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Annotated Bibliographies of Mineral Deposits in Europe

PART 1

Northern Europe including examples from the USSR in both Europe and Asia

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ABSTRACT

These bibliographies and the notes that accompany them have been prepared to aid the economic geologist in his study of ore deposits; they certainly will not do all of his work for him. I have tried to include all ore districts in northern Europe for which sufficient material is available in print to permit a student of ore deposits to obtain a real understanding of each of the northern European deposits in this Volume. Of the deposits selected from the impressive number available from the Soviet Union, only a small sample could be chosen. Those selected are ones for which the literature is sufficient to permit study in some detail by the student but only if he reads Russian with ease.

The references for the deposits in northern Europe are largely in English because the Scandinavian ore geologists now write most of their papers in that language. References in the languages of those countries and in German and French are included where they are thought to be pertinent to the problems being discussed. The large majority of the citations to the literature for the Soviet deposits are, of course, in that language, although references in languages using the Roman alphabet are given where this is possible.

The notes are designed to show: (1) where the deposit is and the grade and tonnage of the ore it contained and, where possible, yet contains; this information is not readily available for deposits in the Soviet Union; (2) the stratigraphy and structure of the rocks of the district, sedimentary, igneous, and metamorphic; (3) the characteristics of the ore bodies in relation to stratigraphy and structure and to ore and gangue minerals; (4) the information bearing on the age of the deposit; and (5) the reasons for the position assigned to it in the modified Lindgren classification (for a full discussion of that classification, see Volume I in this series). These notes should serve to acquaint the student with the deposit and to explain the ideas of those who have worked on it (and mine, where they differ from those readily available in print.) Any student of any deposit must do much more than read these notes.

Several indices are provided in the back of this Volume; these list the authors of the papers cited in the references, the names of the deposits given in the references or in the notes, the metals and minerals produced from each deposit, the ages of the various deposits, the classifications of the various deposits; all indices are arranged alphabetically.

Outline maps are included to show where the deposits are in relation to the general boundaries of the countries in which they are contained.

*ZUSAMMENFASSUNG

Diese Literaturzusammenstellungen und ihre begleitenden Bemerkungen wurden angefertiat, um dem Lagerstättenkundler bei seinem Studium der Erzlagerstätten zu helfen; sie werden sicherlich nicht alle seine Arbeiten für ihn erledigen. Ich habe versucht, alle Erzbezirke in Nordeuropa, für die hinreichend viele Veröffentlichungen vorliegen, einzubeziehen, um dem, der sich mit Erzlagerstätten beschäftigt, zu ermöglichen, jede in diesem Band behandelte Lagerstätte wirklich zu verstehen. Von der beeindruckenden Menge der erfaßbaren Lagerstätten in der Sowjetunion konnten nur wenige ausgewählten Lagerstätten ist immerhin soviel veröffentlicht worden, daß der interessierte Leser auch Einzelheiten erfahren kann, wenn er russisch ohne Schwierigkeiten liest.

Das Schrifttum über die Lagerstätten in Nordeuropa ist im wesentlich englisch verfaßt, hauptsächlich deshalb, weil inzwischen die skandinavischen Erzlagerstättenkundler die meisten ihrer Aufsätze in dieser Sprache schreiben. Arbeiten in den skandinavischen Sprachen, in deutsch und französisch sind dann aufgeführt, wenn sie für wesentlich zum Verständnis der behandelten Fragen gehalten wurden. Die Literaturzitate über sowjetische Lagerstätten sind übersiegend in russisch, wie sich von selbst versteht, aber da, wo es möglich war, wurden auch Arbeiten in Sprachen erwähnt, die das lateinische Alphabet benutzen.

Die Anmerkungen sollen zeigen: (1) die geographische Lage der Lagerstätte, ihren Gehalt und die Vorräte, wenn möglich, auch die heute noch verfügbaren; diese Informationen ist für Lagerstätten in der Sowjetunion nicht leicht zu erhalten; (2) den geologischen Rahmen des Erzbezirks, d.h. die Stratigraphie, die Struktur, ob es sich um sedimentäre, magmatische oder metamorphe Gesteine handelt; (3) die Kennzeichnung der Erzkörper inbezug auf Stratigraphie und Struktur sowie auf Erzmineralien und Gangarten; (4) die für die Alterseinstufung der Lagerstätte wichtigen Angaben; (5) die Begründung für die Einstufung in das modifizierte Schema nach LINDGREN (siehe Band I dieser Serie für eine ausführliche Disskussion dieser Klassifikation). Diese Anmerkungen sollen dazu dienen, den Bearbeiter mit der Lagerstätte vertraut zu machen und ihm die Vorstellungen jener nahezubringen, die die Lagerstätte untersucht haben (und auch meine Vorstellungen, wenn sie abweichend sind von denen, die in der leicht beschaffbaren literatur vertreten werden.)

Jeder Bearbeiter muß jedoch für jede Lagerstätte weit mehr tun als nur diese Anmerkungen zu lesen.

Am Ende des Bandes stehen verschiedene Verzeichnisse: eine Liste der Autoren, die ziteriert worden sind, ein Verzeichnis der Lagerstätten, die in der Literaturzusammenstellung oder in den Anmerkungen erwähnt worden sind, eine Liste der Metalle und Mineralien, die von jeder Lagerstätte gewonnen worden sind, eine Auflistung der Alter der verschiedenen Lagerstätten sowie eine solche der genetischen Einstufung. Alle Verzeichnisse sind alphabetisch geordnet.

Obersichtskarten sind angeschlossen, um zu zeigen, wo sich die Lagerstätten innerhalb der jeweiligen Landesgrenzen befinden.

* Translation provided by Dr Hansgeorg Förster, Rhein.-Westf. Techn. Hochschule Aachen, BDR.

RESUMEN

Estas bibliografias y las notas que las acompanan se han elaborado para facilitar la tarea del metalogenista en sus estudios sobre los yacimientos o les depósitos minerales; sin embargo estos documentos no van a solucionar todo su trabajo. El autor ha intentado incluir todos los distritos mineros del Norte de Europa para los cuales hay suficientesdatos publicados como para permitir, al que se interesa por los yacimientos, tener un conocimiento exacto de cada uno de los depósitos de Europa septentrional mencionados en este volumen, Dado la enorme cantidad de yacimientos en la Union Soviética, el autor se ha visto obligado a elegir un pequeno número de los previamente seleccionados por él. Se han elegido los depósitos acerca de los cuales se han publicado datos suficientes para que el interesado en metalogenia pueda analizarlos con cierto detalle con tal que sepa leer Ruso con facilidad.

Las referencias a los depósitos del Norte de Europa son principalmente en Inglés porque actualmente los metalogenistas escandinavos escriben la mayoria de sus trabajos en ese idioma. Se incluyen referencias en los idiomas de estos paises, asi como en Alemán y en Frances, cuando estas referencias se estiman necesarias para los temas que se debaten. Para los depósitos soviéticos, la mayoria de las citas de obras publicadas aparecen naturalmente en Ruso, aunque se dan referencias -en la medida de lo posible- en idiomas que utilizan el alfabeto latino.

Las notas tienen como propósito indicar: (1) la situación geográfica del depósito asi como la ley y el tonelaje de la mena que contenia y, en su caso, que contiene todavia; esta información no es fácilmente disponible en el caso de los depósitos de la Unión Soviética; (2) la estratigrafia y la estructura de las rocas del distrito (rocas sedimentarias, igneas y metamórfica); (3) las caracteristicas de los cuerpos mineralizados respecto tanto a la estatigrafia y a la estructura como a la mena y a la ganga; (4) la información respecto a la edad del depósito; y (5) las razones que justifican la posición asignada a

este depósito en la clasificación modificada de Lindgren (para un conocimiento más completo de esta clasificación, ver el volumen I de esta serie). Estas notas deben proporcionar al interesado información respecto al depósito en cuestión y presentarle las ideas a las que se ha llegado por investigación de sus propias ideas cuando estas difieren sensiblemente de las recogidas en obras publicadas al respecto. En todo caso, todo metalogenista deseoso de conocer detalladamente un yacimiento determinado debe hacer mucho más que leer estas notas.

Varios indices aparecen al final del volumen y recogen, por oden alfabético: los autores de las publicaciones citadas en las referencias, los nombres de los depósitos señalados en las referencias y las notas, los metales y minerales existentes en cada vacimiento asi como las edades y la clasificación de los diversos depósitos.

Se incluyen mapas esquemáticos para situar los depósitos dentro de los limites generales de las regiones o paises a los cuales pertenecen.

RESUME

Ces bibliographies et les notes qui les accompagnent ont été élaborées en vue de faciliter la tache du métallogéniste lors de ses études sur les gisements ou les dépots minéraux; toutefois ces documents ne vont pas solutionner tout son travail. L'auteur a tenté d'inclure tous les districts miniers du Nord de l"Europe pour lesquels des données suffisantes existent dans la littérature afin de permettre à celui qui s'intéresse aux gisements d'avoir une connaissance exacte de chacun des dépots d'Europe septentrionale mentionnés dans ce volume. Vu l'enorme quantité de gisements qui existent en Union Soviétique, l'auteur s'est vu obligé à ne retenir qu'un petit nombre de ceux qu'il avait préalablement selectionnés. Le choix s'est porté sur les dépots pour lesquels des données suffisantes ont été publiées afin que celui qui s'intéresse à la métallogénie puisse les analyser en detail à condition qu'il sache lire le Russe avec facilité.

Les références serapportant aux dépots du Nord de l'Europe sont principalment en Anglais du fait qu'actuellement les métallogénistes scandinaves écrivent la majorité de leurs travaux en cette langue. Des références dan les langues de ces pays ainsi qu'en Allemand et en Francais sont aussi mentionnées lorsque ces références s'avèrent nécessaires lors de la discussion des thèmes traités. Pour les dépots soviétiques, la plupart des citations des travaux publiés apparaissent naturellement en Russe, bien que -dans la mesure du possible- des références dans des langues utilisant l'alphabet latin sont également fournies.

Les notes ont pour but d'indiquer: (1) la situation géographique du dépot ainsi que la teneur et le tonnage du minerai qu'il contenait et, le cas échéant, qu'il contient encore; cette information n'est pas facilement disponible pour les dépots de l'Union Soviétique; (2)la stratigraphie et la structure des roches du district (roches sédimentaires, ignées et métamorphiques); (3) les caractéristiques des corps minéralisés par rapport a la stratigraphie et à la structure ainsi que les caractèristiques du mineral et de la gangue; (4) l'information concernant l'age du dépot; et (5) les raisons justifiant la position assignée a ce dépot au sein de la classification modifiée de Lindgren (pour une connaissance plus complète de cette classification, voir le volume I de cette série). Ces notes fournissent à l'intéressé des informations concernant le dépot minéral envisagé et lui présentent les idées qui resultent des recherches réalisées sur ce dépot; l'auteur a egalement exprimé ses propres idées lorsque celles-ci diffèrent sensiblement de celles recueillies dans les travaux publiés. Cependant, tout métallogéniste désireux de connaitre en détail un gisement déterminé doit faire davantage que de lire ces notes.

Plusieurs indices apparaissent en fin de volume et mentionnent, par ordre alphabétique: les auteurs des publications citées dans les références,

les noms des dépots signalés dans les références et les notes, les métaux et minéraux existant dans chaque gisement ainsi que les ages et la classification des divers dépots.

Des cartes schématiques sont jointes pour situer les dépots par rapport aux limites générales des régions ou des pays auxquels ils appartionnent.

Translations provided by Professor Dr Georges Monseur, Departamento de Geologia, Facultad de Ciencias, Universidad Autónoma, Canto Blanco, Madrid 34, España

РЕЗЮМЕ

Предлагаемые библиографические обзоры ставят своей целью помочь геологам в изучении рудных месторождений. Я попытался охватить в этой работе все те рудные районы Северной Европы; количество опубликованных сведений, по которым могло бы позволить исследователю составить достаточно полное представление о каждом из месторождений, включенных в этот том. Лишь немногие из обширного списка заслуживающих внимания месторождений СССР смогли быть охарактеризованы в этой работе. Были выбраны те месторождения, по которым в литературе имеются более менее детальные сведения, однако анализ этой литературы доступен лишь геологам, свободно читающим по-русски.

Большая часть ссылок в библиографии по месторождениям Северной Европы дается на работы, написанные по-английски, так как и скандинавские геологи в последнее время публикуют большинство своих статей на английском языке. В тех случаях, когда это существенно для обсуждаемых проблем, даны также ссылки на работы, написанные на немецком, французском и на языках скандинавских стран. При описании месторождений СССР, помимо ссылок на русские работы, указаны и имеющиеся публикации на европейских языках.

Обзор по каждому на месторождений включает следующие сведения: 1. Где расположено месторождение: его масштабы: добыча металла в тоннах, имеющиеся запасы - подобная информация по месторождениям Советского Союза практически отсутствует. 2.Стратиграфия и структура района, типы и последовательность осадочных, изверженных и метаморфических пород. 3.Характеристика рудных тел - их положение в разрезе, строение, состав рудных и жильных минералов. 4. Сведения о возрасте оруденения. 5. Критерии для отнесения месторождения к одному из типов усовершенствованной классификации Линдгрена (Обсуждение классификации дано в 1 томе этой серии). Предполагается, что эти сведения помогут исследователю ознакомиться как с главными чертами месторождения, так и с идеями изучавших его геологов и автора обзора (там, где его представления расходятся с наиболее доступными в литературе). Несомненно, любой геолог, приступающий к изучению любого месторождения, на должен ограничиваться лишь ознакомлением с предлагаемым обзором.

В конце тома приведен ряд алфавитных указателей: авторов цитированных работ; названий месторождений; добываемых металлов и минералов; указатели возраста месторождений и их места в классификации.

Даны обзорные карты, указывающие расположение месторождений каждой страны по отношению к ее границам.

Translation provided by Dr. V.A.Baskina, Institute of Ore Deposit Geology, Petrography, Mineralogy and Geochemistry of the USSR Academy of Sciences, Moscow, USSR.

INTRODUCTION

This is the third volume in a series designed to discuss the major ore deposits of the world for which the published literature is adequate, or better, for a study of each ore body here presented. Some deposits very minor in size are included because of their interesting or even unique geology. Many deposits, the geology of which almost certainly would justify their inclusion here, have been omitted because the available literature is too scant to permit the reader to verify what I claim to be facts.

Originally the third volume of this series was planned to cover all of Europe and a few of the geologically most interesting ore deposits of the USSR. It soon was discovered, however, that the material available from the Soviet Union would permit the discussion with profit of 30 of that huge country's deposits. In addition, the literature on the ore bodies of northern Europe is large enough and sufficiently comprehensive to require a separate volume for that part of Europe and for the 30 Soviet deposits. Volume III, therefore, has been divided into two parts. The first of these, this one, is designated as Volume III, Part I with the title being: Northern Europe and Selected Examples from the USSR. The second part will bear the title: Central and Southern Europe.

This part of Volume III will depart from my original plan of limiting the deposits included to those formed only (or dominantly) by magmatic and hydrothermal processes and contains deposits formed in whole or in major part by surface sedimentary processes as well.

As I have done before, I have not limited the references selected only to those published recently. Although I have been criticized for doing this, I find that many of the older papers contain information useful to the student that is not available in more recent works.

In arranging the bibliographies and notes in this Volume, the basic division is by countries; within the various countries, the organization differs from one to the next. In Ireland, the geology of the ore-bearing part of that island (which largely excludes for the purposes of this Volume) Northern Ireland) is discussed briefly and short descriptions of several deposits, the literature on which does not make possible a more extended treatment. This section is followed by detailed studies of three major deposits, given in alphabetical order, for which sufficient information is available to justify so doing.

For Great Britain (again excluding Northern Ireland), the tin-tungsten-copper deposits of Cornwall are treated first. This discussion is followed by a general statement about British lead-zinc deposits. Then come discussions about six of the major lead-zinc deposits of England with the four Mississippi-Valley type deposits being placed before the two in which higher-temperature processes took part and in which silver is of appreciable economic importance. The single deposit in Scotland and the one in Wales are studied in that order; they are of the higher-temperature, silver-bearing variety.

In Norway, three of the deposits (Sulitjelma, Bleikvassli, and Røros-Kvikne-Killingdal) that have a similar genesis are given first. The order of the remaining four, no one of which has genetic similarities to any of the other three, are arranged from north to south.

For Sweden, a general summary of the metallogeny of that country is given first; in this section are included brief descriptions of certain ore bodies that are of such geologic interest as to justify separate presentations if only the literature for them was sufficient in extent and variety. The 16 Swedish deposits (or deposit groups) considered separately are arranged roughly by geographic areas. The first of these is the one that contains the iron ores of central Sweden in which the numerous ore bodies are of a few divergent types. These deposits are similar enough not to require detailed geologic descriptions, but several of them are considered briefly. A few deposits in the central Swedish area are so different in their geologic characteristics that they warrant discussion in detail; these are: Grängesberg, Långban, Falun, and Garpenberg. Also included in this area is the unusual lead-zinc sulfide deposit of Åmmeberg, which is near the southern margin of the central Swedish ore area.

The next general area to be considered is that of the Skellefte District, all the deposits of which are located in Västerbotten province. This district is here divided into three geographical areas to make the discussions more manageable. Then follow three other deposits in Västerbotten, Lainijaur, Stekenjokk, and Varuträsk, no one of which is genetically similar to the other two or to the Skellefte ores. After these three come the deposits in the province of Norrbotten, Aitik, Laisvall, and Kiruna-Gällivare, that are not in any way genetically connected. The last two deposits are the carbonatite of Alnö Island and the farthest south of all major Swedish metallic ore deposits, Smålands Taberg.

In Finland, a brief description of the ore geology is given, followed by detailed discussions of 12 deposits (or deposit groups) for which treatment in depth is possible. These are arranged alphabetically, although grouping according to genetic characteristics might be possible. Such an organization would put the deposits of the nickel-copper belt (or belts), Ahlainen-Sääksmäki and Hitura-Makola-Kotalahti, together. These would be followed by the copper or copper-zinc-lead deposits [Hällinmäki, Hammaslahti, Orijärvi-Aijala-Metsämonttu, Outokumpu (and its stalleites), Pyhäsalmi, and Vihanti] is a separate group. These deposits show enough differences in geologic character, however, that this was not done. The remaining four deposits (Kemi, Mustavaara, Otanmäki, and Ylöjärvi) are so distinctly different as not to be placed in a single genetic category. The study of Finnish ore deposits, therefore, is begun with a general statement about the ore geology that is followed by a discussion of each of the 12 major deposits (or deposit groupings) in alphabetical order.

The only two deposits in Poland worthy of separate treatment are so diverse in character as to make it reasonable to present them also in an alphabetical arrangement.

The organization of the Soviet deposits is done on the alphabetical positions of the political divisions in which they lie; a table to their arrangement is given in the introduction to the Soviet section of this Volume.

The 80 deposits (or deposit groups) discussed in detail here are located in the following countries:

Ireland	3	Wales	1	Finland	12
England	7	Norway	7	Poland	2
Scotland	1	Sweden	17	USSR	30.

These 80 do not include the four general discussions presented in the Volume.

Probably I shall receive considerable criticism for attempting to denegrate the role of chloride ion in the transportation of ore elements from the source(s) from which they are derived to their site(s) of deposition. My conviction that sulfur must be introduced along with the metal ions is, as I have said many times, based on the complexity of sulfide minerals that must require an intimate and continued association of the various sulfur radicals with metal ions, and cannot be achieved by chance encounters with sulfur ions lying in wait for wandering ions of the various ore metals. Yet experimental work strongly indicates that the pH of ore fluids must be within ranges that preclude the formation of the necessary metal-sulfur complexes. I consider that this conclusion will some day be upset by the "missing experiment;" I only wish that I may live to see it carried out and the results published.

In this connection, I would like to refer to the recent work of Hansgeorg Förster and Rudiger Oberste-Padtberg that was presented during the 12 to 15 January meeting in London of the Contact Group "Geology of Ore Deposits held at the Institute of Geological Sciences. These authors point out that the stability of the pair H_2S/H_2SO_4 is doubtful (Phillips and Williams, 1965) although Barnes contends (quoted by Förster and Padtberg, 1982) that these are the only two species in which sulfur occurs in hydrothermal fluids. Phillips and Williams, however, have argued that, in acid solution, H_2S will tend to reduce any sulfur in any sulfur-bearing radical to the native state. On the basis of this concept, they believe that intermediate oxidized sulfur species (IOSS) can and will participate in dissolving, transporting, replacement, and precipitation reactions that take place as a hydrothermal fluid makes its way upward from the locus of its development to the final stages of its precipitation cycle. During this journey, the solution will deposit not only sulfides but also sulfates and sulfosalts by making use of the IOSS produced in its upward progress.

The opinion of Förster and Padtberg is that IOSS such as SO_3^{-2} (sulfite), $S_2O_6^{-2}$ to $S_4O_6^{-2}$ (thionates), $S_2O_3^{-2}$ (thiosulfate), and SS^{-2} to S_4S^{-2} (polysulfides) may be present in hydrothermal fluids below 600°C and $\lg P_{O_2}$ equal to -40 with the IOSS initially being much more abundant than sulfate. The formation of IOSS is favored by the action of $FeS\text{-}FeS_2\text{-}Fe_3O_4$ and $FeS_2\text{-}Fe_3O_4\text{-}Fe_2O_3$ buffers (Holland, 1965). Sources of such IOSS in natural hydrothermal solutions would be: (1) sulfur transferred from cooling magmas by hydrothermal fluids (internally or externally generated in relation to the magma); (2) sulfate-rich waters such as those provided by sea water or from intraformational brines, these waters being heated by contact with rocks carrying residual heat from magmatic processes; and (3) sulfur-rich waters generated by the oxidization of sulfides encountered by hydrothermal fluids during their upward passage. Within such solutions, the disproportionation of the original sulfate depends on the temperature, pressure, contained catalyzers and inhibitors (such as NH_3, iron minerals, and silica gels) in the hydrothermal fluid. Förster and Padtberg suggest that the IOSS so produced may exist in hydrothermal fluids long enough to be responsible for the dissolution and transportation of sulfur, the sulfur-bearing radicals so formed being available to combine with metal ions to produce both sulfide (in the broad sense) and sulfate minerals of the wide variety found in ore deposits, thereby providing an explanation of the precipitation of barite in temporal conjunction with the usual sulfur-bearing minerals of hydrothermal deposits.

Reading through Barnes' chapter in his "Geochemistry of Hydrothermal Minerals" indicates that Förster and Oberste-Padtberg have somewhat misinterpreted Barnes for he says (p. 408) that:

> "(sulfur) Complexes may be formed by reactions of metals with most of the aqueous species shown in (Barnes') Figure 8.1 (p. 409 in Barnes), including the (generally) subordinate species, polysulfide, thiosulfate, and sulfite. During oxidation of sulfides, these subordinate species may be produced in concentrations . . . . that exceed those of the thermodynamically dominant sulfide and sulfate species. Thiosulfate and polysulfide species can persist metastably at significant concentrations for long periods, and, in the latter case, to above 200°C in near-neutral solutions . . . . There may even be a small field near neutrality where polysulfide species are more stable than sulfides or sulfates at low temperatures . . . . In contrast sulfite species disproportionate relatively rapidly even at ambient temperatures . . . ."

This quotation indicates that Barnes is in essential agreement with Förster and Oberste-Padtberg on the matter of disproportionation of sulfur species. Barnes goes on to say that:

> "The dominant complexing ligands among sulfur species are more likely to be HS^- or H_2S in hot, reducing, hydrothermal solutions where equilibrium is favored by time and high temperatures. Stronger complexing is caused by the bisulfide ionic field than by the dipole field of molecular H_2S. For this reason, it is in, *or close to*, the stability of HS^- that bisulfide complexing is more likely to result in higher solubilities of the sulfides."

However, Barnes goes on to say that:

> "geologically improbable, very alkaline solutions are required above approximately 300°C for HS^- to become the dominant species. For this reason (Barnes believes), bisulfide complexes may be important only at temperatures below roughly 350°C."

This conclusion, though well documented, seems, so far as natural hydrothermal solutions are concerned, to be subject to some question. For example, numerous ore deposits contain the three major ore sulfides (chalcopyrite, sphalerite, and galena) that almost certainly were deposited in a continuous sequence at temperatures from over 500°C down to as low as 200° to 250°C (or locally down to as low as 100°C±. How, if HS^{-1} cannot be a dominant species above 300° to 350°C, does this continuous deposition occur from over 500°C on down without any apparent change in the character of these three sulfides (until the lack of silver in galena is encountered below about 175°C)? Again, it is difficult, if not foolish, to argue against the strong evidence for the truth of what Barnes says about the upper limit of the importance of HS^{-1} in hydrothermal fluids; it seems to me, however, that some important fact (or facts) still is (or are) missing from the experimental picture. It seems reasonable to consider further work to be required before a completely acceptable answer to this problem has been achieved.

Obviously, not all of the questions concerning the species and sequences of sulfides, sulfosalts, and sulfates present in ore deposits are answered by assuming that the disproportionation of oxidized sulfur species into sulfide (and thiosulfide) and sulfate is a valid concept. A thorough study of the production of ions containing sulfur in various valences through disproportionation may go far toward solving the problem of the presence of the various species of sulfur-containing radicals in hydrothermal (in the general sense) sulfide deposits definitely is needed.

I should repeat, from the Introduction of Volume II *"that, because of a lack of space and funds, geologic maps of the various district in this volume have been omitted. Such maps can be found in certain of the references listed for each deposit, and these should be used for a better understanding of the notes that appear with each bibliography.*

In all instances in this Volume, distances, from the smallest to the largest, are given in the metric system even though this system normally is not used in those references concerned with deposits in Great Britain and Ireland.

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In any work such as this, much of the information included is derived from printed papers. These are contained in a wide variety of publications, and these are shelved in less than their entirety in the various libraries to which I have access, I have made the greatest use of the libraries of the U.S. Geological Survey in Reston, Virginia, and, to a lesser extent, in Menlo Park, California. I am most particularly indebted to Barbara Chappell, Chief, Reference/Circulation Section in the Library at Reston. A considerable amount of information was obtained from the Lenin Library in Moscow.

In the early days of the work on this Volume, I made extensive use of the Library of the College of Earth and Mineral Sciences at The Pennsylvania State University, but, following my retirement from that institution, my connection with that source essentially has stopped. During my days at Penn State, Emilie T. McWilliams, Librarian of that College, and her chief assistant, Helen Bokach, were of great assistance.

For the typing of the final draft of this Volume, as it was sent to the publisher, I have had to depend on my own typing; this may account to some measure for the errors that this Volume yet may contain.

The maps have been drawn, from my rough sketches, by the cartographers of Pergamon Press Ltd in Oxford, and to them I owe my thanks. The final typing was also done at Pergamon, as so much accurate typing was beyond my abilities.

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IRELAND (EIRE)

IRELAND (GENERAL)

Late Paleozoic	Lead, Zinc, Silver, Copper, Pyrites, Barite	Mesothermal to Leptothermal

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Notes

Ireland has been the scene of metal working and fabrication as long ago as 2000 B.C. Although it is possible that the metals so used were obtained through a well-organized system of international trade, it is much more probable that the metals used were obtained indigenously. During the period of the Viking invasions (beginning on or about 800 A.D.)., much of early Irish historical and archeological data, including the processes of metal working and fabrication, were lost or forgotten. There are, nevertheless, several written, as well as archeological, references to mineral working in the times of the wars of the Irish with Vikings. The Viking rule in Ireland reached its zenith in 977 but was broken by two remarkable Irishmen, Malachy, King of Meath, and Brian Boru in the late 10th century. By 1002, Brian became the acknowledged High King of Ireland and ended the Viking terror, but he was killed after the battle of Contarf in April of 1014. The high kingship was resumed by Malachy, but he died in 1022, and the rule of the country was divided among many "kings". This turbulent time reached its climax in 1156 with the death of Ireland's last great king, Turloch O'Connor. The first Norman-English invaders of Ireland were the FitzGeralds, but the first Norman to assume a position of power in Ireland was Richard FitzGilbert (Earl of Pembroke); he became King of Leinster in 1171. Pembroke (or Strongbow) was finally victorious over the more or less combined Irish after first defeating the last Viking invaders. Strongbow, however, thought it the better part of valor to prefer being the principal English nobleman in Ireland to being a king in that country, so he swore allegience to King Henry II of England who had landed with an army at Waterford in October of 1171. From this time onward (until independence in 1921) control of almost all facets of Irish life, including mining and metallurgy, passed to the English. Written records of these metal-related activities still existent date from the 17th century when mining appears to have been a viable and profitable industry in Ireland. Copper, sulfur, and iron were recovered from the Avoca area as early as the 16th century and lead and silver from Silvermines from the 17th (of course, zinc is now the main metal produced from that district). Other early mines were those for lead at Glendalough and for copper in the west of County Cork. Considerable prosperity in mining continued until about 1880 when lower metal prices drove the Irish mines out of business. It has been reported that, over the centuries, over 500 localities actually underwent at least some mining, and some of the localities were, for the times when they were worked, of considerable size. During all this time, however, methods of processing were primitive, so that hand-cobbing was the principal means of cencentrating the ores. This resulted in mining being confined to coarse-grained ores, preference being given to monomineralic ones, nor was it possible to process the complex sulfides even if they were easily concentrated by hand.

Even after the independence of Ireland was achieved in 1921, attempts to attract foreign companies to prospect in Ireland were of limited success until the Minerals Development Act was passed in 1940. Obviously, this was an unfortunate date, for few mining companies had personnel or capital to justify exploration at that time in a country of no greater promise than Ireland. After World War II, however, some small operations were successful to a minor extent, mainly in the areas where small mines had been operated for many years - for example, Silvermines, Abbeytown [close to Sligo (N54°17'N, 8°37'W)], and Benbulben [about 10 km north of Sligo]. In the late 1950s, however, the Irish government offered generous fiscal benefits for the exploitation of deposits discovered. The first deposit discovered under the terms of this act was Tynagh. Attention was thereby brought to the lower Carboniferous limestones of central Ireland, and the exploration boom of the following years lead to the discovery of Gortdrum (about 5 km north of the town of Tipperary in the county of that name) and of earlier unknown ore bodies at Silvermines. Interest in the 1960s shifted to the more deeply buried deposits in the north-central part of the country (although still in the lower Carboniferous limestones) with the most impressive discovery being made at Navan (An-Uaimh - N53°39'N, 6°41'W) in County Meath, about 45 km NW of Dublin. Although considerable mineralization has been located at Keel and Ballinalack (west of Navan) and Mallow in County Cork (52°08'N, 8°39'W), economic conditions have not been such as to permit mining of any one of these. At least a dozen other potentially worthwhile prospects have been more or less delimited, but work apparently is not active at any of them. In addition to those deposits named here and the dozen not named, several geophysical and geochemical anomalies are known in Ireland, but the meaning of them has not yet been determined. Recent changes in the mining laws of Ireland, designed to return more of the profits from any mining operations to the government and less to the companies working them, have made the country less attractive to explorationists and, if unmodified, may result in the full mineral potential of Ireland not being achieved. The government, however, has done some things to aid prospecting by increasing the capabilities of the Geological Survey of Ireland and by undertaking to provide full air-photographic coverage of the country. Plans also exist for an air-borne magnetometer survey of Ireland which may provide clues to deposits as yet unknown.

As Evans points out (1976), the main elements of Irish geology provide the most westwardly extensions of some of the principal structural units of both Great Britain and the European continent proper. The metamorphic portion of the Scottish and Norwegian Caledonides is present in northwest Ireland, and this area is composed of both Precambrian and Cambrian rocks; a continuation of these strata also is found farther southwest in northwest Mayo and western Connemara. This section of the Caledonian fold belt contains as important rock types both pelitic and semipelitic schists and quartzites, plus much marble and metabasites, that correspond to the Precambrian and Cambrian Dalradian series of Scotland. Igneous rocks of Dalradian type, both migmatitic and igneous granites, are located among these metamorphic rocks. In addition to these metamorphic Caledonian rocks, sedimentary rocks of Ordovician and Silurian age, similar to those in the Southern Uplands of Scotland, are located in southeastern Ulster and the northeastern portion of Ireland. Again, the lower Paleozoic rocks of the northwest of England and of north Wales, both volcanic and sedimentary, are similar to those in southeast Ireland; this terrane includes the largest granite mass outcropping in the British Isles. Considerable outcrops of these lower Paleozoic rocks also are known in the west of the Republic and are present as inliers of various sizes in the central Irish plain. Between the lower Paleozoic rocks of the inliers and the lower Carboniferous rocks that provide the bulk of the bedrock of the Central Irish Plain normally are annular bodies of the continental facies of the Old Red sandstone. These inliers rise up through the lower Carboniferous beds as mountains or low ranges of hills; not all of them are eroded enough to expose the lower Paleozoic strata.

The rocks of these lower Paleozoic inliers (Evans, 1976) are normally geosynclinal shales, mudstones, siltstones, and graywackes that have been folded and in which some slaty cleavage has been produced. The trend of the folding is northeast in the eastern part of the Central Plain but rotates to an east direction in the western part. These rocks may be as much as 3000 plus m thick and probably underlie the entire Central Plain. Evans points out that the Precambrian and lower Paleozoic rocks provide the basement of much of Ireland and formed the foundation of the foreland during the succeeding orogeny. Evans believes that a knowledge of this fact is necessary if the regional tectonic environments of the main Irish metalliferous deposits are to be understood.

Ireland, as well as much of Europe, was affected by the formation of the Variscan geosyncline (Evans, 1976); rocks formed in the northern part of this geosyncline are exposed in southwestern Ireland (as well as southwestern England and northern France, Belgium, and Germany). These rocks underwent several phases of folding and regional metamorphism with, at the end of the Carboniferous, the introduction of granite and various types of ore mineralization, the Hercynian epoch of mineralization, the effects of which were felt from Ireland to Czechoslovakia and Poland. In Ireland, as in other northern European countries, the northern foreland of this geosyncline is weakly developed. In the Devonian, the rocks of the foreland are the terrestrial facies of the Old Red sandstone whereas the rocks of the geosyncline are marine Devonian. In the lower Carboniferous, the marine rocks are clastics but the foreland rocks are principally carbonates. Although the Variscan geosynclinal rocks are well-established in southwestern England, they are lacking in Ireland, but it is suggested that the northern edge of the geosyncline was only a short distance south of the present south coast of Ireland. On the contrary, the principal orogenic front of the fold belt of the Variscan is north of the northern boundary of geosynclinal rocks and usually is placed along the line of the Killarney-Mallow thrust front (52°03'N, 9°30'W, 52°08'N, 8°39'W). South of this line cleavage folding dominates, whereas north of it concentric folding is impressively developed. The northern limit of this zone of concentric folding is located at about the latitude of Dublin (53°20'N); beyond this line, the upper Paleozoic, mainly shallow-water carbonate, rocks are only slightly folded. The major mineral finds of the last 20 years or so have been made in these two zones of the Variscan foreland, an area that makes up the Central Plain of Ireland. The Devonian beds south of the Killarney-Mallow line are, at least for the present, not known to be appreciably the host of ore mineralization.

Although the upper Paleozoic rocks in the geosyncline may be as much as 10,000 m thick, those of the Central Plain are much thinner. In the south of this area, the Carboniferous is conformable on the Devonian; the base of the Carboniferous becomes younger as the Plain is crossed from south to north with the northern half of the Plain not being covered by the sea until the first of the Visean. Here the Carboniferous rocks lie on lower Paleozoic and Caledonian metamorphics. The maximum thickness of these Carboniferous sediments, with their gentle dip to the south, have a maximum thickness of about 1000 m. The basal strata of the lower Carboniferous are calcareous siltstones, sandstones, and shales (so-called "mixed-beds") that were produced in tidal zones and shallow water. Upward these beds grade into thick accumulations of argillaceous and bioclastic limestones. Above these strata are the thick mudbank limestones ("Waulsortian"); locally these Wualsortian rocks are interbedded with the older limestones; the Waulsortian rocks are composed of generally pale-colored, fine-grained carbonates that include, among other types, micrites, biomicrites, and encrinites (crinoidal limestones). Although these Waulsortian rocks were once designated as "reef limestones"; this term largely has been dropped. Locally, the Waulsortian rocks may be missing or sporadically developed, their stratigraphic place is occupied by oolites, dolomites, and/or bioclastic limestones. Above the Waulsortian beds, a variety of rock-types, such as dark compact limestones ("Calp"), are quite common or, locally, cherts, cherty limestones, shales, or other rocks may be present. Running west from Dublin across Ireland in the central part of the Plain is a so-called "basinal" facies. The main rocks of this "basin" are dark argillaceous and carbonaceous limestones and shales, low in fossils; locally Waulsortian rocks-types are present in the basinal facies.

No Hercynian intrusives of major size are uncovered in Ireland, but several minor intrusions are known both in the southern Hercynian fold belt and in the Central Plain. In a few places, Hercynian volcanic activity produced lava flows and pyroclastic deposits. Evaporites are, at best, uncommon, and it is suggested that they are more probably sulfates than chlorides.

In the various belts into which the rocks of Ireland are divided, the degree of contained ore mineralization is quite varied. The Caledonian metamorphic belt (in the west of Ireland) has only a few small lead, zinc, and copper deposits that have not been worked for many years; they certainly, so far as is known, are too small for economic operation under modern economic and technological conditions.

Aside from a few small veins of lead and zinc, the Silurian region of southeast Ulster is not mineralized. In the volcano-sedimentary belt of southeast Ireland, both vein and stratiform deposits are known. Only two ever were of major importance. One of these, the Bunmahon deposits, have not been operated since the early 1900s and information about them is sparse. They are a system of ramifying quartz veins that cut Ordovician volcanics and sediments. Chalcopyrite was the only important ore mineral; other sulfides were sparsely developed. Ore bodies on these veins were up to 20 m thick. Because the eroded top of one vein is overlain by a Devonian conglomerate, the ore must be older than the Devonian. Locally, lead-zinc veins occur in the Bunmahon area and in much of the area of lower Paleozoic rocks in southeast Ireland, but they are of no known or expected economic importance. The character and source of the ore-forming fluids for these lower Paleozoic ores is currently unknown.

The deposits at Avoca are discussed in some detail later in this volume.

In the Variscan fold belt, the rocks are old Red sandstone facies in which a few, uneconomic red-bed copper deposits are present. Several vein-type deposits (certainly epigenetic) have been discovered but only Allihies, near the tip of the peninsula between the Kenmare River (north) and Bantry Bay (south) on the west coast of County Cork, had an output worth mentioning. The area was explored to a considerable degree in the 1970s (and perhaps later). Despite this area being a westward continuation of the metallogenic province of the European Variscanides, no new deposits appear to have been found. The Allihies deposit, however, has been followed down for at least 800 m with reserves of all categories being perhaps 2 million tons of 1.6 per cent copper ore. Evans suggests that this depth of copper mineralization in the Variscan orogenic zone may indicate a tin mineralization at greater depths than have been explored at Allihies. The possibility of granitic plutons or larger bodies in the area is suggested by the presence of a faulted continuation of the Cornubian batholith in the Atlantic about 160 km south of southwest Ireland and negative gravity anomalies measured on, and tectonic breccias found in, the rocks of this area.

The Central Plain of Ireland, however, is (thus far) the major site of present, and probably future, Irish ore metalliferous mining. First production in this area, not from vein deposits, was at Abbeytown (in northwestern Ireland, about 50 km south-southwest of Donegal) principally between 1950 and 1961. This stratiform deposit probably was known before 1645 (Williams and McArdle, 1978), but no certain information of actual mining is dated prior to 1785, then monks mined the *argentiferous* galena. In 1857, the ore was reported to contain over 360 grams of silver per ton. The ore contains more sphalerite than galena, but zinc was not recovered until 1917, when the old dumps were retreated and several hundred tons of zinc and lead were recovered and about 550 kg of silver. Some minor exploration at this time found that the ore extended below the old workings, but nothing more was done until 1949-1950 when over 750,000 tons of "profitable" ore were proved. During the open-pit mining of this ore, exploration at lower levels found two new ore bodies that were worked underground between 1950 and 1961. When these two ore bodies were worked out (1961), and exploration concurrent with the mining failed to find more ore, the mine was closed. Total output from Abbeytown is estimated (Williams and McArdle, 1978) at about 1 million tons in which the grade was 2.5 per cent zinc; 1.5 per cent lead; and "a significant content of silver."

The Abbeytown mine is located on the northern limb of a shallow east-west syncline in Visean limestones and clastic sediments; the structure plunges gently to the west. The syncline is bounded by faults on both sides the mine area is, respectively, 1600 m from the south bounding fault and 400 m from the north boundary fault. Both of these faults have brought quartzites and schists of the Ox Mountains metamorphic belt into contact witl Visean rocks, and both faults are parts of the Ox Mountain fault system. To the east of the area of the mine, the Visean beds are unconformable on metamorphics similar to those of the Ox Mountains belt. The lower Visean rock sequence is typical of the Central Plain, the thickness of which is somewhat greater than 150 m.

The mineralization is mainly silver-bearing galena, sphalerite with common pyrite and minor chalcopyrite; two main mineralized zones exist, Index Bed and Lower Ore. In the Index Bed, the mineralization is as stringers and grains of sphalerite and galena with the better ore on the upper side of the bed where it occurs in breccia zones, vugs, and veins; in other beds, the ore is developed along bedding planes and joints. Dolomitization seems to be associated with the ore. The lower ore bodies, found by the 1950 to 1953 exploration, are within a crinoidal limestone in the main but also extend into a grit below the limestone to a limited extent. The ore bodies are in elongated breccia zones that trend NNE and are located along the flanks of minor folds developed late in the folding. Some NNE faults and minor NNE folds are connected with the breccia zones. The ores definitely were introduced after the containing rocks had been folded, faulted, and brecciated and must, therefore, be epigenetic. What the ore fluids were or where they came from is not known; they probably did not come from either of the boundary faults as the ore does not extend to them either to the north or the south.

Old mines on small, but high-grade lead-silver vein deposits were worked near Ballyvergin during the 19th century. Exploration by extensive geochemical and geophysical work in 1960 to 1961 was carried out by Irish Base Metals Ltd. in the east-central part of County Clare to look for a downward extension of these vein deposits. A strong induced-potential anomaly was found in the vicinity of the old Ballyvergin mine, and diamond drilling on this site found a deposit of disseminated but low-grade copper sulfides in lower Carboniferous limestones. This was the first discovery of disseminated sulfides that were associated with minor vein sulfides. The geologists of Irish Base Metals believe that the ore is contained in a small flexure in the folded limestones, producing a dome-shaped deposit. Unfortunately, the proved reserves amount to only about 150,000 tons of 1.0 per cent copper and 16 grams/ton of silver; another 70,000 tons of similar ore may underlie the known ore in an incompletely drilled rock volume. Ballyvergin, therefore, is uneconomic for as far in the future as can be predicted, but its discovery is an encouragement for examination of other targets.

In 1961 to 1962, Rio Tinto-Zinc investigated an area in County Longford, about 65 to 70 km west of Navan, which was followed by an induced polatization study of a limestone-sandstone contact. Drilling, based on this work, discovered a lead-zinc deposit at Keel in that county. Within the basal sandstones and conglomerates of the lower Carboniferous and adjoining the major Keel fault, disseminated sulfides of these two metals were found. The associated, east-dipping normal faults of Hercynian age apparently directed the ore fluids into nearby breccias and fractures; such structural features also control the disseminated mineralization in such a manner as to give it (the mineralization) a stratiform appearance. The main sulfide is sphalerite, and it contains 1 per cent cadmium in solid solution. The silver appears to be contained in the zinc, probably as inclusions (exsolution blebs ?) of silver-bearing tetrahedrite. The other sulfides, in order of decreasing abundance are: galena, pyrite, marcasite, and chalcopyrite; barite also is present but minor in amount. The main gangue minerals are quartz and carbonates. The deposit almost certainly is epigenetic. The diamond drilling indicated reserves of 1,700,000 tons of 7 per cent combined lead and zinc. Even when a 180 m shaft had been sunk and considerable subsurface exploration had been done, the total ore found was thought too small to justify mining.

In 1963 to 1964, a potential copper-silver ore body was located and outlined by Dennison Mines Ltd. near Galbally at the western end of the Vale of Aherlow, County Limerick. It is reported to contain nearly 5.5 million tons of mineralized rock with a grade of 0.89 per cent copper and 41 grams silver to the ton. The lower 3 million tons of the deposit has a grade of 1.7 per cent copper and 47 grams of silver to the ton. At least at present, this deposit is not economic to mine.

Here the lower Carboniferous limestones are folded into an east-west-trending syncline, on the northern limb of which the Aherlow deposit is located. The syncline plunges gently to the west and is bordered to the north and south by slices of Devonian sediments that have been overthrust into the area. The mineralized limestones are muddy bioclastic rocks that grade downward into more muddy and more sandy beds. The principal sulfides are chalcopyrite, bornite, and chalcocite with minor quantities of sphalerite, galena, and pyrite. The sulfides are very fine-grained and occur as disseminations and in veinlets and breccia cements. Native silver and stromeyerite are known, but Williams and McArdle report that silver probably also is present in the lattices of the sulfides; no tetrahedrite-tennantite have been seen in the deposit. To some extent, lithology controls the mineralization in that chalcocite and bornite are concentrated in the more shaly rocks. The intensity of mineralization, however, is directly connected to the degree of rock deformation with the ore body occurring in a steeply dipping zone related to a flexure in the north-dipping beds. This ore zone has been followed from east to west for more than 600 m and down for 250 m; the ore zone is as much as 30 m wide.

Another copper-silver prospect has been located in County Cork, 6.5 km north of the town of Mallow, in an area where no mining had been carried out prior to the 1950s. This deposit lies at about the southern margin of the Central plain and is contained in basal Carboniferous limestones; the rocks overlie the sequence of Old Red sandstone rocks and are on the northern flank of the Kilmaclenine anticline. Here, the Old Red rocks are made up of interbedded red beds, shales, and quartzitic sandstones. In turn, the basal carbonate rocks are covered by the Transition beds that include shales, sandstones, and calcareous sandstones that are from 36 to 45 m thick; next above are the Lower Limestone shales (shaly and oolitic calcarenites, 37 to 50 m thick); next the Ballyvergin shale only 1.0 to 1.2 m thick; followed by the Tullacondra limestone, 35 m thick, and the Kilmaclenie limestone (60 m thick) and composed of pale massive calcarenite. All of these beds above the Old Red are Tournaisian, older than most of the mineralized beds in the mines of the Central Plain. Mineralization is found from the Transition beds through the Tullacondra limestone. In the surrounding area, the ENE-trending Kilmaclenine anticline is the principal structure, and it is cut by thrust faults parallel to its axis and by later north-south dip-slip faults - these structures all are Hercynian in age.

Despite earlier unsatisfactory exploration efforts in the Mallow area, a major program in 1972 found more or less coincident geochemical and geophysical anomalies in that region. On the surface, weakly disseminated chalcopyrite (0.18% Cu) was found in the steeply dipping basal Carboniferous limestones. Later diamond drilling showed a body of about 4 million tons that contains 0.7 per cent copper with about 20 grams to the ton of silver; of these 4 million tons, a continuous block of about 1 million tons has a grade of about 1.0 per cent copper.

The Mallow mineralization has been followed along the strike for about 380 m and, from the sub-outcrop, has been followed downward for 150 m. The ore minerals are chalcopyrite, bornite, chalcocite, and tennantite; these are as fine disseminations and veinlets in zones parallel to the bedding, normally in shales and shaly limestones. The mineralization also occurs in narrow, nearly vertical calcite-filled vein systems. A northerly trending normal fault cuts off the western end of the deposit. It would appear that the deposit is not economic under present economic and technological conditions.

In County Westmeath, 105 km west of Dublin, a lead-zinc deposit (called Ballinalack) is 23 km northwest of Mullingar, on the road from that town to Longford; it is some 15 km slightly north of east from Keel. The area has no previous history of mining but was found by a considerable program of exploration carried out by Syn-gen-ore Explorations, Ltd. The first clue to the possibility of ore in the area appears to have been some pyrite-bearing boulders excavated from adjacent drainage ditches. In 1969, however, reconnaissance and then detailed geochemical soil surveys confirmed base-metal anomalies in the 3 to 20 m cover of glacial till. Later induced polarization studies produced other anomalies that were not in the same locations as the geochemical ones. Drilling on the IP anomalies was useless, but such work on the geochemically favorable areas cut the shallow-lying sub-outcrop of the mineralized area. Further drilling outlined the main mineralized zones down to 275 m. Unofficially, the deposit may contain as much as a few million tons of a combined grade of about 8 per cent in lead and zinc.

The lower Carboniferous limestones, shales, and sandstones that contain the mineralized bodies are unconformable on the lower Paleozoic basement, the uppermost rocks of which probably are Silurian. The stratigraphic sequence above the Silurian basement consists, from bottom to top, of lower and upper Mixed beds, lower Bioclastic limestone, and Reef limestone, the last marking the uppermost Tournaisian. The Visean above is made up of the upper Bioclastic limestone, Calp limestone, and Lough Derravaragh cherts.

The lower Paleozoic shales are strongly folded. The lower Carboniferous beds that are hosts to the ore are gently folded by post-depositional Hercynian deformation with the main structure being a north-south fault with a maximum vertical displacement of more than 180 m. This fault may have been active during the deposition of the Reef limestone; this concept has been put forward for some of the other Central Plain deposits but is not firmly proved for them. The ore mineralization is located on the down-throw side of the fault, and there seem to be no good reasons to assume that it did not do much to control the location of the mineralization.

The deposit is in somewhat diffuse lenses that are strata-bound within the Reef limestone. The main ore minerals are sphalerite and pyrite with some lesser fine-grained galena; the gangue minerals are barite, calcite, and dolomite. These minerals fill fractures and cavities in the Reef rocks and in places replace that bed. This much evidence indicates that the ores, though strata-bound, are epigenetic to the bed that contains them.

The six deposits of the Central Irish Plain (Ballinalack, Mallow, Aherlow, Keel, Ballyvergin, and Abbeytown) that have been briefly described here have been so cursorily treated because: (1) the lack of literature about them does not permit them to be covered at the length that is made possible for Silvermines and Tynagh by the extensive studies that have been published about them and (2) because none of the six is of present economic importance. There remain two mines in the Central Plain that would be worth describing at length if the literature about them were sufficiently abundant and specific. These two are Navan and Gortdrum. The lack of printed material about Navan is largely due to legal problems that have constrained Tara Exploration and Development Ltd. from permitting much information to reach the scientific journals. The paucity of the printed papers about Gortdrum appear to be due mainly to its short productive life. They will be described below in such detail as it is possible to obtain.

The Gortdrum mine is located in County Tipperary, 5 km north of the town of that name (52°29'N, 8°10'W) and about 15 km nearly due north of Aherlow. In this area covered by 3 to 6 m of drift, the underlying bed rock is underlain by a gently folded series of lower Carboniferous limestones and shales that have been appreciably faulted. Before the discovery of the ore body, the surface expression was a low hill in which a quarry showed small quantities of copper sulfides (0.1 to 0.5 per cent copper). The underlying ore body was found by geochemical and geophysical prospecting by a subsidiary of Northgate Exploration, Ltd. The ore body is contained on the south limb of the east-trending Pallas syncline and is directly east of the Limerick volcanic basin in which are major amounts of lower Carboniferous (Visean) volcanics. The ore body is fault-located, this Gortdrum fault being a steep-dipping, ENE-trending structure that Williams and McArdle say has transcurrent movement, that is, it is a major strike-slip fault, the fault surface of which is steeply inclined. The result is that the Tournaisian limestone-shale sequence on the north side is downthrown against Devonian quartzites and thin red beds. The Devonian beds involved are some 270 m thick; above them lie various lower Carboniferous strata: Transition beds - 30 m; Dark limestone - 21 m; Ballyvergin shale - 3 m; Pale limestone - lower 27 m, middle 110 m, upper 88 m; and Waulsortian Reef Complex - 150 m. The Tournaisian rocks are conformable on, and grade into, the Devonian beds beneath. The Waulsortian beds outcrop several kilometers west of the mine as does the Visean volcanic group farther west.

The mineralization was found to descend to depths of 75 to 150 m, and it is confined to the Dark and Pale limestones, but these rocks are irregularly veined and replaced by copper-bearing sulfides. All of the ore is within 100 m of the Gortdrum fault with the ore nearest the fault having the highest copper content; the length of the ore along the fault is 600 m. The ore-enclosing limestones have been considerably altered, being recrystallized, dolomitized, and having veins of carbonate and barite. The ore-bearing sequence is cut by irregular vertical dikes and networks of fine-grained intrusive breccias (known as the Buff Alteration Zones). The ore body, in plan, is in two separate parts, eastern and western; each is a rough oval, and they are connected by a narrow band of mineralization. The west half is quite broad but is less than 100 m deep; the east part is narrower and is 150 m deep. The deeper portion is in a down-faulted wedge of Dark limestones in which the copper grade is between 2 and 5 per cent. The cut-off on the Gortdrum fault footwall is quite sharp, but the hanging wall of the ore is determined by assays.

The ore is low in pyrite with the Dark limestones having disseminations of bornite and chalcocite, whereas the overlying Pale limestones are more competent and contain their ore in stringers and veins of chalcopyrite and tennantite. The tennantite holds unusually large amounts of mercury, arsenic, and antimony, but the other, tennantite-free portions of the ore are quite lacking in these minor elements. Copper sulfides not only replace the limestone but also the intrusive breccias. Williams and McArdle contend that the structural control, the wall-rock alteration, and the replacement textures show that the ore was emplaced epigenetically. They report that the temperatures of ore formation probably were less than 200°C, determining this by the thermal stability of the mercury- and silver-bearing minerals; this estimate may be somewhat low as the sulfosalt is arsenic-rich tennantite and not tetrahedrite.

When the drilling at Gortdrum, begun in 1963, was finished in 1966, 3.8 million tons of ore had been blocked out that averaged 1.19 per cent copper and 23 grams of silver per ton. The ore body was of such shape that it could be mined by open-pit methods. Concentration commenced in late 1967, and the concentrates produced were trucked to the port of Limerick and shipped from there to a Spanish smelter. When production was initiated, the firm found that metal grade, particularly that of silver, was appreciably higher than the original estimates, but arsenic and antimony also were higher than this made the concentrates quite difficult to smelt. The level of mercury in concentrates produced from certain volumes of the ore also were so unacceptably high that they could not be sold until a mercury roasting-plant was constructed, and the mercury, partly as cinnabar and partly in solid solution in the tennantite, was recovered. A total of nearly 7500 flasks of mercury was produced over the life of the mine. Also over the nearly nine-year life of the mine, the copper concentrates averaged almost 900 grams of silver per ton. In the final year of mining, the copper grade rose to 2.01 per cent as the deepest benches were extracted. There remain nearly 650,00 tons of ore with a grade of 1.56 per cent copper that cannot be extracted by open-pit methods, and the company believes that this grade, even though arsenic, antimony, and mercury contents are low, is not high enough to make underground mining profitable.

The ore body at Navan is one of the major zinc-lead deposits of Europe; it was discovered by Tara Mines, Ltd. in 1970 although no record of mining within 8 km of the site has been recorded. The deposit is about 45 km northwest of Dublin, and the surface over the ore body is crossed by the Blackwater River. The property has been divided along this line, and legal problems as to the exact ownership of the ore body, while not preventing mining south of the river in the Tara-owned portion where mining is by underground methods, has delayed, at best, publication of the details of the geology of the Navan ore. It is planned to mine the ore north of the river (belonging to Bula Ltd.) by open-pit; by mid-1980 at least, mining had not commenced in that part of the ore body.

The Navan ore body is located south of the prominent Longford-Down massif that consists of lower Paleozoic sediments and volcanics; these rocks were both deformed and cleaved during the Caledonian orogeny. The southern flank of this massif is overlapped by a sequence of lower Carboniferous sedimentary rocks that dip gently toward the south and southwest. The Carboniferous strata are overlain in the area by outliers of Namurian (immediately younger than the Visean) shales, and there is an impressive Namurian-Triassic outlier north of Navan and within the lower Paleozoic massif at Kingscourt (also an area of minor ore mineralization); this massif is bordered on one side by a north-south fault, and such faults are common in the area. Numerous Tertiary dolerite (diabase) dikes and sills are known in the Navan area.

When exploration began at Navan in 1969, it was based on the concept that the most favorable location for such activity was in lower Carboniferous limestones near faulted contacts with Devonian and older rocks. Soil samples taken in that year around Navan outlined a definite zinc-lead anomaly, and some poorly mineralized limestone was exposed during this exploration. Further samples on a tighter grid marked out a southeast-trending area in which the zinc contents were as much as 25,000 ppm. Induced polarization studies were made that reinforced the soil-sampling results and defined more accurately targets for drilling, that began in November of 1970. By 1972, Tara Exploration reported reserves on both sides of the Blackwater River to be 70,000,000 tons with a grade of 10.9 per cent zinc and 2.63 per cent lead. No mention is made of silver in the galena, which may be because of legal problems or because Navan appears to be the only one of the major zinc-lead mines now being, or recently having been, worked in Ireland (Tynagh, Silvermines) that truly seems to be of the Mississippi-Valley type. The Bula organization had their property north of the Blackwater examined by RTZ Consultants, Ltd. and then reported their reserves to be 19.6 million tons with a grade of 6.7 per cent zinc and 1.2 per cent lead, appreciably lower grades than those announced by Tara, if they can reasonably be compared.

The oldest rocks in the Navan stratigraphy (Williams and McArdle, 1978) are undifferentiated lower Paleozoic beds of unknown thickness that consist of strongly colored shales and intrusives that are syenite or andesite. Unconformably above these are the Devonian (perhaps) and Carboniferous (certainly) Red beds to that up to 50 m thick and are made up of quartz-pebble conglomerates, red sandstones, and red shales. Over these are the Tournaisian Laminated beds, some 40 m at a maximum, that are composed of laminated argillites, argillaceous limestones, siltstones, and sandstones. The higher the position of these beds in the sequence, the greater will be the proportion of limestone. The succeeding beds are those of the Muddy limestone that is composed of bioclastic muds and is about 20 m thick. Next in the conformable sequence come the Pale beds (the hosts of the ores); they are light, sparry calarenites and oolites, plus minor quantities of argillaceous material. Where the lenses of ore occur, they normally are surrounded by dolomite, and the ore-gangue also has been dolomitized. The Tara geologists have found a series of shaly and sandy markers within these beds. The succeeding Shaly Pale beds consist of shales, argillaceous calcarenites and oolites with locally calcareous sandstone members. These beds are found on both sides of the Blackwater. Although generally eroded from the area of the Navan deposit, Reef limestones are found to the west and northwest and are made up of typical Waulsortian facies (micrites and biomicrites). Over the Reef beds are the Boulder conglomerates. Locally, the Reef beds overlie the Pale beds south of the river and demonstrate that an important erosional period affected the Navan region during the lower Carboniferous. These conglomerates are of the polymict limestone type and, in places, have a shale matrix; they are the youngest Tournaisian rocks in the area. The oldest rocks of the Visean are the Upper Dark limestones in which thin-bedded calcarenites are interbedded with black shales and graded pebble conglomerates. The Upper Pale beds are up to 150 m thick. The minor Tertiary dolerite (diabase) sills (previously mentioned) and dikes cut this rock sequence.

The local Navan structure is that of an anticline that plunges gently to the southwest; the anticlinal core is composed of lower Paleozoic rocks. The Navan ore body is located on the northwest flank of this structure. The ore-containing Pale beds (and those Tournaisian rocks associated with them) dip south and west at 20°; these beds are unconformably overlain by the Visean rocks. The folding of the Pale (and accompanying) beds is gentle, but two normal faults (C and B) and one reverse (A), cut the Tournaisian sequence. These normal faults stop against the unconformity (so thus are pre-Visean); the reverse fault (A) does penetrate the overlying Visean rocks; it also cuts through fault B at depth.

The ore bodies make up a series of stacked lenses in the Pale beds, the lenses dipping 20°SW. The axial trend of the ore lenses is to the northeast where the lenses join to form one ore body with a thickness greater than 60 m. Correlation of the various lenses south of the Blackwater has been greatly simplified by the recognition of the marker beds mentioned above. On the north side of the river, the best mineralization is contained in a dolomitized micrite unit at the base of the Pale beds (the Lower ore body); somewhat higher in the sequence is patchy mineralization known as the Upper ore body.

Throughout the area of Navan ore development, an irregular envelope of dolomite surrounds the ore. The major ore minerals are sphalerite and galena (Zn:Pb = 5:1) lesser quantities of marcasite and even lesser of pyrite are present; the gangue minerals are dolomite, barite, calcite, fluorite, and silica. Only traces of silver are in the galena and of cadmium in the sphalerite. The ore ranges from massive to fine-grained disseminations, and the sphalerite shows colloform, banded, and crustiform textures, particularly where it is accompanied by pyrite and galena. The lesser amounts of galena occur as small dispersed grains with sphalerite. Pyrite and marcasite usually have banded structures, but these minerals may be in free grains in or with sphalerite. In the sections of massive sulfides banded and colloform textures can be seen; these sulfides may have been brecciated and cemented by later sulfides. In the sections that contain disseminated sulfides, the ore is in narrow ramifying veinlets (Williams and McArdle, 1978).

Williams and McArdle (1978), whose work is practically all that is in print in regard to the Navan, think that too little is known about the deposit to reach a firm conclusion as to how it was formed, say that they have had no opportunity to examine large exposures of the ore body. They do point out, however, drill-core studies have given reason to support both syngenetic and epigenetic concepts of ore genesis. It is certain that the ore is strata-bound in that it is all contained in the Pale beds and that the fine-grained sulfides, located in brecciated bands, do not appear to have been fed through entry channels as an epigenetic mode of genesis would require. These factors argue in favor of syngenetic manner of emplacement. On the contrary, the intensity of mineralization and its marked change in character from one side of a fault to another, the envelope of dolomite surrounding the ore, and the numerous ramifying veinlets of ore sulfides all point to epigenetic emplacement of the ore and gangue minerals. After the IGC in Paris, I visited the Navan mine and am still not in a position to be dogmatic about how its ore was emplaced. I refer the reader to the appended Table I in this discussion that shows that I am convinced that all other Irish base-metal deposits in the Central Plain were epigenetically introduced into the rocks in which they now are contained. This is hardly evidence, since at least two of those listed in that table (Silvermines and Tynagh) have been firmly classed as syngenetic by several workers who have studied in each in detail. From my visits to each of these three mines (Navan, Silvermines, and Tynagh), I am convinced that each is epigenetic. In further support of this last statement, I am as certain as any ore geologist can be that the other seven deposits in the appended table are epigenetic. It would, therefore, be surprising if the three largest (Navan, Silvermines, and Tynagh) would have been syngenetically emplaced whereas only the small ones were epigenetic. Incidentally, Field Excursion 087 of the 26th IGC visited Navan, Tynagh, (and Gortdrum), yet the guide to these deposits has essentially the same discussion of Navan (translated into French) that was published in Volume I, Northwest Europe of Mineral Deposits of Europe. Since the latter was published in 1978 (and the included articles written much earlier), it is surprising that the guide to Navan was not, in any real sense, up-dated from that in Northwest Europe.

BASIC CHARACTERISTICS OF ORES IN THE LOWER CARBONIFEROUS LIMESTONES OF THE CENTRAL PLAIN OF IRELAND

Deposits	Ore-Containing trata	Tonnage Estimates of Reserves and/or Production	Grade of Reserves and/or Production				Manner of Ore Emplacement	Strata-bound or Stratiform
			Zn	Pb	Cu	Ag		
Abbeytown	Visean (Dinantian)	1,000,000[1]	2.5	1.5	---	±300gm/ ton	epigenetic	Yes
Aherlow	Tournaisian	5,400,000	---	---	0.89	41gm/ ton	epigenetic	No
Ballinalack	Tournaisian (Reef Ls)	few million	8.0 (combined)		---	?	epigenetic	Yes
Ballyvergin	Tournaisian	150,000	---	---	1.0	?	epigenetic(?)	Yes
Gortdrum	Tournaisian (Dark & Pale Ls)	3,800,000[2]	---	---	1.25[3]	30gm/ ton+	epigenetic	Yes in a broad sense
Keel	Tournaisian	1,800,000	7.0 (combined)		---	?	epigenetic	Yes in part
Mallow	Tournaisian (Courceyan)	4,000,000[4]	---	---	0.7	20gm/ ton	epigenetic	Yes
Navan	Tournaisian (Pale Beds)	70,000,000+	10.9	2.63	---	minor	epigenetic or syngenetic	Yes
Silvermines	Devonian (Old Red Ss) Tournaisian (upper) Visean (lowermost)	11,000,000+[5]	8.9	2.8	0.3	minor[6]	epigenetic syngenetic (in part ?)	Yes in a broad sense
Tynagh	Tournaisian (uppermost) Visean (lowermost)	9,400,000[7]	3.2	3.0	0.3	27gm/ ton	epigenetic	Yes in a broad sense

[1] total production (est.); [2] production thru 1975 3,350,000 tons; [3] est. grade, plus 7500 flasks Hg; [4] outlined by drilling, 1,000,000 tons at 1% Cu; [5] almost certainly appreciably more has been added; [6] plus minor Cd; [7] now shut-down, production probably was in excess of this tonnage

At least five authors (Grieg, and others, 1971; Morrissey, and others, 1971, Taylor and Andrew, 1978, Evans, 1976, and Boast, and others 1981) have discussed the bearing of sulfur-isotope analyses on the manner of genesis of the ore deposits of the Central Plain of Ireland. The work of Greig and his colleagues was done on samples from the Silvermines property, specifically the Lower G and Upper G ore bodies. The delta-^{34}S values for the sulfides range from +4 to -7°/$_{oo}$ in the lower ore body and from -8 to -35°/$_{oo}$ in the upper one. They observed a pronounced trend in delta-^{34}S values in that the sulfides become progressively lighter upward along the Silvermines fault and away from the fault in the stratiform ore of the Upper G ore body. Barite, which occurs above, and at the far end of, the stratiform ore body, gives consistently delta-^{34}S values that all are near +18°/$_{oo}$. These authors believe that the trend in delta-^{34}S values that they observed, as well as the presence of barite in the marginal areas of the Upper G ore body, can best be explained by an increase in pH and Eh of the hydrothermal fluids through reactions with the host carbonates. They believe that, as the hydrothermal fluids moved upward and outward away from the fault, reduced sulfur species (in relation to oxidized ones) became less abundant and the delta-^{34}S values of the reduced sulfur became more negative. On the contrary, the oxidized sulfur species became more abundant and more positive. These authors achieved similar results from samples taken from the Tynagh and Gortdrum mines.

Morrissey and his colleagues say that the results on the two G ore zones at Silvermines are in agreement on the following points: (1) the delta-^{34}S values of sulfides in the upper, stratiform ore zone are uniformly negative and show a wide spread, whereas those in the lower, discordant zone have an appreciably narrower range and are both positive and negative. Morrissey and his co-workers remark that Graham (who did a thesis on the Silvermines ore at the University of Western Ontario but has not otherwise published his results to my knowledge) explained the coincidence of the delta-^{34}S values in barite and those of Carboniferous sea water by assuming that the sulfate in barite did indeed come from that source whereas Greig and his colleagues thought that the coincidence of barite sulfate and sea-water sulfate was just that, the delta-^{34}S values of the barite sulfate having been the result of a long series of oxidation reactions on sulfide sulfur as the hydrothermal ore fluids in which it was contained moved upward through, and outward from, the Silvermines fault. Greig and his colleagues demonstrated that the observed levels of enrichment of light and heavy sulfur, respectively, in the sulfides and barite of the upper zone could have resulted from abiologic reactions between rock carbonates and hydrothermal solutions with an initial delta-^{34}S value of about 0°/$_{oo}$ and a starting temperature of about 250°C. Morrissey and his co-workers believe that Greig's concept is the more reasonable, particularly when it is applied to Tynagh and Gortdrum as well because the rocks of these two deposits do not confirm the idea that large quantities of sulfides had been remobilized in them.

Taylor and Andrew (1978) examined this same evidence, plus an unpublished report by Coomer and Robinson, 1975, and reached rather different conclusions. These are: (1) that the sulfur came from two sources, lower Carboniferous sea water and a deep-seated source; (2) that the sulfides in the Devonian and lower Tournaisian rocks were precipitated from deep-seated solutions at about 150° to 300°C with delta-^{34}S in the range of +4 to -8°/$_{oo}$; (3) barite in the stratiform and Shallee ore bodies was precipitated directly from sea water and the sulfur in the pyrite in the stratiform ore zones was formed by biogenic reduction of sea-water SO_4^{-2} during early diagenesis; and (4) galena and sphalerite generally precipitated later than barite-pyrite-siderite at temperatures between 120° and 200°C. This last (4) statement certainly raises more questions than it answers. By implication, Taylor and Andrew indicate that the metals came in to the Silvermines space volumes as chloride complexes; how the deep-seated sulfur got there, they do not consider. This problem, as it concerns Silvermines in particular, is considered further in the discussion of that deposit.

Evans (1976) essentially agrees with Greig and his colleagues, accepting their explanation of why very different values between sulfate and sulfide sulfur do not require the complicated (and unreal) hypothesis that the sulfur had two different sources.

Boast and his colleagues (1981) base their study of the stable-isotope evidence as it bears on the genesis of the Tynagh deposit on the assumption that the various ore and gangue sulfide/sulfates were deposited syngenetically. They are convinced, despite the calculations of Greig and his co-workers, that the sulfate in the Tynagh barite was obtained from sea water and that the sulfide ions in the various sulfides of their second stage were produced by bacterial open-system reduction of sea-water sulfate. In their third stage, however, reduced sulfur from a deep-seated source was mixed with bacteriogenic sulfide migrating into bank limestones to precipitate stage 3 sulfides and sulfosalts. All this neglects the point made by Williams and McArdle that the fossils that are closely associated with the ore sulfides are normal in size and morphological characteristics; had the sulfide sulfur been formed by the bacteriogenic reduction of sulfate sulfur, the fossils would have been stunted and deformed. Yet the entire fossil assemblage at Tynagh shows every evidence of having been formed under perfectly normal conditions. This point is considered at greater length in the discussion of Tynagh.

That the interpretation put on their sulfide-isotope data by Boast and his colleagues can be explained in a different manner than they have done can be seen from a study of the work of Greig and his co-workers. Certainly, the conclusions of the Boast group cannot be considered as firm as they believe they are.

Certainly, the genetic story of the ore deposits of the Central Plain of Ireland is well summarized by Evans (1976). He states truly that the original interpretation of the manner of formation of these deposits was that they were of magmatic-hydrothermal origin. This definitely was the thought of Rhoden on Silvermines, work done before the newer Central-Plain deposits were discovered. These discoveries occurred shortly after the publication of Oftedahl's paper (1958) on the volcanic-exhalative manner of ore formation that attracted instant acceptance by many ore geologists. As the Central-Plain deposits were largely (see attached table) strata-bound or stratiform, it was easy to see them as syngenetic with the sediments that enclosed them. Today, the synsedimentary hypothesis has lost much of its former popularity [despite the recent (1981) paper of the Boast group] because of the clearly demonstrated epigenetic character of most of the mineralization in this area. It no longer seems possible or reasonable to explain every epigenetic feature by remobilization of syngenetic sulfides, particularly when it is remembered that the Central-Plain rock suffered only minor deformation and metamorphism. Further, the Irish deposits are quite similar to those in the Variscan foreland in Great Britain and northwestern Europe.

Evans emphasizes that the stratiform character of many of the Irish deposits is due to several factors other than a possible syngenesis of the sulfides with the primary sedimentary minerals of the rocks in which they are contained. These are: (1) the permeable nature of arenaceous and conglomeratic beds intersected by feeder faults, (2) the secondary permeability produced by dolomitization of the originally limestone host rocks, (3) the ponding of rising ore solutions, and (4) the marked preference of the ore fluids for certain stratigraphic horizons. Even where such factors did not obtain, the ore zone is, in some deposits, radically discordant. Further, to repeat, the normal character of the fossils in the various ore-containing beds, argues against their having been developed in an environment polluted with bacterially generated hydrogen sulfide.

If neither the magmatic-hydrothermal or the synsedimentary hypotheses are to be accepted, what concept should be? Present day opinion appears largely to favor the polygenetic hypothesis by which metal-bearing solutions (the metals in chloride complexes) rise of depths and encounter large volumes of essentially quiet waters in which hydrogen sulfide is being bacterially produced, the interaction of these two fluids is presumed to have resulted in the formation of the ore sulfides. The various difficulties of producing the rather complex mineralization of several of these Irish deposits by this method of rapid precipitation are neglected.

These rising fluids with the metals in chloride complexes are assumed to have obtained their chloride content as they worked their way downward through the crust of the earth to depths (upwards of 10 km) where they can have been heated to high-enough temperatures to have picked up metals on their way up (and perhaps during the latter part of their downward journey). Also neglected is the apparent shortage (not to say general lack) of chloride-bearing evaporite beds where the downward moving meteoric waters could have obtained the huge amounts of chloride ion they need. Surely, a more reasonable mechanism for the acquisition of the needed chloride ion would be from crystallizing magmas. Chloride ion is not readily acceptable in more than a few magmatic minerals, these never are common in igneous rocks. Thus, any ore fluid of magmatic origin could be expected to contain an appreciable content of chloride ion (at least 5 per cent) and, if boiling of the ore fluid occurred, much more. As I have said elsewhere in this volume, I think that the chloride ions have as much to do with the ore-forming process as the passengers in an aircraft have on its ability to fly. The main purpose that the chloride ions have is to supply salt to the sea; a supply that is kept relatively constant by the precipitation in favorable localities of evaporite beds.

This brings us to the sulfate ion in the barite. I have approached this problem in a somewhat different manner than Greig and his colleagues, but the result, I think, is essentially the same. The sulfate ions that unite with barium ions to form barite are, in my opinion, the result of the greater concentration of ^{34}S than of ^{32}S in the sulfide ion because ^{34}S is more readily oxidized than ^{32}S. It probably is not coincidence that the delta-^{34}S concentration in the sulfate in ore fluids and in the sulfate of sea water is essentially the same because much of the sulfate ion in sea water probably is produced by reactions undergone by sulfide ion in magmato-hydrothermal ore fluids on their way to the surface. The remainder of the sulfate ion in the sea comes from the oxidation (in the weathering process) of sulfide ions in sulfide minerals exposed at the surface. Apparently this process fractionates the sulfur in much the same way as it is fractionated by the reactions carried out in the ore fluid. Thus, it is not necessary to appeal to one (or two) sources for sulfide ion and an entirely different one for sulfate.

The polygenetic mechanism would obtain the metals for the ore fluid by leaching them from the wall rock through which it passed on its long journey from the surface to depths of up to 10 km and back to the surface. Yet, wherever rocks are exposed from which it is suggested that the metals in the ore fluids came, they are not found to show any evidence of leaching. The metals in ore deposits are exactly those that either are largely unacceptable in silicate minerals (Zn, Pb, Ag, Au, Hg) or are so abundant in the parent magmas as to have them (Fe, Cu, Mg, Ca) left over for incorporation in the ore fluid developed in the late stages of magma crystallization. Thus it is much more simple and much more reasonable for the ore metals to have been given off from a magma in the late stages of its crystallization than for them to have been picked up from the rocks through which the mildly heated ore fluid has travelled. At the temperatures to which a descending ore fluid can be heated, particularly one that deposits telethermal type ores, the chances of that fluid being able to dissolve any appreciable quantity of such elements is small.

Although the 10 Irish base-metal deposits in the Central Plain (those included in the preceding table) often have been referred to as Mississippi-Valley-type deposits, most of them do not fit in that category. In the first place, only six of the ten deposits listed in the attached table are zinc-lead or zinc-lead-copper deposits. Secondly, four of those six certainly are too high in silver to fit into the Mississippi-Valley type, and the other two (Ballinalack and Keel) probably contain too much silver to have been deposited within the telethermal temperature range, although the necessary data are not available to confirm this. Of the four copper deposits that do not have enough zinc and lead for the contents of the elements to be reported (Aherlow, Ballyvergin, Gortdrum, and Mallow), no one is appreciably rich in copper with Gortdrum (1.25% Cu) being the only one that was mineable and that to some extent because of its silver and mercury contents. In the mineralization at Silvermines and Tynagh (the two zinc-lead-copper deposits) (Silvermines certainly was not mineable because of the copper it contains), the copper minerals do indicate that they were deposited at temperatures above the telethermal range. Of the zinc-lead deposits in the Central Plain, only two (Ballinalack and Keel) may be of the Mississippi-Valley type, but more data are needed before this statement can be confirmed.

As for the age of the deposits, lead-isotope data suggest that they were deposited over a wide range of ages, but the difficulty of being certain what lead-isotope data mean as to ages, these suggestions cannot be, as yet, substantiated. All of the ore bodies are entirely or mainly (Silvermines) in lower Carboniferous rocks, and their mineralization probably was related to the Variscan orogeny. They here are classified as Late Paleozoic.

The mineral suites present in those deposits that have been studied in depth definitely indicate that they were formed at moderate temperatures (i.e., in the mesothermal to leptothermal ranges) and they are categorized here as Mesothermal to Leptothermal with the proviso that it may be necessary to include telethermal in the classification of a few of the deposits, though not for the three that are disscussed in detail in what immediately follows.

AVOCA, COUNTY WICKLOW

Middle Paleozoic	Copper, Pyrites minor Zinc, Lead	Mesothermal

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Ridge, J. D., 1973, Volcanic exhalations and ore deposition in the vicinity of the sea floor: Mineral. Dep., v. 8, p. 332-348

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Notes

The copper-pyrites deposit of Avoca is located in the southeastern part of County Wicklow near the village of Avoca (52°52'N, 6°13'W); the mine is slightly over 70 km south of the city of Dublin and is in one of the most beautiful areas of Ireland. Avoca is only 8 km from the seaport of Arklow from which location concentrates can readily be shipped abroad.

The valley of the Avoca River is the site of road and rail lines and follows the major north-south fault that cuts the ore bodies at Avoca into east and west sections.

Avoca is the only operating mine in the Caledonide terrane of Ireland, and the mine has been operated intermittently for well over 200 years. Since 1971, however, the mine has been operated continuously and has produced in that time over 6.5 million tons of 0.8 per cent copper ore (Sheppard, 1980). The overall production during the entire history of mining at Avoca is estimated by Downes and Platt (1978) as 15 million tons with a grade averaging 1.2 per cent copper. On the contrary, Williams and McArdle (1978) state definitely that, before World War II, total production of ore from Avoca (mainly from higher-grade massive sulfides) was nearly 4 million tons. To this must be added the 6.5 million tons reported by Sheppard to give a total of about 10.5 million tons of ore that should have had an average grade of appreciably above 1.0 per cent; since, however, no grade is given for the 4 million tons, no exact average can be calculated.

Despite its more than 200-year history, interest in publishing the results of such geological studies as have been carried out on the property has been small until the last less than four years. Of the 10 references listed in the bibliography for Avoca, eight have been published no earlier than 1977. This is not to denegrate the contribution of M. V. O'Brien in the early 1950s who not only awoke interest in Avoca but also in the lower Carboniferous limestones of the Central Plain of Ireland where the bulk of the country's mineral wealth (so far as it now is known) is situated but to make certain that credit is given where it is due.

In 1955, over two centuries after the first ore had been discovered in outcrops along the banks of the Avoca River, mining had been extended for some 4 km along strike. Into this area, exploration teams of the Mianrai Teoranta

[Mineral Treasure (of Ireland)] examined old workings and carried out extensive diamond drilling, on the basis of which they estimated that the ore reserves of the West Avoca area only were 14,000,000 tons of ore that averaged 1.12 per cent. In 1956, St. Patrick's Copper Mines, Ltd. conducted further evaluation work and raised the reserve total to 18,700,000 tons of copper ore at the same grade as the Mianrai Teoranta estimate. [Apparently the theory being that the good saint may have driven the snakes out of Ireland but left the ore minerals in place.] This estimate included 5,500,000 tons of near-surface ore in a zone of old workings. The St. Patrick's staff undertook to mine the near-surface ore first (presumably in an effort to generate cash flow quickly). Unfortunately, for many reasons, both projected grade and production goals could not be met. So, after seven years (in 1962) the mine was closed after recovering 2,850,000 tons of ore running 0.74 per cent copper, plus a minor amount of pyrite sold as concentrates. Toward the end of 1966, a Canadian group began a new set of evaluation studies that produced an estimate of 6,500,000 tons of copper ore that assayed 0.89 per cent Cu. The present ownership of the property is complex, but basically it is Canadian. Under this new management, the first ore was delivered to the rehabilitated mill in late 1970. At present, two concentrates are made, the copper concentrate being shipped from the port at Arklow and the pyrite one being used locally in the production of fertilizer. Although mining still continues (1980), the estimated 6.5 million tons of ore have been mined and the grade maintained at 0.8 per cent copper.

As has been discussed (see Ireland - General), the deformed volcanic rocks of southeast Ireland can be divided into three belts from south to north: (1) Portrane-Lambay-Donabate-HIll of Allen-Kildare inlier; (2) Donard-Baltinglass-Rathdangan-Aughavannagh in southeast Leinster and on to Borris in County Carlow; and (3) from Tramore in County Waterford to Arklow in County Wicklow. A subdivision of (3), designated (3a), is a volcanic tract north of Arklow; the Avoca mines are located in the center of this tract. The rocks of belt (3) probably form the core of a syncline and (3a) is in a subsidiary syncline.

In southeast Ireland, the lower Paleozoic rocks appear to have been deposited on a metamorphosed Precambrian basement. A small area of these Precambrian rocks is located on the Irish coast just south of Rosslare; elsewhere, in that part of Ireland between Waterford on the south and nearly as far as Dublin on the north, the volcanics mentioned in the preceeding paragraph are included in lower Paleozoic sediments; these volcanics are of Caradocian [middle to late Ordovician] age. The lower Paleozoic sediments were derived from a major source to the southeast and were deposited in a NE-SW oriented trough developed on the metamorphosed sialic Rosslare complex basement. These lower Paleozoic sediments are referred to by Platt (1977) as paratectonically metamorphosed rocks [i.e., low-grade metamorphic rocks (greenschist facies) that have steep cleavages and are contained in orogenic belts]. The upper Orodovician volcanics in this part of Ireland followed, after a period of no volcanism, an earlier mafic to intermediate volcanic activity. The second stage of volcanics was calc-alkaline and basaltic, andesitic, and rhyolitic in the Waterford area, but at Avoca, however, the volcanics of this age are rhyolitic. Pointon (1980) says that this volcanism in the paratectonic Irish Caledonides probably was of the Andean type and occurred near the margin of the European plate.

According to Pointon, the upper Ordovician volcanism in southeastern Ireland was directly connected with regional fracturing and the differential subsidence of small basins and the regional alignment of volcanic centers. In the Avoca area, a linear control appears to have existed that affected not only the volcanism but also later tectonic events. This direction is that of the NE-SW orientation of the local volcanics and is a response to this control. The presence, in the Avoca formation, of bodies of massive rhyolite, either as flows or shallow plugs, is assumed by Pointon to indicate that these were extruded close to the original vents. The NE-SW orientation of the rhyolites also is followed by later intermediate intrusives and the even later granite. This same trend governs the orientation of the axes of the regional synclinal folds in which the volcanics have been protected from erosion. Pointon is of the opinion that the regional cleavage, that is offset from this trend by as much as 15°, may have been produced by folding under shear stress during the Caledonian orogeny.

Williams and McArdle describe the Avoca volcanic belt as a distinctive unit of the Leinster Caledonides, the belt extending for 30 km from Coraghan Kinshella Mountain to the coast at Wicklow Head and paralleling the regional NE-SW cleavage. They consider the volcanics to be Ordovician silicic tuffs and lavas that have been intruded by minor granite and diorite sheets and by dolerite [diabase] sills. Within this belt is a distinct band of chloritized tuffs, in all places less than 300 m wide, that follows the NE-SW trend for more than 10 km and includes the well-defined zone in which the chalcopyrite-pyrite mineralization of Avoca is contained.

Platt (1977) says that the volcanics in the Avoca belt are massive sheets of felsic eruptives that contain interbedded sediments and pyroclastics, the deposition of which occurred under shallow marine conditions. He says that previous workers have classified the volcanic rocks of the Avoca belt as calc-alkaline and have considered them as having been formed in an island-arc setting. He believes this interpretation is compatible with recent models for the plate-tectonic evolution of the Appalachian-Caledonian orogen. In this evolution, it is suggested that a southeast-inclined zone of subduction along the southeast margin of the proto-Atlantic Ocean produced island-arc volcanism. This volcanism, during Ordovician time, was in an extensive, complex, and semi-continuous belt along the active southeast margin of this orogen. The belt developed through what is now the Scandinavian peninsula, down through the English Lake district, to southeast Ireland, Newfoundland, and New Brunswick and into the Appalachian belt of eastern North America. Paleozoic volcanism was highly developed in this belt, and it is there that the greatest number of lower Paleozoic deposits of economic value are to be found.

Platt points out that the geological understanding of the Avoca volcanic and mineral belt is complicated because the rocks and the ore deposits [that he thinks were contained in them during their metamorphism] have undergone several episodes of deformation in which the rocks were brought to the greenschist facies of low-grade regional metamorphism. The tectonic fabrics resulting from these processes have in many places effectively concealed or destroyed the primary mineralogy and volcanic or pyroclastic textures. According to Platt's interpretation of the literature, the Avoca belt is localized in a broadly synclinal structure - the Wicklow syncline - that at present is considered to be a tightly folded recumbent syncline that, in turn, has been folded into an antiform, the axis of which plunges gently to the northeast. This results in the outcrop of the structure having a horseshoe-like appearance in which the limbs trend NE-SW and the closure is to the northeast.

Platt reports that the volcanic sequence at Avoca is between 1400 and 2400 m thick. The volcanics are both underlain and overlain by slaty rocks with some limestones in the footwall of the beds. The rhyolites of the sequence are calc-alkaline and, at their base, are associated with appreciable magnetite. The rhyolite layers are rhyolite lavas and breccias overlain by a wide range of silicified, sericitized, and chloritizied pyroclastics; although the pyroclastics are schistose, still several varieties - crystal and 'quartz-eye' tuffs and porphyries, lapilli and lithic tuffs, coarse agglomerates and flow and slump breccias - are known. Within these pyroclastics are beds of non-volcanic mudstones and carbonaceous sediments; these are most commonly the lateral stratigraphic equivalents of the ore horizons. These carbonate sediments are highly dolomitized and carbonized and contain appreciable Mn-rich carbonate. The environment in which these carbonate beds were developed is considered to have been a back-arc basin.

From southwest to northeast, the ancient mines of Avoca are: Ballymurtagh, Ballygahan, Tigroney, Cronebane, Connary, and Kilmaco. The first, and probably the second of these are located west of a major fault that separates the Avoca mineralization into West and East Avoca. Only one of the ancient mines in East Avoca has been mined in recent years - Cronebane. This ore body was secondarily enriched and was mined in the early stages as an open-pit for its higher grade. In the West Avoca area, mining has been confined to the South and Pond lodes. These two mines are located immediately adjacent and parallel to each other with the South lode being sheet-like in form in the hanging wall of the Pond lode; both lodes dip to the southeast and at about 55°. The South lode, therefore, is actually to the south of the Pond lode. The western edge of the South lode plunges southwest at 30° and the eastern limit plunges at 60°, also to the southwest; the lode has been traced to a depth of 300 m and is about 20 m wide. It has been followed along strike for 1000 m and is concordant with the regional cleavage, 55°SE. The ore minerals are chalcopyrite and pyrite that are disseminated in quartz-chlorite schists that were silicic tuffs before their deformation and alteration. Veins of quartz that contain coarse stringers of sulfides cut through the altered volcanics; the accepted theory of the origin of the ores by volcanic syngenesis requires that these veins were the result of remobilization. The grade of the ore normally is below 0.8 per cent copper. The competent rocks in the hanging wall are unmineralized quartz-chlorite schists, whereas the footwall rocks are composed of less competent sericitic tuffs; these tuffs separate the South lode from the Pond lode.

The Pond lode also is a sheet-like body that is 6 m wide (and is parallel to, and in the footwall of, the South lode). It differs a great deal from the South lode, being composed of massive sulfides as alternating bands of chalcopyrite and pyrite; its grade usually is about 1.15 per cent copper. The contact between the ore and the footwall rocks is determined by assay with the massive sulfides gradually changing to disseminated ore minerals; as this process took place, the grade away from the massive sulfides dropped until the rock no longer is mineable. As the ore is followed downward, the proportion of massive to disseminated ore decreases; the disseminated ore is best developed at the southwestern end of the ore body . The hanging wall of the Pond lode is delimited by a narrow zone of friable graphitic schist, overlain by normally competent sericitic tuffs. The Pond lode has been followed along the regional strike for more than 750 m; its depth is about 750 m also. The dip of this lode is 55°SE and, at both ends, it plunges at 30°SW. At its northern end, it outcrops over a considerable area.

The South lode was mined underground by longhole open stoping, whereas the outcrop area of the Pond lode was mined from an open-pit in which selective mining kept the grade at about 0.88 per cent copper. The surface volumes of the South lode were mined long ago, and these have been kept separate from the underground workings by a major crown pillar. The underground (and main) mining of the Pond lode was first by cut-and-fill methods that later were changed to sublevel caving.

The Cronebane mine is at a highter elevation than those west of the Avoca fault; it is in the East Avoca area and is about 1.5 km from the mill. This ore body also follows the regional strike and extends for about 600 m in total distance. The Cronebane ore is in two zones: (1) a near-surface one composed of secondary plastic clays in which the grade ran 1.63 per cent copper and (2) a deeper zone of primary disseminated sulfides in which the grade is about 0.65 per cent copper. In the secondary ores, the higher grade was produced by the developemnt of chalcocite and covellite from the chalcopyrite; primary pyrite and sphalerite remain largely unaffected by the near-surface processes; the primary ores are made up of pyrite and chalcopyrite with quite minor sphalerite and galena. The secondary ore was concentrated in a boat-shaped body that grades sideward and downward into the primary ore. The rock containing the primary ore is competent chloritic tuffs of much the same character as those associated with the South lode; such tuffs also make up the footwall of this ore body. The hanging wall is composed of decomposed rhyolites.

Production at Cronebane began in 1971 with the easily mined but difficultly concentratable secondary ore (a separate mill had to be constructed to handle it). From late 1971 until 1973, this secondary ore was the only type mined at Cronebane. Gradually a transition was made from the mining of the secondary ore to that of the primary; by 1974 all mining was done by conventional cut-and-blast methods. The pit, however, was closed in 1975 because, with the low-grade primary ore, profitable grade ore could not be recovered. The entire production from Cronebane was nearly 575,000 tons with a grade of 1.41 copper, 0.97 per cent lead, and 1.43 per cent zinc.

Platt (1977) quotes Sangster as defining two types of ore as typical of volcanogenic ore deposits: (1) massive sulfide ore that is markedly conformable with the rocks that surround it and (2) stringer or disseminated ore that cuts across the stratigraphic column. Both types of ore, Platt reports, are known and mined at Avoca, and these he considers are comparable to two of the Kuroko types: (1) Oko or yellow ore, normally massive cupriferous pyrite, and (2) Keiko or siliceous ore, composed of cupriferous, silica-rich ore that is either in disseminated bodies or in stockwork veinlets. These two types should be described in more detail. Type (1) at Avoca is broken down by Platt (1977) into two subcategories, these are: (a) chalcopyrite-pyrite ore with minor galena and sphalerite that is described as having a banded, sedimentary appearance with the host rock being a fine-grained and chloritized and possibly slightly carbonaceous tuff. Layers of sulfides are separated, in many places, by appreciable distances of altered host rock; the total sulfide content may be no more than 15 to 20 per cent of the total of the whole rock, and (b) dense, fine-grained pyritic lead-zinc-copper ore with associated silver-bearing sulfosalts. This type (b) may include more than 90 per cent of total sulfides with the highest contents of lead and zinc being near the hanging wall. Platt points out that sedimentary textures are difficult to recognize in hand specimens of this type of ore.

The massive sulfides of both types have appreciable carbonate mineralization developed both laterally from the ore and in its hanging wall. The carbonate usually is nodular or oolitic dolomite. Platt thinks that both kinds of massive sulfide ore are [in the Pickwickian sense] integral parts of a single composite massive-sulfide ore body. The massive sulfide ore bodies have sharp contacts with the country rock, most especially in the hanging wall.

Of the two types, type (a) is stratigraphically above type (b), and the maximum thickness of both types (plus their included country rock) is about 21 or 22 m with the long dimensions of the ore bodies being parallel to the bedding planes of the country rock.

As for type (2) (stringer) ore, it lies under the massive type, and it (type 2) seldom has more than 20 per cent sulfides included in it. These stringer ore bodies at Avoca are located within the pile of major chloritized pyroclastics and are close to, or in contact with, the overlying massive sulfide ores. These stringer ores are made up of ramifying and braided veinlets of quartz and sulfides that cut across the bedded volcanics for many tens of meters. In the upper part of the stringer ore, the change to massive type (1) ore is quite sudden; this relationship is in stark contrast to the diffuse and gradational contacts of the stringer ore in its pipe-like bodies with the altered rocks beside and below them. Where the stringer ore has not been tectonically deformed, the pipes are cylindrical or funnel-shaped; these pipes or funnels have caps of massive sulfides that finger out laterally into carbonate-rich beds.

Features that are common to both types of deposits [(1) and (2)] are: (1) carbonate-rich gangues that, in many places, are dolomitic or manganiferous and are associated with shaly tuffs and carbonaceous shales; (2) introduction into a general area of felsic volcanic rocks of the calc-alkaline type - only locally are these interbedded with basaltic beds; (3) obvious emplacement in a volcanic pile that was generated by explosive volcanism related to dome-shaped centers - locally these volcanics may have been extruded subaerially; (4) association of the volcanic series with the prevalent clastic sediments; (5) occurrence of the massive sulfides in clastic sediments near volcanic rocks rather than in the volcanic rocks proper; and (6) compositional zoning of the ore bodies as the stratigraphic sequence is followed upward or downward - Pb/Zn decreasing downward and Cu decreasing upward.

Platt believes that ore bodies in such arrangements tend to be located in provinces or clusters, and at least two such units occur in the Avoca area (West Avoca and Cronebane). If Pb-Zn-Cu ore should be discovered in West Avoca, Platt believes that this would have important consequences for mineral exploration in the Caledonides of southeast Ireland.

Platt thinks that the designation of Avoca as a "copper" mine is what he calls an accident of history; recent detailed examination of the deposit, particularly on the east side of the Avoca fault, has shown that it is poly- rather than mono-metallic and that it bears more resemblance to the Kuroko-type ores than it would have been thought possible before the 1970s. Platt says that the deposit clearly was produced by volcanogenic processes. This well may be true, but it would be more accurate to say, until more evidence is in, that it is a deposit that certainly was formed in an area where volcanic porcesses were strongly developed and that almost certainly was derived from fluids given off by the same magma chamber that produced the silicic volcanics with which it is so closely associated in space and time. Exploration for other deposits such as Avoca in the Caledonides of southeastern Ireland (or anywhere else in the world for that matter) should seek for geologic environments similar to that surrounding the Avoca ore bodies. Exploratory work should not attempt to apply a single theory of formation for Avoca and any other deposit that may be similar to it in stratigraphy, structure, ore minerals, associated alteration, ore body characteristics in relation to the mineralogy, and structure of its host rocks. This applies no matter at what stage in the development cycle of the totality of the rock and ore sequence the ore may have been introduced.

The major point of difference between the authors listed in the references for Avoca and me is that I believe it possible that the ores were introduced into the sequence of structurally and mineralogically altered rhyolitic rocks after those rocks had been lithified and not either syngenetically or diagenetically. I think the same about the Kuroko deposits that I have had a greater amount of time to examine than I did those at Avoca. On the basis of these, admittedly cursory examinations, I am convinced that the evidence exhibited by the rocks in both areas does not eliminate the possibility that the ores were emplaced at a considerable period of time after the ore had been lithified and, at least to some extent in the case of Avoca, metamorphosed. It is true that the rocks at Avoca that contain the ores show some evidence that slumping occurred at some time during their geologic history, but this slumping has not been fully demonstrated to have occurred after the ores had been introduced. It is further generally agreed that the veins and veinlets of sulfides that cut the massive ore in the various Avoca ore bodies were the result of remobilization, yet no evidence has been put forward to show that such a process affected the Avoca ores or, if it did, how it managed to dissolve ore minerals in one place and redeposit them in another only a few meters or even centimeters away. Perhaps these authors believe that the minerals in these veins moved into position by plastic flow. These points will be discussed further after considerations of Platt's volcanogenic model and of Wheatly's ideas on ore mineral fabrics in the Avoca polymetallic ores (1978).

Platt's volcanogenic model is quite straightforward. The ores entered the area through a stockwork feeder or pipe-like body that cut through the local rhyolitic volcanics and volcaniclastics while they were still essentially horizontal. This feeder pipe was capped on the sea floor by massive sulfides deposited there at the same time essentially as the rhyolitic material was being extruded, from other vent(s) on the sea floor. Thus, the ores are interbedded with rhyolitic materials in their various forms in proportions dependent on the amounts of rock and ore materials being supplied to the area during the time when the ores were being introduced. Thus, as Platt points out, the massive sulfide body tongues laterally into clastic, carbonate-rich basinal rocks and itself is capped by shale and fine-grained volcanic sediments. In addition to the introduction of rhyolite and ore minerals, the normal processes of sedimentation supplied clasts and dissolved materials from the surrounding surface areas. The rocks of the pipe and of the pyroclastics containing the ores were silicified and chloritized by presumably the same solutions from which the ores were deposited. The brecciation of the basal rhyolite sheet of the volcanic pile probably occurred before the ores were introduced so that the breccia fragments now are cemented by sulfides, mainly massive pyrite. Of course, it remains possible that these breccias were cemented by remobilized massive sulfides long after their original deposition on the sea floor.

A considerable objection to the efficacy of Platt's model stems from data that I published in 1973. The presence of exsolution blebs of chalcopyrite in sphalerite and at grain boundaries strongly suggests that the sphalerite was emplaced at temperatures at least above 300°C and probably considerably higher (Wheatley 1978). Further, the diagram in Platt (1977, Fig. 3) and the rocks that were deposited concurrently with the volcanics indicate that the depth of the sea at the time of ore emplacement (granted the ore was poured out on the sea floor) almost certainly was not more than 300 m (1000 feet approximately). From these data of mine, it follows (Ridge, 1973, Table 2) that the ore fluid (assuming the NaCl content to have been 20 per cent) would have boiled at a depth of at least 100 m below the sea floor. This would have resulted in the deposition of a variety of copper, lead, and zinc chlorides (assuming that the commonly made assumption that the metals were being transported in chloride complexes) 100 m at least beneath the sea floor. No evidence of any such depositon of metal chlorides in such locations has been reported for Avoca (or any other Irish base-metal deposit). On the contrary, the depths of deposition in the East Pacific Rise or the Kuroko area appear to be measured in 1000s of meters, so it is unreal to attempt to compare the processes of formation of the shallow-depth Irish deposits (granted again that they were syngenetic in deposition) with such deposits as those of the Pacific Rise or the Kuroko Sea.

It appears extremely unlikely to me that, granted the Avoca ores were deposited syngenetically on the sea floor, metamorphism could have taken sphalerite deposited at temperatures low enough for the ore fluids to have resisted boiling and added enough chalcopyrite to that mineral that, on cooling after metamorphism, chalcopyrite would have exsolved from the sphalerite. Since Wheatly does not show plates of chalcopyrite from which the presence or absence of sphalerite blebs can be judged, conclusions cannot be drawn as to the temperature of deposition of the chalcopyrite, although it probably was quite comparable to that of sphalerite.

Also, it is apparent from Wheatley's plates that the metamorphism to which the Avoca ore minerals were subjected was minimal. Wheatley seems to consider that one result of metamorphism, was the development of porphyroblastic undeformed sphalerite in sericite schist, this sphalerite in places being associated with galena in 'pressure fringes'. To him this illustrates the ductility contrast between the two sulfides. In ores where the primary textures of pyrite are almost certainly unchanged from their original shapes, it seems unreasonable that the crystals of sphalerite in sericite should be considered as porphyroblasts when they could as easily be classed as replacement structures around which minor amounts of galena had been deposited subsequent to the emplacementof the sphalerite crystals.

Recent work by National Oceanic and Atmospheric Administration (NOAA) geologists, working under Alexander Malahoff, have discovered in the East Pacific Rise area huge deposits of copper sulfides containing appreciable silver (and probably lead and zinc). Malahoff reports that these sulfides are on the sea floor and surround old volcanic vents, but these deposits lie at depths of over 2500 m, so the conditions of their deposition cannot be compared with those of the shallow-water conditions obtaining in the lower Carboniferous Irish seas. Further, just what will happen to these East Pacific Rise deposits with the passage of an appreciable span of geologic time? Present opinion, in France at least as expressed verbally at the 26th IGC in Paris in 1980, holds that, if sulfides reach the sea floor (as of course they could at Pacific Rise depths) they would be so thoroughly oxidized in a geologically short period of time that they would never be incorporated as such in a permanent stratigraphic record. Also, it would seem to follow that sulfides pouring out of volcanic vents as "sulfide smoke" hardly could have acquired their sulfur from the sea water they met on leaving their vents. This is particularly true since these sulfides appear to have been such before they left the vents and must have had their sulfur with them when they were formed at some measureable depth beneath the sea floor. The current explanation of such a situation would be that sea water, moving beneath the sea floor would have had their sea-water sulfate reduced to S^{-2}, and these sulfide ions would have reacted with the metal-chloride complexes to produce the sulfides in the "sulfide smoke". No one can deny the presence of the chlorides in the ore fluids (either beneath the sea floor or beneath land masses as well), but I still contend (as I did in 1973) that the chlorides are just along for the ride and have no more to do with the ores being formed than the passengers in an airplane have with keeping that aircraft in flight. The weakest point in the theory that produces sulfide ore bodies from volcanic exhalations is the lack of a theoretically sound mechanism for producing enough sulfide ion by the reduction of sea-water sulfate to provide all the sulfur needed for the sulfides emplaced in the ore bodies in the short period of time available. Further, the problem (even if bacteria in sea water can work in a most efficient manner at elevated temperatures) still is real as to how to produce all the sulfide ion needed as fast as it is required when such ore fluids are invading solid rock where no sea-dwelling bacteria can be expected be be present. In short, the concept of the metals in sulfide ore deposits having been transported as metal-sulfide (or bi-sulfide) complexes and having been deposited as sulfides and sulfosalts is most reasonably explained by these complexes reacting among themselves and with available uncomplexed metal ions to produce the typical sulfide-mineral suites of the various types of sulfide ore deposits throughout the world. This applies whether the ore deposits were formed in rocks beneath the sea floor or within rocks not covered by sea water at the time of ore emplacement.

Wheatley's paper on the ore mineral fabrics at Avoca raises a number of points. The first of these is that the primary colloform, melnikovitic, and zoned crystalline pyrite indicate that the pyrite probably formed in pore waters in the sediments. This suggests some degree of lithification in these sediments; if this assumption is correct, then the pyrite was emplaced at an appreciable time after the rocks containing it, long enough for the stability of the pores to have been established. As a second conclusion, Wheatly thinks that the minor carbonaceous horizons in the Avoca sediments at times produced reducing environments at the sediment-water interfaces, and in such positions, aqueous ionic sulfur [S^{-2}?] could have been produced through bacterial reduction of sea-water sulfate. This sulfide ion reacted with ferrous ions [from iron-chloride complexes (?)] to produce framboidal pyrite. [Why such conditions would have been particularly effective in producing framboids is not explained.] If pyrite was so produced, the dominant available sulfide ion must have been S_2^{-2} and not S^{-2}, which means that the reduction did not proceed as far as S^{-2}. Wheatley admits that the siliceous stockworks stratigraphically below the pyrite-rich zones do not contain appreciable colloform pyrite, that they are demonstrably epigenetic, and that they probably formed from chloride-rich, metal-laden hydrothermal brines. He does not say where the sulfur in this pyrite came from. As a correlary point, he says that porosity increases in the basal area of these pyritic zones and that this may have resulted from leaching by circulating pore waters following pyrite formation. Wheatley claims that these pore waters would have been appreciably acidic in a sulfide-rich environment [presumably this sulfide ion concentration was developed from sea-water sulfates although why waters rich in sulfide ion should have been more acidic than ones rich in sulfate ion is not clear]. He goes on to say, however, that these pore waters, on mixing with hydrothermal brines, would have been highly corrosive of primary pyrite. [This seems to suggest that Wheatley believes that the primary pyrite was definitely older than the copper, lead, and zinc sulfides that presumably were deposited from these hydrothermal brines. Yet the pyrite can only have come from iron brought into the system by an earlier stage (?) of these same hydrothermal brines.] He is of the opinion that this corrosion of pyrite caused a rapid increase of reduced sulfur species [I suppose he means that the S_2^{-2} of the pyrite was reduced to the S^{-2} needed to form the lead, zinc, and copper sulfides. [Though how the rather complex Cu-Fe-S chalcopyrite was produced by this process does not seem clear.] He makes the statement that "effects of fluid mixing [by which I suppose he means the mixing of hydrothermal brines with pore water], coupled with decrease in temperature and pressure [why, with greater depth, should the hydrothermal brines decrease in temperature?] gave stockwork zones at lower levels." He considers that "corroded, unzoned crystalline pyrite indicates ore deposition in the presence of a corrosive medium; possibly chloride-rich." Wheatley discusses the ore mineral fabrics under three headings: (1) pyrite zones; (2) siliceous zones; and (3) lead-zinc zones.

In the discussion of the fabrics in pyrite zones, he states that a complete gradation exists between massive and disseminated pyrite [as has already been said] and that massive seams consist of euhedral to subhedral crystals aggregates (0.1 to 1.5 mm grain size) these crystal aggregates are seen normally only after etching. Zoning is developed almost everywhere; subgrain zone boundaries have lesser hardness and etch preferentially. Colloform textures are common, especially in the upper levels where they occur in conjunction with botryoidal and reniform types. Wheatley accepts the pyrite morphology as a primary feature so, where truncations of crystal outlines or core or zone boundaries are noted, he considers that they resulted from replacement. Brecciation of pyrite took place only where the pyrite is adjacent to slickensides. Melnikovite [which Fleischer says is greigite - $Fe^{2}Fe^{3}_{2}S_{4}$ -] reacts strongly to weathering and can be distinguished by its brown color. The chalcopyrite in pyrite zones is present as an interstitial cement for pyrite grains and aggregates and fills pore spaces. Wheatley says that chalcopyrite "is more ductile than pyrite" [although what this has to do with its emplacement, I do no see since it does not seem to have been moved by plastic flow]. Chalcopyrite is concentrated in siliceous zones relative to pyrite ones. Sphalerite is interstitial to pyrite grains, filling open spaces and replacing zoned pyrite crystals. Wheatley thinks that the larger sphalerite crystals in the sericite schist are porphyroblasts [as has already been stated] and galena fringes these crystals, demonstrating [to Wheatley] that the galena is more ductile [this has been commented on above]. Wheatley holds that metamorphism and deformation [apparently at Avoca] cause galena to migrate into volumes of low pressure or zones of dilation and to replace other major primary minerals. [Of course, such relationships can be interpreted as the results of later deposition of galena than of pyrite and sphalerite; the lack of other major evidences of impressive deformation suggests that metamorphism is not responsible for the development of galena as just described.]

In the siliceous zones pyrite is disseminated in a siliceous-chlorite matrix with its grain size depending on the degree of aggregation. Pyrite crystals may be embayed by gangue minerals or other sulfides, almost certainly as the result of replacement. Mechanical deformation appears to be confined to slickensides. Wheatley says that more ductile chalcopyrite takes up the bulk of the strains so produced by glide twinning and recrystallizing. Because of the minor amounts of chalcopyrite in the ore in relation to pyrite, the amount of strain to be so taken up does not suggest major metamorphism. Chalcopyrite, sphalerite, and galena show fabrics similar to those of pyrite but are enclosed in a siliceous matrix.

In the lead-zinc zones, pyrite provides a fine-grained, banded matrix to sphalerite and galena; replacement of pyrite by sphalerite is common. Where pyrite is weathered, colloform textures are important, but radiating, reniform, and botryoidal and worm-like and atoll structures are known in the pyrite. Wheatley explains fringes of sphalerite or galena around pyrite as being due to their greater ductility. [Such textures also may have been produced by later deposition.] The textures of chalcopyrite, sphalerite, and galena correlate with those of pyrite.

Wheatley argues that the textures of pyrite indicate growth at less than 100°C in an environment in which pH was low and the solution supersaturated with iron. The material in which the pyrite is contained shows lateral continuity and the fabrics show no evidence of slumping. The pyrite does not appear to have been moved by bottom currents, and the pyrite grains show no abrasive effects. Wheatley also says that no suitable site for pyrite deposition existed above the sediment-water interface. He thinks that the zoned growth of the pyrite crystals argues for slow growth and changes in the composition of the fluid medium; this he considers is compatible with sediment pore waters. [Such development also probably is compatible with pyrite introduced by replacement of the already present gangue minerals.] There is no question but that the ores were deposited in a volcanic environment, the only question being were the ores emplaced during the deposition of the enclosing sediments or after they had been essentially lithified? Wheatley insists that the non-ferrous sulfides were deposited by reaction between metal-chloride-rich brines and zoned and colloform pyrite. This may be true, but this means that all the sulfur must have come from pyrite and that the sulfur-isotope characteristics of the pyrite sulfur would have determined by the isotope ratios of the pyrite and would tell nothing about where the metals came from.

As is the case with the Silvermines deposits, I am convinced that the ore minerals were emplaced after the lithification of the enclosing sediments, that these were appreciably altered, mainly to sericite, before and during the period of ore emplacement, and that the temperature of emplacement of the non-ferrous-metal sulfides was above 300°C as is shown by the exsolution textures of chalcopyrite in sphalerite. The emplacement temperature of the pyrite may have been considerably lower, although this has not been convincingly demonstrated.

I think, therefore, that the ores were introduced largely by replacement under mesothermal conditions in middle Paleozoic time, that is, in connection with the volcanism associated with the Caledonian orogeny. The ore fluids (of whatever composition they may have been) undoubtedly came from the same source as the volcanic rocks with which the ores at Avoca are interbedded.

SILVERMINES, COUNTY TIPPERARY

Late Paleozoic	Lead, Zinc, Silver	Mesothermal to Leptothermal

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Williams, C. E. and McArdle, P., 1978, Silvermines, in Bowie, S. H. U., and others, Mineral Deposits of Europe, Volume I, Northwest Europe: Inst. Min. and Met. and Mineral Soc., London, p. 333-336

The known ore occurrences in the Silvermines district lie in an east-west-trending belt that runs from Silvermines village (52°48'N, 8°4'E) on the east as far as the abandonned Shallee mine on the west, a distance of about 3.5 km. Silvermines village is in the northern part of County Tipperary and is 8 km south of the town of Nenagh and 30 km northeast of the city of Limerick. The mines are located on the northern slopes of the Silvermines Mountains, and the crest of these mountains is some 375 m above the valley of the Kilmastulla River, a small tributary flowing more or less southwest to the Shannon.

In 1645, a report was circulated that described the discovery of a lead mine in the Silvermines area some 40 years previously. The "ora" was said to contain three pounds of silver to each "tun", but tun of what was not stated. Some mercury also was said to have been recovered, but no other by-product was mentioned. Early in the 19th century, a geologist named Weaver described four vertical north-south veins in the Old Red Sandstone at the Shallee mine (at the far west end of the present district). Weaver was convinced that these veins were formed by segregation that took place contemporaneously with the sedimentation.

The first real peak in mining activity occurred during the 1860s with lead and silver from the galena, zinc from hemimorphite [$Zn_4Si_2O_7(OH)_2 \cdot H_2O$], and copper from (possibly) chalcopyrite. By 1870, competition from foreign lead and zinc forced the mines to close. Not until 1917 did exploration recommence and then only on a minor scale. Some mining of galena and sphalerite was undertaken between 1949 and 1952; exploitation of oxidized zinc ore was unsuccessful. Still further exploration between 1952 and 1957 (results reported by Rhoden, 1958, 1959) produced a considerable fund of information about the various deposits in the district. The next exploration near Silvermines was that in the early 1960s of the long-known Ballynoe barite deposit that is centrally located in the Silvermines district. This exploration located several million tons of high-grade barite ore, the discovery of which coincided with the greatly expanded use of that mineral for oil-well drilling muds; production from Ballynoe now (1978) runs about 200,000 tons a year.

The discovery of the Tynagh ore body in 1961 increased the interest in exploration for base metals in central Ireland, and the Silvermines area was one of the first targets studied because of the history of prior mining in the district. A joint venture for the exploration of an 80 km^2 rectangular block along the Silvermines fault was organized by International Mogul Mines and Silvermines Ltd. and began there in late 1962. A great deal of geophysical and geochemical work was done; the latter method appears to have been far the more effective of the exploration tools. Diamond drilling was commenced in mid-1963, and, after some false starts, success was achieved; by the end of 1964, 280 holes for a total length of 43,000 m had outlined 11,000,000 tons of ore that ran above 10 per cent zinc and lead combined. By 1968, production of lead and zinc concentrates had begun and has continued at least until 1980; 7,200,000 tons of ore with a grade of 8.9 per cent zinc and 2.8 per cent lead had been recovered by the end of 1976. Concentrates recovered contained 114,000 tons of lead metal and 494,000 tons of zinc metal. The lead contained "minor" silver and zinc "minor" cadmium.

The deposits of the Silvermines district occur on the north slopes of the long ridge comprising the Silvermines Mountain and is on the south side of the Kilmastulla Valley. The ore is at an elevation of less than 100 m above the valley under a gently sloping surface of glacial drift that in places is more than 120 m thick. The valley is underlain by an open syncline of lower Carboniferous limestones, the axis of which plunges gently to the northeast. In the low hills on either side of the valley are inliers of older rocks, both Silurian siltstones and graywackes and Devonian sandstones.

Taylor and Andrew (1978) report that the following deposits were available for production in 1968:

Ore Zone	Stratigraphic Position	Indicated Reserve Short, Dry Tons	Pb%	Zn%	Ag, oz/ton
Upper G	Waulsortian	9,332,100	2.40	9.20	0.74
K	Old Red Sandstone	264,000	2.40	7.35	----
B	Waulsortian	4,591,600	2.50	6.50	1.01
Lower G	Lower Tournaisian	2,068,800	4.50	3.40	1.13
K	Lower Tournaisian	2,083,500	1.38	4.58	0.50
C	Upper Old Red Sandstone	396,400	0.97	5.28	----
Shallee Mine	Upper Old Red Sandstone	1,468,200	1.95	0.10	0.60
Magcobar Deposit	Waulsortian	±2,500,000 @ 85% $BaSO_4$			

The total of lead-zinc ore in this table is 22,204,600 of which (Taylor and Andrew, 1978) only 11,000,000 ran more than 10 per cent lead and zinc combined. Any ore estimate carried to the nearest 100 tons on the basis of diamond drilling gives the impression of being far more accurate than any estimate of this type can be. How the total of 11 million tons above 10 per cent lead and zinc combined (Williams and McArdle, 1978) was achieved is not stated, but the total of 7.2 million tons in 8.5 years at an average grade of 11.7 per cent lead and zinc combined indicates that this estimate was quite good. This tonnage in 8.5 years is at an average rate of about 850,000, not far from the target of 1 million quoted by Taylor and Andrew, but this means, if the original estimate of 11 million tons has not been added to, that, at the end of 1976, only 2.5 million tons (or three years production) remained to be taken from the ground. Obviously more ore has been encountered or the mine certainly now would be shut down, which it apparently is not.

As I remember it, I was told when I visited the Silvermines district in 1967 that drilling beneath the earlier-known Magcobar ore had encountered sulfide mineralization. I have heard nothing more about this, and, if it were true, surely Taylor and Andrew would have mentioned it. Larter and his colleagues (1981), however, [based on their recent work on Magcobar (Ballynoe)] suggest that an epigenetic sulfide mineralization may remain to be discovered under Ballynoe. Perhaps their thinking was influenced by the discovery, under the early known massive barite at Walton, Nova Scotia, of a polymetallic sulfide deposit that had an almost certain epigenetic manner of emplacement. The change from massive barite to sulfides at Walton was a gradual one, and the deposit as a whole is classified as mesothermal to leptothermal (as are the Silvermines deposits). Thus far, at least, no sulfides have been discovered under the Magnet Cove massive barite deposit, but its manner of genesis probably was appreciably different from that of Ballynoe.

In the area around Silvermines, surface exposures and underground drilling have made known rocks of Silurian to Carboniferous (Visean) age (Table I). The Silurian beds make up the core of the Silvermines Mountain and constitute a sequence of gray, green, and purple mudstones with subordinate grits and sandstones that have been isoclinally folded. Although no fossil evidence has been found, they are lithologically much the same as rocks of Upper Llandoverian-Ludlovian age that outcrop in the general area.

TABLE I. STRATIGRAPHY AT SILVERMINES (after Taylor and Andrew, 1978)

AGE	LOCAL DESIGNATION	THICKNESS	LITHOLOGY	MINERALOGY
Visean (Chadian)	Calp-like dark ls	Top not seen	Ptly silicified massive nodular dark argillaceous ls, shale partings, chert nodules twd base	
	Supra Reef		Mixed sequence of fine-dolomite, nodular ls, subordinate sh	
	Chert Breccias	30-60 m	Black massive or banded nodular chert, locally cem'd w gray dol, brecciated at base	
	Upper Reef Ls	0-30 m	Pale "Waulsortian" reef ls and ls breccias w grayish massive chert locally developed at base	
	Pale Reef Ls and Dol Breccias	30-126 m	Interfingering sequence of stromatacid reef ls, reef ls br and dol br w massive py, barite br & siderite br twd base	Stratiform ore or Upper G and B zones Magcobar barite
Tournaisian (Courceyan)	Mudbank Reef Ls	15-45 m	Poorly bedded light nodular micrite w dark gray sh and light nodular biomicrite - of crinoidal ls, dol sh, green sh, silic ls and chert	Up to 5 m of econ Pb-Zn minzn in silicified strata under ore in B and Upper G ore zones
	Muddy Biolcastic Ls	15-60 m	Thin interbeds of dark calcareous, fossiliferous sh and light nodular biomicrite, ls increases downward	
	Lwr Dolomite	30-90 m	Gray massive rextlzd fine dol biomicrite, part nodular, wavy sh laminae	Stratabound Lwr G and K zones

TABLE I(cont.)

AGE	LOCAL DESIGNATION	THICKNESS	LITHOLOGY	MINERALOGY
Tournaisian (Courceyan)	Basal Fragmental Beds	0.40 m	80% irreg nodules and lenses of light dol biomicrite, 20% wavy dark sh laminae	
	Ballyvergin Sh	1-3 m	Light mudstone w silt lenses	
	Basal Sh	21 m	Thin-bedded black dol sh w lenses and laminae of biomicrite, more silty twd base	
Devonian (Upper Old Red Ss)	Transitional Ss	6 m	Thin interbeds of fine gray qtzites, dark gry & grn impure ss, sandy sh, & siltstones	
	Upr Shallee Ss	24 m	Gry to white fine-med grnd qtzites w minor qtzose grits, gry & grn sh & grn ss	py-barite vein minzn at Shallee & stratiform horizons of C zone, Breccia ore of K zone
	Middle Grn Ss	21 m	Interbeds of gry & grn fine qtzites & grn sh	
	Lwr Gry Ss	27 m	Med-grnd gry X-bedded qtzites w scattered mudstone & qtz frags & local qtz-mudstone conglomerates	
	Lwr Grn Ss	6-26 m	Fine-grnd grn ss, sh, & siltstones	
	Basal Gry Ss	12-24 m	Gry & white med-grnd qtzites w scattered qtz & grn mudstone cherts w interbeds of qtz mudstone grits & conglomerates	
angular unconformity				
Silurian (Upr Llandoverian-Ludlovian)		Base not yet known in Silvermines area	Gry, purple or grn shales, mudstones & siltstones, pt altered to slate w intercalated graded grn & gry ss	Galena, sphalerite & minor cp in qtz-cal vns

The Devonian beds in the Silvermines district lie on Silurian rocks above an angular unconformity; they are facies of the Old Red Sandstone and are some 120 m thick. They are principally quartzites of various colors, mostly quite clean; they are classified, on the basis of color, into six units. The strata show features diagnostic of their deposition under deltaic and coastal-litoral environments. Where followed up, they change from gritty quartzites with lenses and interbeds of conglomerates passing through finer, but impure, quartzites with mudstones and shales into mainly clean quartzites. The last 6 m of the Old Red is composed of thin-bedded fine, clean quartzites, sandy shales, and mudstones that Taylor and Andrew consider to be transitional to the overlying and conformable Carboniferous carbonate beds.

In the Carboniferous, the lower 250 m is a shallow-water sequence of thin-bedded, generally nodular bioclastic limestones and shales; their abundant fauna is Tournaisian and equals the 'Lower Limestone Shales' of the Irish Survey. In the mine area, this sequence is broken down further on the basis of the proportions of limestone and shale in the various beds (Table I). The mineralization of the Lower G and K zones is in the lower Dolomite unit which Taylor and Andrew think is the completely dolomitized massive lower portion of the Muddy Bioclastic limestone. Near the base of the Carboniferous series is the Ballyvergin limestone that is the only district-wide marker bed; it has been called "water-lain tuff", but these authors see no evidence of this near Silvermines. The uppermost Tournaisian strata are those of the Mudbank Reef limestone; these are much varied but are mainly poorly developed "off-mudbank" lagoonal facies that have compositional connection with the Waulsortian (Visean) reef limestone. Green shale beds that may be present between the Mudbank Reef limestone and the Visean ore zones do not appear to Taylor and Andrew to have been derived from volcanic debris. These beds probably were mud slowly brought in from a terrestrial source.

Above the Mudbank Reef rocks (deposited in a stable quite-water environment) were strata developed under turbulent conditions caused by dynamic and tectonic disturbances evidenced by overlying breccias. As far as the fringes of the B and Upper G ore zones, the enclosing rocks are 'Waulsortian' stromatacid carbonate mudbank reefs. This sequence is largely a series of dolomitized clastic breccias, pseudo-breccias, and what Taylor and Andrew refer to as "tectonically triggered in-situ shake breccias"; these are the Silvermines equivalent of the 'Waulsortian Reef' facies. This sequence from the base of the Visean upward through the Pale Reef beds may be as much as 250 m thick with the greatest thickness (much thicker than the 30-126 m given in Taylor and Andrew's Table II, here Table I) being on the downthrow sides of northwest-trending faults that Taylor and Andrew believe were active during sedimentation. Toward the south, the 'Waulsortian' beds thin down to about 100 m, and there the whole series is completely brecciated and dolomitized. To the northwest (in the thicker parts of the sequence) the proportion of unbrecciated and undolomitized rocks grows larger. In the lowermost portion of these beds, the main rocks are massive pyrite, barite breccia, and siderite breccia; these contain the stratiform mineralization of the Upper G and B ore zones.

Above this series of breccia-dominated rocks is a series of interfinering reef limestones and reef limestone breccias designated as the Upper Reef limestone. Followed upward and laterally, this sequence gradually becomes higher in chert, forming massive cherts and chert breccias; this is the 30 to 60 m of Chert breccias (Table I). Above the cherts are an irregular series of limestones called the Supra Reef and are thought to be the transition between the Waulsortian beds and the overlying Calp-like limestones. The Calp beds are the youngest rocks still present in the mine area; they are massive blue-gray argillaceous limestones as is the usual Calp.

So far as is at present known, all the ore bodies in the Silvermines area are located on the south limb of the broad and asymmetrical Kilmastulla syncline, the axis of which strikes ENE and which plunges gently in that direction. The dips on the north (unmineralized limb) average about 10°S, whereas those on the south limb (where the ore bodies are) range between 10° and 30°N except in the drag zone next to the Silvermines ENE-trending fault complex where dip may be as high as 60°. Taylor and Andrew consider the ENE fold-trend as being of Amorican age, although the typical folds of this character are more usually east-west in the County Cork area. In the Silvermines fault zone, the total movement amounts to 200 to 350 m of down throw on the north side of the fault zone. It is the major structural feature of the district. Although Rhoden (1958) placed the faults in the Silvermines district into seven groups, Taylor and Andrew think that the description of the three dominant sets is enough for a study of the structure of the district. These are: (1) the ENE-trending faults, the main one of which is the Silvermines fault proper on which the north side is downthrown and the dip is between 70° and 90°N: this fault runs for 40 km through northern Tipperary and Limerick; they consider that the fault originated in the pre-middle Old Red Sandstone deformation; (2) the NW-trending faults that cut and displace the Silvermines fault; this gives the trace of the Silvermines fault the sawtooth shape that is particularly apparent in the G zone and the eastern K zone; these are normal faults with displacements of 50 to 65 m; the north side is downthrown and the dip is between 70° and 90°N; and (3) NNW-and North-trending faults; these normally are cut by trends (1) and (2) and are prominent as block faults in the vicinity of the Shallee mine. Taylor and Andrew believe that the manner in which these three sets of faults are associated together fits the concept of a major dextral strike-slip system that later was modified by a dip-slip movement.

If all the known ore bodies in the Silvermines district were exposed at the surface, they would lie in an irregularly east-west-trending belt, with the secondary Calamine mine (or P zone) on the east end, directly south of Silvermines village; next would be the K ore body along the trace of the Silvermines fault, then the Magcobar barite deposit, then the two G ore bodies (with the Lower G directly against the fault and the Upper G slightly north of it), then the S zone (also along the Silvermines fault), and finally the Shallee mine (or C zone) south of the Silvermines fault in Old Red sandstone.

Taylor and Andrew report three distinct styles of mineralization in the Silvermines district: (1) the stratiform ore bodies that were emplaced in the Waulsortian carbonates; these are the Upper G zone, the B zone, and the Magcobar barite open-pit - these authors regard these as having been formed both syngenetically and syndiagenetically; (2) the stratabound ore bodies that are located in the Tournaisian Lower dolomite (about one-third of the way above the base of the Tournaisian); these are the Lower G zone and the K zone (which had not been fully delineated when Taylor and Andrew wrote in 1978); and (3) the tectonically controlled epigenetic vein and breccia ore bodies, the best known examples of which were in the Shallee mine (abandonned in 1958) and in the K and C zones which were known in 1978 only by drillhole data.

The stratiform ore body of the Upper G zone is contained in massive, partly brecciated pyrite (but with 20 per cent of the ore being disseminated breccia ore and 5 per cent fine-grained sulfide muds). These ores are at the base of the dolomite-breccia series that lies essentially at the base of the Visean sequence immediately above the Tournaisian Mudbank Reef limestone. This ore has a maximum thickness of 30 m and this is reached in the south end of the body as controlled by the northwest-trending faults (Taylor and Andrew, 1978). Older authors consider that the Upper G zone is bounded on the south by the Silvermines fault. To the south, however, the boundaries are sharp, being provided by those of the massive sulfides; to the north, the ore follows the irregular contact with the footwall Reef limestone against which the pyrite ore pinches out. The ore body changes abruptly from massive pyrite to footwall sediments that contain practically no sulfides. The sulfides of this ore are made up of 20 per cent sphalerite and 4 per cent galena with traces of chalcopyrite, tennantite, boulangerite, bournonite, jordanite, and pyrrhotite, but the bulk (75%) of the ore is pyrite and marcasite. The gangue minerals are dolomite, calcite, chert, clay minerals, barite, and quartz. The sulfides and gangue are located in bands of various compositional proportions. Taylor and Andrew report a wide variety of colloform textures such as spheroids and framboids, plus concentric banding, and dendritic structures. The breccia portion of the ore normally is unstratified and poorly sorted and consists of fragments of pyrite and sphalerite with considerable differences in size in a matrix of fine-grained sulfides, and less dolomite and chert. The overlying Dolomite breccia is an unsorted mass of rounded to subangular dolomitized limestone clasts in dolomite-mud matrix that shows some sedimentary features. Minor pyrite is present in the lower 15 m of the breccia as detrital and replaced fragments. The B zone ore body also is a stratiform ore body and is located in the same stratigraphic horizon as the Upper G zone; the dolomite breccia is less abundant than it is in the Upper G with the main host rock being reef limestone. Williams and McArdle think that the structural situation of these two ore bodies is essentially similar with each being bounded along the footwall by a normal fault (NW-trending according to Taylor and Andrew) and with the dip of the containing beds being appreciably steeper near the fault than farther away from it. The thickness of the zone ranges from nearly 0 to 30 m and averages about 5 m. The footwall contact of the ore is defined by the stratigraphic break between the Mudbank Reef limestone and the overlying breccias; locally, however, silicified portions of the footwall may include laminae of fine-grained, economically viable sphalerite that are as much as 5 m thick. In the massive pyrite portion of the B z one, the actual contact is not easily determined as pyrite is present in appreciable amounts in the footwall Reef limestone. The host rocks of the ore zone are divided into three types: (1) barite breccia, (2) pyrite breccia, and (3) siderite breccia; these breccias also include lenses of Reef limestone breccia, and Dolomite breccia that interfinger with the three breccia types just described.

The major sulfides are pyrite, sphalerite, and galena with minor marcasite; the galena contains spindle-shaped intergrowths of silver-bearing boulangerite. Normally, these ore minerals are fine-grained, although coarse-grained patches do occur. At the northern extremities of the B zone, the contact of the ore-hanging wall is not easy to distinguish because of the fine grain of the ore and the lack of marked lithological differences in the enclosing rocks. The barite breccia is mainly fine-grained breccia fragments but contains inclusions of hematite and jasperoid near its base; its mineralization is largely fine intraclastic disseminations of galena and pyrite with minor fine-grained sphalerite. The pyrite in that zone is either massive or brecciated; the massive variety is finely crystalline or colloform (in a number of textures). Sphalerite and galena are present as fine disseminations or coarser replacements. The breccia matrix is argillaceous or dolomitic and is partly replaced by fine sphalerite and galena. Williams and McArdle say that the siderite breccia is the main ore type, the gray siderite matrix giving the rock a dirty gray color; the siderite clasts in the breccia are of varied size, angularity, and packing that locally are stratified with interfingering beds of unbroken shale. Sphalerite occurs in the clasts, commonly in the matrix, and locally filling irregular cavities in more massive volumes. A series of NNE-trending basins are present in the siderite zone in which basins the siderite is thicker than the average and contains a strong, but fine-grained, sphalerite mineralization. Taylor and Andrew consider that these basins are an expression of the initial paleotopography and provide impressive evidence of syngenesis of the ores. The presence of stromatacid Reef limestone on the top of the Mudbank Reef limestone provided a barrier to mineralization so that it pinched out against the Stromatactid Reef limestone along its northern and eastern boundaries.

The Magcobar (Magnet Cove Barite) ore body (the Ballynoe mine) is in the same stratigraphic level as the stratiform Upper G and B ore zones, but it occupies a higher structural level, although being located geographically between them. The Magcobar ore lies immediately to the south of the area in which the Upper G and B ore bodies connect. The barite ore is directly above the layer of impersistent green shale that locally separated the Visean beds (above) and the Tournaisian beds (below). This ore zone has an irregular shape, being about 650 m along strike and about 250 m down dip; it ranges between 0 and 15 m thick and averages about 85 per cent $BaSO_4$. Above the green footwall shale, there is a basal sequence of layered hematite, hematitic barite-jasperoid bands that change to brecciated pinkish to white barite. This barite is covered by a cap of sulfides composed of as much as 10 m of pyritic dolomite breccias with framboidal pyrite and melnikovite plus some fine-grained galena and pale sphalerite. At its margins, the barite zone commonly interfingers with siderite breccias; above the barite may be either massive siderite breccias or a thin chert band overlain by pyritic dolomite breccias. The basal sequence in the barite zone (Taylor and Andrew, 1978) shows slumping and brecciation of the type expected of soft sediments, and what they call syneresis cracking is present in the jasperoid bands. They believe that these "sedimentary" features in the main barite ore body indicate that the barite was deposited as a microspherulitic-cryptocrystalline 'mud' that was remobilized, recrystallized, and stylolitized during diagenesis. They believe that barite texture, thickness, and lateral extent can be correlated with the paleotopography of the Mudbank Reef limestone of the footwall. They also consider that the barite is the lateral equivalent of the massive siderite and pyrite rocks that hosts to the zinc-lead mineralization of the Upper G and B zones. They find it reasonable that these facies variations were developed in a restricted lagoonal environment in which appreciable differences in Eh and pH existed from one part of the lagoon to another.

The Lower G zone is enclosed in the Tournaisian Lower dolomite and is bounded by two northwest-trending cross faults. The best mineralization, in grade and thickness, appears to be between the faults that are beneath the Upper G ore body and the Silvermines fault. Along this line, the two G ore zones are directly adjacent to each other with the Dolomite breccias and the Lower dolomite seeming to have pinched out the Muddy Bioclastic limestone near the fault. North of the fault, the Upper and Lower G ore zones diverge. the zones being a considerable vertical thickness of barren rock apart only a short distance from the Silvermines fault; the Lower G ore zone is confined to the basal portion of the Lower dolomite.

Unlike the Upper G zone, the mineralization in the Lower G is far less abundant with the content of sulfides in the Lower G being less than 20 per cent total; galena is more abundant in the Lower G, sphalerite less common, and pyrite no more than a minor fraction of what it was in the Upper G. Other sulfides present in minor to trace amounts are chalcopyrite, marcasite, tetrahedrite (and other similar silver-bearing sulfosalts), arsenopyrite, and löllingite. The main gangue minerals are dolomite, quartz, and barite; the sphalerite contains what appears to be exsolution blebs of chalcopyrite.

Where the Lower and Upper G bodies are in direct contact (at the south end of the two ore bodies, against the Silvermines fault), the Lower G mineralization is normally massive and fine-grained and contains abundant colloform textures. Farther down dip, it is made up of coarse-grained open-space fillings of veins and other cavities and local brecciated zones; also considerable disseminated sulfides are present. To Taylor and Andrew, these ore arrangements indicate widespread remobilization. They emphasize the considerable variations in zoning (or lack of it) in the anastomosing veins, the different types of veins margins, the mono-or polymineralic character the vein fillings, and the varied vein thicknesses, probably as providing confirmation of the remobilized nature of the vein fillings.

The K zone lies appreciably farther east than any of the other ore zones, except the long abandonned Calamine zone. The southern limits of the K abut against the Silvermines fault (sensu stricto as Taylor and Andrew say) with its stratigraphical setting being similar to that of the lower G zone. It reaches its maximum thickness in the steeply dipping Lower dolomite, splitting into two separate ore horizons away from the fault. The upper horizon is generally some 3 m thick but is as much as 10 m in places; it extends for about 75 m north of the fault. The lower horizon, also in gently dipping beds away from the fault, has a south-north length of 275 m; its thickness is normally between 4 and 7 m. Where northwest faults cut the lower and eastern part of the K zone in the basal section of the Lower dolomite, the ore zone increases to 12 m in thickness. The K zone is completely concordant with the contacts of the Lower dolomite, being, therefore, truly stratabound. The Lower dolomite overlies the Basal Fragmental beds, and this contact normal sets the lower limit on the ore although the Fragmental beds do contain traces of sulfide mineralization. In the lower 60 m of the Lower dolomite, the boundaries of the ore are determined by assaying - lithologic boundaries to mark the upper limit of K-zone ore are lacking. The sulfide mineralization is made up mainly of sphalerite with minor galena and pyrite and traces of chalcopyrite and pyrrhotite.

Taylor and Andrew, who emphasize the importance of the northwest-trending faults in controlling the boundaries of the ore, point out that, in the western part of the zone, where no such faults are present, the sphalerite usually is a fine-grained dark type that occupies the margins of cavity fillings and fills anastomosing veins that Taylor and Andrew consider to have been produced during diagenesis. Some light sphalerite and galena may occur in the central parts of some veinlets and cavities, surrounded by dolomite or barite. In the fault-dominated part of the K zone, the sulfides are more coarse-grained and show textures that Taylor and Andrew says are "indicative of remobilization". The sulfides are present in spaces among clasts, in thick tensional veins, the fillings of which are symmetrically banded, and in minor en-echelon zones of gash veins; all of these structures are, as Taylor and Andrew point out, related to movements on the faults. The margins of the veins are composed of fine-grained dark sphalerite then comes galena and pyrite, and then a central core of dolomite or barite that surrounds coarse crystalline sphalerite, in orange or brown colors.

The P zone (the Calamine zone or Ballygowan South mine) was mined for that zinc mineral until 1958; it is considered by Taylor and Andrew to be the down-dip extension to the north of the K zone and is located away from the main northwest-trending fault. The ore is in the Lower dolomite, which, despite its distance from the northwest fault, is oxidized down to 85 m below the surface. The ore minerals in this material are hemimorphite (identified by my great-great uncle in 1859), smithsonite, and cerussite; unoxidized galena and sphalerite are encountered in increasing amounts with depth. Above the primary protore, the grade of zinc indicates that the metal has been redistributed, dispersed, and, overall, enriched.

The Shallee mine is the farthest west of the known ore bodies; these ores occur in gently dipping upper Devonian beds in a part of the district that is bounded by the WNW-trending Main fault and the northwest-trending Western Reefs fault which faults have brought the Carboniferous beds on the north side of that fault down to the level of Devonian Upper Shallee sandstone; some mineralization is present in the underlying Middle Green sandstone. The major sulfide is galena with barite and pyrite in moderate amounts and chalcopyrite, sphalerite, arsenopyrite, and tetrahedrite (and other silver-bearing sulfosalts in trace quantities). Although disseminated galena is quite common in the sandstones, the principal manner of mineral emplacement is as open-space filling of north-south-trending ramifying veinlets that are as much as 15 cm in width. These veinlets contain crustiform galena with lesser barite and pyrite. In the veinlets near the main northwest-trending faults, the mineralization is coarser and has a network-like form; this is similar to the Devonian mineralization of the K zone. The block-faulting of the beds in the Shallee area has been accomplished by N-S and NNW-trending faults; these post-date the mineralization but affect only minor displacement of the ore.

When Taylor and Andrew studied the deposits at Silvermines, surface diamond drilling cut fractured, brecciated, and mineralized Devonian beds located between two northwest-trending faults that both intersect and displace the ENE-trending Silvermines fault. These two faults dip about 80°NE; the throw of the more northerly fault is down for 15 m but that of the southerly fault could not be determined. The extent of this mineralization had not been located in 1978; the maximum thickness attained is 35 m at a depth of 100 m but higher up drops off to 5 m - no lower limit had been set.

This mineralization in Devonian beds is located in quartzites, grits, and shales that are shattered or even more strongly brecciated. The cracks in these rocks are filled with veinlets of fine-grained variously colored sphalerite and pyrite and minor galena. Included in this sequence is a series of thick brecciated horizons composed of poorly sorted and closely packed angular clasts of clean quartzites and angular shales that contain symmetrically banded mineralization of variously colored sphalerite and pyrite and locally coarse galena. Most calsts do not appear to have been transported because the heavy mineral laminae can be traced from one clast to its neighbor. Other quartzite and shale clasts essentially are separate from each other.

The C zone is the most easterly of the ore bodies (Taylor and Andrew); it is located in the siliciclastics of the Upper Shallee sandstone. At the time of Taylor and Andrew's work, it was the least understood of all the ore zones. This ore body has a strike length of 275 m and has the shape of a rough triangle. It lies against the hanging wall of a major northwest-trending fault that brings Devonian rocks down against Silurian flagstones and mudstones. The ore body is composed of two or three thin (3 m) stratiform mineralized sandstone horizons that are upturned and coalesce against the fault zone. The main part of the ore body is about 10 m thick and the beds in which the ore is located are drag-attenuated, fractured, and brecciated pale sandstones and grits with which are interspersed minor Silurian slate breccias. The sandstones contain disseminated galena and pyrite; these are cut by anastomosing veinlets up to 15 cm wide of dark to pale, fine-grained sphalerite with some galena and pyrite. The veins show well-developed colloform textures.

Several other, but minor, occurrences of ore are known in the area, but these are of no economic importance, at least at present.

Taylor and Andrew believe that their major contribution to the geology of the Silvermines ore bodies was their determination that the controlling structural feature in the location of the ore bodies was the northwest-trending faults. These faults not only set the pattern of ore-fluid introduction but also contributed largely to the breaking of the ground in which a large portion of the ore bodies are contained. From this it follows that the east-west (or ENE-trending) Silvermines fault (or fault zone) exerted (in their opinion) only a rather minor and quite early influence on the distribution of the ore bodies.

They put forward the following items of evidence that the locations of the ores were related to these northwest-trending faults. The first is that the ore bodies are in close spatial proximity to the NW faults; the Upper G actually lies against a major NW fault and the B zone is cut by one; the Lower G and K zones lie directly against such a fault. Further, the lead-rich vein deposit at Shallee was cut by such NW faults. The second evidence is derived from their contouring of the zinc and lead percentages and of the zinc-lead ratios in the ores. In the upper portions of the G and B zones, the ore bodies changed from a lead-rich area through to a zinc-rich one on both sides on a NW fault. Only one major trend ran in a northeasterly direction, and these authors think that this was due to troughs in the topography of the Carboniferous sea floor and not to fault directions. In the K zone, the trends showed by the ratio plots are definitely northwesterly; further, the K zone had a linear trend that was parallel to the NW fault. They also contend that the element zonation in the ore was related to the northwest trend. The third evidence was that of what these authors believe to have been remobilization. In the K zone in the Lower dolomite, they report that remobilization occurred next to the NW fault. At the fault, the ore mineralization was more or less 40 m thick, but it had dropped to 3 or 4 m where the ore was emplaced in breccias or veins, such structures having been produced by later movements. The evidence in this zone, however, is far from fully developed, so the authors think that more examples of NW faults remain to be proved. The fourth line of evidence was the minor mineralization in veins in the Silurian and Devonian rocks, these veins being related to NW-trending faults; faults with other trends did not have mineralization associated with them. They believe that these NW trends were active as early as Waulsortian times.

In the 1979 discussion of their 1978 paper, Andrew reports that, after publication of that paper, a new high-grade pod of polymetallic silver mineralization has been discovered; it cuts the massive pyrite of the B zone and assayed generally up to 1 per cent silver and locally as much as 6.5 per cent. This ore included several of the arsenic-ruby silver minerals such as proustite, xanthoxonite [Ag_3AsS_3], smithite [$AgAsS_2$], and argyrodite [Ag_8GeS_6] in association with galena, sphalerite, barite, and argentite. This ore also contained 1 per cent nickel; this is obviously quite different material from what had previously been known from the Silvermines ore bodies and merits further study.

In the 1979 discussion of the 1978 paper, C. J. Morrissey pointed out that, to him, the patterns of metal distribution pointed to a subtle interplay of two or more directions of structural control of which he believed the ENE direction to be the most important. This suggests that the dominance of NW-fault control has not yet been firmly established. He contends that the concentration of the ore at Silvermines along only a short portion of the 40 km length of the Silvermines fault indicates that the mineralization may have been concentrated by NW faults but also may have been localized where the main ENE fault had its greatest throw and/or its greatest disruptive effect on the rock adjacent to the fault.

Taylor and Andrew propose a rather complex genetic scheme to account for the various types of mineralization present in the Silvermines district and for the possibly different times at which the ore-forming events took place. They believe that, in Tournaisian time, Zn-Pb-bearing solutions in which the metals were carried in chloride complexes came up along, and out from, the Silvermines fault zone. These solutions are presumed to have met sulfate-rich connate water when they entered pore spaces, solution cavities, and diagenetic features in the Lower dolomite. Obviously, the precipitation of sulfides in such sites requires that the S^{+6} of the sulfate ions have been reduced to the S^{-2} of the various sulfides (or the S_2^{-2} of pyrite). Yet they also say that in Devonian and Lower dolomite ore zones the sulfides were precipitated from deep-seated solutions at about 150° to 300°C in which sulfides the delta-^{34}S values ranged between +4°/₀₀ and -8°/₀₀. Now certainly these two statements are difficult to reconcile. The sulfur is said to have come from a deep-seated source yet also to have come from the reduction of sulfates present in solution in various types of openings in the Tournaisian (Lower dolomite) rocks. To have been lying in wait for the metal-chloride complexes, the sulfate solutions (granted they came from deep-seated sources) must have arrived in the Tournaisian rocks of the Silvermines district at some measureable time before the chloride complexes and must have been found by these chloride solutions with unerring accuracy. If the sulfides ion already were present in these water-filled openings, then the amount of sulfides that could have been deposited would have depended on the amount of sulfide ion available. Even if these solutions were saturated with such ion, the quantity of it available would have been capable of precipitating sulfide in sufficient amount to have filled only a small fraction of each water-filled cavity available. Further, if these sulfate ions were from a deep-seated source, there seems no compelling reason why they should have been concentrated in open space only in Devonian and Tournaisian Lower dolomite rocks. Surely, the entire system of cavity-contain rocks from Devonian to Visean age would have been filled with these sulfide-rich solutions; such filling cannot be expected to have been confined to the older rocks specified by Taylor and Andrew. Even if these sulfate-rich solutions were present and were found by the metal-chloride-brine-rich deep-seated (and later) ore fluids, what mechanism would have been available to have reduced the sulfate-sulfur to sulfide-sulfur? Certainly, the mechanism could not have been bacterial for Taylor and Andrew say that the sulfides were deposited at temperatures ranging between 150° to 300°C, temperatures at which no sulfate-reducing bacteria could have survived. None of the questions posed immediately above has been answered by Taylor and Andrew.

Setting these deposits in older rocks aside for the moment, Taylor and Andrew say that the stratiform ores bodies present in the early Visean beds were precipitated in an early Visean tectonically unstable reef-bounded lagoonal complex in which variations in Eh and pH account for the differences in sedimentation and ore-mineral precipitation obvious in these Visean rocks. Here, perhaps, appeal can be made to sulfate-rich solutions trapped in cavities in the early Visean rocks, ineffective as this mechanism almost certainly would be to supply the amount of sulfide ion needed to attack the metal-chloride complexes and convert them to the sulfides and sulfosalts known in the ore bodies in these rocks. This is even less possible as a mechanism when it is remembered that Taylor and Andrew consider that this precipitation of sulfides took place during early diagenesis. The only way out of this contretemps is for the sulfides deposited during early diagenesis to have been remobilized later in the lithification cycle (or, more probably after that cycle had been completed). Actually, Taylor and Andrew do explain all ore-mineral structures that might be taken to indicate epigenetic emplacement as having taken place by this remobilization process. They do not, however, explain the mechanism of such remobilization; they do not answer such questions as how large amounts of sulfides were dissolved, moved only short distances (measured in fractions of a meter in most instances), and then reprecipitated. This is a particularly difficult problem when the presence of sulfosalts in the ore is considered. At the temperatures proposed by Taylor and Andrew for this remobilization (120° to 200°C) the solid solution of, for example, chalcopyrite in sphalerite hardly could have taken place. Yet the common presence of exsolution blebs of chalcopyrite in most sphalerite shows that temperatures were high enough for an appreciable solid solution between these two sulfides to have occurred.

Sulfur-isotope studies, as would be expected, show that the delta-^{34}S values of the sulfur in the barite is between about +15°/₀₀ and +21°/₀₀, which may be accounted for in one or a combination of both of the following mechanisms. The delta-^{34}S values of the sulfur in barite are such that some (though probably not all) of it could have been produced by reaction between Ba^{+2} provided by the deep-seated ore fluids and sulfate ion present in near-surface sea water or by reaction between Ba^{+2} from deep-seated ore fluids and SO_4^{-2} developed by the oxidation of the S^{-2} ions that may have accompanied the deep-seated ore fluids. It must be remembered that by far the major portion of the sulfate ion in sea water is produced by the oxidation of the sulfur in H_2S or in metal-sulfur complexes when they reach the surface or when they are oxidized on their upward journey by the wide variety reactions possible within the ore-forming fluid or between that fluid and minerals deposited by earlier-passing volumes of the ore fluid or between that fluid and the rocks through which it passes. In short, the presence of sulfur in the sulfate ion of barite with delta-^{34}S values higher than +15°/₀₀ does not tell much about where the sulfur came from except that it ultimately came from igneous activity within the earth's crust.

Taylor and Andrew claim that all the barite in the stratiform and Shallee ore bodies was precipitated directly by reaction with sea-water sulfate. Further, they argue that the pyrite in the stratiform ore zones was formed early in the diagenetic stage by biogenic reduction of sea-water sulfate. They believe that the presence of appreciable quantities of siderite in portions of these deposits depended on the Eh-pH relationships in the lagoonal basin. The temperature range that apparently obtained during the formation of this pyrite (120° to 200°C) almost certainly means that biogenic reactions could not have been the mechanism for producing the S_2^{-2} of pyrite. It is much more likely to have been developed in the upward journey of the deep-seated fluids. The recent work by Larter and his colleagues (1981) resulted in the discovery in the pyritic footwall horizon of the southeast corner of the Ballynoe barite open pit what they call "preserved hydrothermal chimneys" that they consider to be similar such chimneys now known in the East Pacific Rise area at 21°N latitude off the southern tip of Baja California. These modern chimneys appear to be physically and chemically unstable and were presumed to be ephemeral (Larter and his colleagues say "ephermal"). Other recent work, too complex to be cited here, appears to indicate that minerals similar to those of which massive sulfide deposits are composed, when poured out on the sea floor from such "chimneys" probably are too easily oxidized to produce deposits of the magnitude of major massive sulfide deposits (though they may bear a genetic relationship to manganese nodules). At any event, these chimneys beneath one corner of the Ballynoe barite deposit indicate that they probably were produced by ore fluid that, on moving near to the surface, deposited these pyrite chimneys. At Ballynoe, these chimneys consist of circular to petaloid pyrite rods; these rods are 0.1 to 20 mm in diameter and up to 200 m high. In some places, rods will coalesce to form sheaves or curtains. The rods and sheaves normally are enclosed in matrices of crystalline barite. In addition, this matrix of barite and disseminated pyrite, contains patches of crystalline and microcrystalline pyrite, galena, and sphalerite. The pyrite and sphalerite also are present in framboids; the galena always is euhedral. Barite normally is in tabular crystals more than 10 mm across that, in many places, stand apart from the matrix; Larter and his colleagues suggest that this is due to the removal in solution of anhydrite that once cemented them.

When these authors studied polished sections, they found that some pyrite rods had a central 'canal' zone of fine framboidal pyrite; in other examples, the 'canal' is empty. The majority of the chimneys have a clear concentric zoning of radiating crystals of pyrite, separated by zones of fine crystalline and framboidal pyrite. In some rods where the matrices are dominantly sphalerite and galena, these do not have the concentric zoning that is seen in rods in which barite is an important mineral.

Larter and his colleagues believe that these chimneys in the Ballynoe barite are so much like those being developed on the East Pacific Rise that they think it reasonable to suppose that the two types were formed in the same way. The chimneys began to grow over microvents, in either case, supplying a vent for the rising plume of sulfides (and at Ballynoe of barite). As these chimneys grew, a dense uncemented pyrite and water mixture dripped down the outside of the chimneys and formed globular masses on the outside of that structure. On coming out of the chimneys, the solutions that rose through them contained barium ions that reacted with sea-water sulfate to form barite that deposited as a hydrated barite gel. This gel was then moved by bottom-hugging currents to form a sedimentary mass of the gel that was later to be dehydrated. What sulfide may have accompanied the barite probably was, in the authors' opinion, oxidized rapidly (as has been suggested above) and was taken out of the barite-rich area. These authors further suggest that the lead and zinc ions so liberated were redeposited when they reached a deeper part of the lagoonal basin where large amounts of hydrogen sulfide were available. This may have been a sufficient mechanism to deposit the lead and zinc sulfides but something must have oxidized the S^{-2} to produce S_2^{-2} required for pyrite. The authors make no distinction between water depths in the Ballynoe sea and those of the ocean around the East Pacific Rise.

Of course, and in my opinion, an equally acceptable possibility, is that the solutions coming out of the pyrite chimneys remained as solutions and attacked the carbonate rocks into which they were vented, massively replacing them.

Obviously, the majority opinion is that the Silvermines ores were formed syngenetically despite the evidence of the introduction of most of the sulfides epigenetically into the positions they now occupy. This apparent epigenesis of sulfides is, of course, explained as resulting from remobilization of sulfides deposited originally in the unlithified beds and later, after lithification, dissolved and redeposited epigenetically. I am of the opinion that original epigenesis is the better explanation. On this basis, the Silvermines ore bodies were formed under mesothermal to leptothermal conditions and are so classified here. This classification is in agreement with Evans (1976) who says that, at Silvermines, the Upper G ore zone presents the nearest approach to a stratiform massive sulfide deposit. He goes on to say, however, that the lower ore body in that zone is more discordant, coarser-grained, and largely as fracture fillings. He thinks that the ores in the older mines at Silvermines are even more epigenetic in character.

If, on the contrary, the Silvermines ores were deposited syngenetically and later considerably remobilized, they should be classified as Sedimentary-A1a & b, plus B2.

Because the Silvermines ores in Devonian rocks were introduced after those beds had been lithified and faulted and because the major mass of mineralization, which may have been syngenetically or epigenetically introduced, is located in lower Carboniferous rocks, the ores must be categorized as Late Paleozoic.

TYNAGH, COUNTY GALWAY

Late Paleozoic (primary) Late Tertiary (secondary)	Lead, Zinc, Silver, Copper, Barite	Mesothermal to Leptothermal (primary) (Chemically concentrated from fluid igneous emanation - possibly) Ground Water-B2 (secondary)

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Notes

The unusual base- and precious-metal mine of Tynagh (53°09'N, 8°22'E) is located near the village of that name, some 48 km east-southeast of Galway on the Atlantic, 65 km north-east of the airport at Shannon, and 145 km west of Dublin. The area was one of those recommended to prospectors by Murrogh O'Brien when he was Director of the Geological Survey of Ireland. Irish Base Metals (an Irish subsidiary of Northgate Exploration Limited of Canada), using soil sampling, outlined a strong and large-sized anomaly and an electromagnetic survey confirmed the existence of a conductive body. Two inclined diamond drill holes passed through dark, unconsolidated mud that assayed well in lead, zinc, and silver with a minor quantity of copper; the bed rock under the unconsolidated mud was limestone that contained only minor amounts of lead and zinc sulfides. Later core drilling, in the plastic muddy material, developed the volume of this mud as about 1000 m long with a maximum width of 200 m and a maximum depth of about 75 m. Much diamond drilling into the bed rock indicated several million tons of primary ore consisting of galena, sphalerite, pyrite, and subordinate copper minerals with maximum contents of 5000 ppm Pb and Zn. The cover of glacial overburden above the ore-bearing mud ranged from about 3 to 15 m in thickness. This overburden was uncovered, and mining began in this most unlikely ore in 1965. In January 1974, mining in the secondary ore was stopped when the ore came 80 per cent from primary material and 20 per cent secondary; mining in 1978 was entirely underground. By the end of 1974, production had amounted to 317,000 tons of lead, 226,000 tons of zinc, and 284,000 kg of silver, plus small quantities of barite. Barite is now being recovered at a rate of 50,000 tons annually from the mine tailings. At the end of 1970, reserves were reported to be nearly 6,000,000 tons of ore with a grade of 5.37 per cent lead, 4.51 per cent zinc, 0.37 per cent copper, and 60 g/ton of silver. A further 3,000,000 tons of somewhat lower-grade ore had been blocked out at that time (Williams and McArdle, 1978).

The primary ore is reported by Boast and his colleagues (1981) to have been composed of two sulfide deposits (zone II and zone III); these were separated by a barren zone of dolomitized limestone. The combined tonnage of these two ore zones was 9.4 million short tons; average grade of this ore was 3.0 per cent Pb, 3.2 per cent Zn, 0.3 per cent Cu, and 1 oz. Ag per ton. The block of Waulsortian limestone containing the ore was 200 m vertically, 1.6 km in length parallel to the strike of the Tynagh fault, 130 m perpendicular to that fault.

Although the ores of the central plain of Ireland are contained in Carboniferous rocks, they are underlain by those of the lower Paleozoic, including parts of the lower Devonian, that were folded strongly during the Caledonian orogeny (Schultz, 1966). The erosion of these Devonian rocks left a surface of minor hills and valleys over which the Carboniferous rocks are thought by some to have been draped. The Carboniferous sea advanced from the south to lay down basal clastics and then the main limestone rocks of the lower Carboniferous that principally are shelf-type carbonates of Dinantian (Tournaisian plus Visean) age. The mid-Dinantian also was a time when volcanic eruptions took place in several areas of Ireland; these flows were largely from basaltic to intermediate composition with intercalated pyroclastic sediments, both types being interbedded with the limestone sequence. The closest of these volcanic areas to Tynagh was in County Limerick, nearly 70 km south of Tynagh. Local structural disturbances about this time produced submarine slide conglomerates and breccias; these lead to moderate regressions of the sea and an increase in the supply of clastic sedimentary material from the uplifted source areas. Schultz believes that this volcanic episode was the climax of a much longer period of crustal instability and moderate earth movements. The Tynagh iron formation (of vital importance in the understanding of the geology of the Tynagh ores) appears to have been developed during the volcanic epoch as a result of the various processes taking place during that time.

The Carboniferous period ended with the Hercynian orogeny; Ireland was on the north front of the Variscan (i.e., Hercynian) folding, but the most intense folding in Ireland took place well south of the Tynagh area. In the part of Ireland in which Tynagh is situated, the folding was appreciably less severe and was strongly influenced by the structure of the underlying Caledonian rocks. What regional metamorphism took place was confined to the fold belt of southern Ireland (Schultz, 1966).

In detail (McDermot and Sevastopulo, 1972), the base of the pertinent portion of the Tynagh stratigraphic column is provided by the Old Red sandstone (the base of which is not observed); it is from 185 to 370 m thick. This is overlain by nearly 50 m of Lower limestone shales (of which the basal portion is 4.3 m of Ballyvergin shale); the Lower limestone shales proper are 45 m thick. Over the Lower limestone is 3 m of the Tynagh sandy bed, and that is followed by 125 m of the Bioclastic limestone and 200 m of Lower Muddy limestone. These last two formations appear to be the equivalent of the Lower limestone and Lower limestone shale of Schultz (1966); the total thickness given by Schultz for these two formations is between 230 and 290 m, not a huge difference from the figures just given especially if the 328 m total is a maximum thickness. These two formations are Tournaisian in age. The rocks immediately above the Lower limestone are known variously as the Waulsortian, Bank, and Reef limestones with the lower portion being still Tournaisian and the upper part being assigned to the lower Visean. The thickness of this variety of limestones is stated by McDermot and Sevastopulo (1972) to be, at a maximum, 170 m. In his generalized stratigraphic section, Schultz (1966) gives the thickness of the Waulsortian or Bank limestone to be 0 to 185 m. McDermot and Sevastopulo divide the Reef limestone into four parts, to which they do not assign (or Williams and McArdle do not quote) actual thicknesses. The lowest part is a slump breccia, the next the main calcilutite (in which is contained an horizon of slump breccia) then comes another (apparently more wide-spread) horizon of slump breccia, and finally a bed of crinoidal biomicrite, about as thick as the bed of calcilutite. The last formation given by McDermot and Sevastopulo is the Calp limestones (more than 250 m thick). This appears to correspond to Schultz Middle limestone. The Calp is divided into a lower non-siliceous Calp and an upper siliceous Calp. Williams and McArdle say that the Calp is more than 250 m thick, whereas Schultz says that his Middle limestone is up to 240 m thick.

Schultz reports the Waulsortian to be analogous to a barrier reef that is generally thick and spread over much of west-central and southwestern Ireland. The Waulsortian separated basin facies in the south and west from lagoon facies to the north and east. The rocks in the lagoonal facies are similar to the reef materials, but they occur in discrete knoll or bank deposits. These limestones are poorly stratified and are made up of calcite mudstones that include various proportions of bioclastic material that ranges from finely broken debris to coarse crinoidal and shelly material. Characteristic of these limestones are carbonate muds deposited by internal sedimentation with sparry calcite as cavity fillings. Although bryozoan fronds locally are rather common, Schultz does not consider them to be abundant enough to have provided rigid frameworks for the reefs. He also does not think that these rocks are true reef-facies limestones because they do not form rigid wave-resistant structures. Lees (1961) describes the reefs as "carbonate mudbanks that were formed by accumulation of carbonate mud and bioclastic sand which were trapped and held in place by organisms, presumably plants or sponges, which have not been preserved in fossil form". The Tynagh area proper is located in the northern part of a large bank complex that reaches from this area almost to the city of Cork, and covers 7750 km^2. The overlying Calp limestone is composed of massively bedded clastic limestones that include interbedded shales and themselves contain much argillaceous and carbonaceous matter. Although the bulk of the Calp lies above the Waulsortian, the two formations also are partly contemporaneous. The Middle (Calp) limestone contains a few tens of meters of cherty beds; in places, thin beds of pyritic black shale occur through the formation. Some of the Calp limestones are red (due to fine-grained hematite intimately mixed with the carbonate mud) and green (due to chlorite). Both hematite and chlorite may occur together. In the Tynagh area, these colored limestones are associated with ironstone sediments. Schultz believes that the hematite was transported as colloids or in true solution and was deposited independently of the carbonate material.

The district (Schultz, 1966) is part of the moderately deformed foreland of the Variscan fold belt slightly over 110 km north of an east-trending thrust fault that marks the northern limit of the highly deformed rocks of the Armorican geosyncline.

The principal structure in the Tynagh area is an inlier (the irregularly shaped Slieve Aughty mass of Old Red sandstone) that contains some Silurian inliers as well. The major fold axes strike between east-west and NE-SW with dips that, in only a few places, exceed 15° except on the north slopes of Slieve Aughty (the slope away from Tynagh) where dips up to 50° have been measured in the Lower limestone.

In the Tynagh area proper, the reef limestones are a complex group of in situ bank limestone, reworked and related limestones, and detrital material derived elsewhere than from the Waulsortian banks. In this area, the Bank limestone is in contact with Old Red sandstone along the Main (North) Tynagh fault.

Within the lower part of the Bank limestone, away from the Main fault, the limestones grade into (intertongue with) ironstones that are at the same level as reddish or greenish limestones in the Bank limestones. Schultz considers the ironstone to be a facies of the lower Bank limestone; its southern edge is about parallel to the trend of the Bank limestone for about 1 km, but then it swings in a more northerly direction than that followed by the limestone. The minimum dimensions of the Iron formation (Schultz, 1966) are about 1.5 km east-west and at least 0.8 km north-south; it may be much larger. The maximum thickness of the ironstone is up to 55 m north of the Tynagh fault. The Iron formation is composed of alternating and interfingering layers of ironstones and limestone with graded bedding. There are a few thin beds of tuffaceous rocks in the upper parts of the Iron formation. The ironstones are (to Schultz) clearly of sedimentary origin and were formed by the chemical (biochemical?) precipitation of iron oxide and silica in a restricted marine basin. Most students of the district consider the ironstones to have been formed by volcanic-exhalative processes. The accompanying clastic sedimentation was provided by intermittent, short-lasting influxes of limestone detritus carried in by subaqueous slides and turbidity currents. In addition to the hematite and chert, dolomite is common in aggregates of small grains. The original carbonate probably was calcite that was diagenetically domomitized.

An iron-rich chlorite is present as a rather abundant constituent of the ironstone; stilpnomelane is a minor mineral in the ironstone and is difficult to distinguish from chlorite in hand specimens, but is easily recognized in thin sections. Minnesotaite also is found in subordinate amounts, generally with, or close to, stilpnomelane. Both of these two minerals have replaced chert and turbid dolomite. Some short sections of some ironstone beds contain enough magnetite and maghemite that they are highly magnetic. Late calcite occurs as minor fillings of pores in coarsely crystalline hematite; some calcite replaces chert and/or chlorite. Schultz believes that this $CaCO_3$ was derived from adjacent limestones. Pyrite is present in the ironstone in very small quantities except for one section of minor extent at the top of the formation. Overlying this section is limestone breccia that also has disturbed the ironstone beneath it.

Hematite and chert are the essential minerals of the ironstone, whereas the turbid dolomite, chlorite, and stilpnomelane are varietal constituents.

The ironstones show different forms of layering: (1) finely laminated, wrinkly structure (algal bedding), (2) poorly layered structure, characteristically showing strongly developed colloform features, and (3) massively bedded structure.

Interbedded limestone, a minor part of the Iron formation, was formed from the same detrital material found in the Waulsortian unit; reworked pyroclastic material makes up a very subordinate part of the total formation.

The exact age of the iron formation is unknown, but it probably is mid-Dinatian, about the boundary between the Tournaisian and Visean. Development of an off-shore barrier somewhere north of the Tynagh area cut off that region from free communication with the open sea and, thereby, from the entry of muddy and silty sediments from the north. Clastic sedimentation was confined to areas near shore or where older sediments had been uplifted and then reworked by wave action and currents; some parts of the ironstone basin were largely untouched by clastic sediments except for intermittent supplies of slide-transported material.

Schultz does not believe that the iron and silica of the Iron formation came from a volcanic-exhalative source (or sources). Nevertheless, Schultz says that the possibility of minor hot springs in the basin floor cannot be disproved by present knowledge. For what it is worth, the large majority of students of the Tynagh ores believe that the iron formation was produced from volcanic exhalations (Boast and others, 1981, Russell, 1975, Riedel, 1980, Derry and others, 1965).

Derry and his colleagues (1965) and Derry (1966) have referred to the greenish-gray silty beds that grade laterally into the Reef or Bank or Waulsortian limestone as "tuff". Schultz objects to this (1966) because he says that, in thin section, he finds no specific mineralogical or textural evidence of "volcanic provenance". Nor does he find that any appreciable amount of this mineral is different from the terrigenous debris such as may occur in impure limestones at any level of the lower Carboniferous succession at Tynagh. Schultz says that, out of the 60 m of total thickness mentioned by Derry and his colleagues (1965), no more than 1.5 m can be classified on reasonably firm petrographic grounds as being of pyroclastic origin. However, Derry (1965) says that all he wants to do is point out that there was some volcanic activity in the area at much the same time as the formation of the ores.

Another area of disagreement between Schultz and Derry and his colleagues is as to the structural sequence of events. Schultz claims that tectonic folding (of Hercynian age) was influenced by the Caledonian grain of the older (Devonian and Silurian) rocks and that this folding, as well as normal faulting, accounts adequately for the present structure. He says that there is clear evidence of Hercynian folding, although, as Derry points out, he does not say what the "clear" evidence is. Further, Schultz holds that the nature and facies distribution of the Old Red sandstone and the Carboniferous sediments do not fit the postulated "ridge and valley" structure that should have been imparted by the pre-Old Red sandstone topographic relief. Moore (1975) contends that minor structures associated with the (North) Tynagh fault suggest that there have been several episodes of movement along the fault after the initial major and normal displacements. The major normal movements were followed by reverse oblique-slip faulting that was accompanied by subsidiary thrusting, cross-faulting, and minor strike-slip displacements. The introduction of the primary ore bodies, Moore claims, preceded the reverse movements that probably were Hercynian. These movements caused minor and local redistribution of small amounts of the gangue and ore minerals. Granted that these post-ore fault movements did take place after the ore had been introduced, the age of the mineralization must have been before these minor fault movements, but this does not prove that the major fault was not an early expression of Hercynian earth movements, and the later minor faulting no more than "after shocks" that followed the earlier major one.

The Tynagh fault proper trends about east-west to E10°S and dips to the north at 60° to 65° at the mine workings, but decreases to 45°N to the east. In this area, the Tynagh fault seems to have been offset in several locations. In the western part of the open pit (from which the residual ore was mined), the fault splits, and movement took place on several slip surfaces. In the western part of the surface workings, a tectonic slice of Muddy limestone (the formation directly underlying the Reef or Bank) has been intercalated within the fault zone between Devonian sandstones in the footwall and the Reef and Calp limestones of the hanging wall. The ore bodies are contained in younger rocks of the hanging wall.

There has been a net normal displacement across the main fault zone as is shown by the lower Carboniferous rocks that overlie the Devonian beds in the footwall. Although the displacement differs somewhat along the fault, in the mine area the slip exceeds 600 m in dipslip; this net slip is the result of the summation of all movements that took place along the fault since the first movement that affected the Carboniferous beds. For a considerable length of the fault, the footwall is composed of an inlier of Old Red sandstone. On the north side of the fault, there are 150 m minimum of normal Reef limestone between the fault and the southernmost tongue of the Iron formation. The Reef limestones generally show some evidence of slump brecciation. Moore thinks that tectonic uplift during sedimentation may have induced the slumping. From this Moore believes that it follows that the Tynagh fault may have been a pre-Carboniferous basement structure reactivated so that it affected Carboniferous sedimentation and was influential in the localization of an ore body of syngenetic or diagenetic origin.

Williams and McArdle (1978) say that the Carboniferous and Devonian rocks on the southern side of the Old Red sandstone inlier are conformable. Thus, they contend, the main Tynagh fault is a trap-door fault (a curved fault bounding a block that is hinged along one edge). They are convinced that there were at least three main phases of movement, all of which affect the mineralization; localized gentle folding also was spatially associated with the faulting.

Moore says that few identifiable minor structures of tectonic origin remain to indicate what was the early history of the Tynagh fault. He says there are two possible explanations for this situation: (1) that the normal fault movements occurred between the sandstone of the footwall and unconsolidated Carboniferous sediments on the north side of the fault and (2) that the large number of younger, superimposed minor structures, which were the products of later movements, obscured the earlier relationships. Under (1), of course, the usual structures found in lithified sediments would not have formed in the soft calcareous sediments or would have been destroyed by differential compaction during and after the movement. Moore attempts to explain essentially all the veinlets in the Carboniferous rocks as having been produced during diagenesis. On the contrary, these minor veins may be the result of the later and minor movements on the fault, after the Reef limestone was essentially, if not entirely, consolidated.

The Reef limestone, cut by the Tynagh fault, contains the ore. These micritic and bioclastic limestones usually are separated from the surrounding muddy and carbonaceous limestone of the Lower Muddy limestone and the Calp by transitional slump breccias. Near the core of the reef, the breccia contains numerous limestone blocks in a shale matrix, a condition that suggests that the original material prior to brecciation was lithified interbedded limestone and shale and, on being subjected to tectonic stresses, the limestone broke into solid fragments and the shale flowed into the spaces between them. This explanation hardly conforms to the concept of a mass of unconsolidated sediment (carbonate and shale) having slumped before lithification. Of course, movement along the fault after the interbedded limestone and shale had been lithified would definitely have a major effect on the newly lithified rocks of which the slump breccia seems to be a reasonable outcome. It also is possible (Williams and McArdle, 1978) that the slumping reflected a regional depositional instability such as has been noted in upper Tournaisian and lower Visean rocks in Ireland.

Schultz (1966) considers the contention of Derry and his co-workers that, because individual formations tend to thin out as well as steepen in dip as the Tynagh fault is approached from the north, fault movement was more or less contemporaneous with sedimentation and accelerated subsidence in the northern area. Schultz thinks this argument is fallacious because the southward thinning of a given bed (that has its maximum thickness in the north) is compensated for by the thickening of the contemporaneous beds of the southern facies; thus, there is no change in the total thickness of the larger stratigraphic units. He holds that the steeper dips in the vicinity of the fault are due to stronger deformation in that area.

Two strongly contrasting views as to the manner and time of ore emplacement are given by Schultz (1966) and Boast and his colleagues (1981). First, to summarize that of Boast; sulfides occur at the same stratigraphic level and were deposited at the same time as the Iron formation (as many geologists, of whom Derry was the first and Boast and his colleagues the latest, contend). Schultz, however, reports that sulfides occur not only at the same stratigraphic level as the Iron formation but also well above and below it in both the Reef limestone and the Lower limestone and the Calp. The total stratigraphic range of the base-metal sulfides and barite (not always in ore grade and dimension) along over 2100 m of fault length is well over 300 m, essentially throughout the entire lower Carboniferous sequence in the Tynagh area. Minor sulfide-barite mineralization also is observed in the Old Red sandstone of the footwall of the fault. Although mineable bodies of ore, thus far, have been found only in the Reef or Bank limestones, Schultz believes that, except for the shale and very argillaceous limestone, any stratigraphic level that abuts the fault may be well mineralized.

Schultz also points out that the Iron formation and the sulfide-barite deposits show quite different modes of depositon. The Iron formation almost certainly (in both the opinion of Schultz and of Boast and his colleagues) was sedimentarily accumulated on the sea floor, there undergoing later diagenetic modifications. The hematite in the red limestones also is sedimentary (syngenetic) with the carbonate rocks. Further, authigenic pyrite must bear a sedimentary-diagenetic relationship to the limestones.

Schultz adds, however, that, in the primary sulfide-barite deposits, no such syngenetic relationships to the host limestones are to be seen. As described by Derry and his co-workers (1965), the ore minerals normally exhibit features typical of cavity fillings, and these cavity fillings are in post-lithification fractures, breccia voids, and solution cavities that cross cut Stromatactis and other primary and diagenetic features, including late-diagenetic recrystallization fabrics. Layered structures in the sulfides exist at any orientation from parallel to normal to observed geopetal fabrics in the primary sediments, including those structures typical of those in the Iron formation. Whereas deposition in the Iron formation is repetitive, that in the sulfide-barite ores is not.

Also, there has been much actual replacement of limestone (normally first altered to iron-bearing dolomite) and of older ore minerals by younger ones. When the ore bodies are followed from the weakly mineralized fringes to the high-grade centers, all gradations can be seen from thin sulfide linings in calcite-filled solution cavities and veinlets in unreplaced limestone to massive ore in which most of the original limestone has been replaced by sulfides and barite.

Schultz considers the order of deposition to have been: (1) replacement of limestone by iron-bearing dolomite, probably partially contemporaneous with cavity-filling pyrite-marcasite; in this altered limestone, authigenic and framboidal pyrite have been segregated, with other non-carbonate impurities, at dolomite crystal boundaries - this iron-bearing dolomite is older than the ore minerals that replace it, thus making the ore minerals even older than the authigenic pyrite; (2) pyrite-marcasite mainly as colloform and crustified cavity linings and fillings - much fractured before the deposition of the other sulfides - marcasite now very minor but much of the original marcasite has been converted to pyrite; (3) sphalerite - microgranular aggregates are common but some cavity fillings have colloform textures; adjacent to the cavity fillings both the iron-bearing dolomite and limestone have been extensively replaced by sphalerite as have those minerals where they form the wall rocks of fractures; (4) barite, overlaps somewhat with sphalerite - locally achieves very extensive replacement of carbonates - weight percentage of barite about double that of sphalerite and galena combined; (5) copper minerals, first chalcopyrite, then the much more common tennantite-tetrahedrite - locally enough copper minerals to make copper ore; (6) galena - very broad overlap with barite, and (7) calcite, locally late calcite has replaced in part all sulfides but most affects sphalerite. This late calcite (7) was followed after post-mineral fracturing, by an even later calcite - these fractures cut both mineralized and unmineralized rocks equally. Minor arsenopyrite precipitated near the end of the iron di-sulfide stage. Many sulfosalt minerals (such as bournonite, enargite, meneghinite, pyrargyrite, polybasite, and stephanite, plus others not identified) are in microscopic quantities largely in galena and tennantite-tetrahedrite.

Sulfide-barite mineralization was most intense in pure limestones, particularly the fine-grained type, that lack argillaceous and carbonaceous matter; these last are poorly mineralized throughout the district except near main and minor faults. The close association of the sulfide minerals to their Carboniferous host rocks argues against the syngenetic depositon of the ores. Had the ores been so deposited, the high sulfide-ion content of the sea water would certainly have reduced and modified the shells produced by the life forms attempting to live in this inimical environment if it had not obliterated them altogether. It must be emphasized (Schultz, 1966) that the ores are not strictly strata-bound, much less stratiform, as is true of the Iron formation. Further, in contrast to the beds of the Iron formation, the distribution of the sulfide-barite mineralization is co-extensive with the plane of the Tynagh fault rather than with the sedimentary facies. Moreover, the extent and the intensity of mineralization decrease with distance from the fault (to the north); near the fault even the least favorable rocks may be fairly heavily mineralized, whereas away from it even the most favorable beds are barren. The ore and sedimentary-rock features are disharmonious to each other on a small scale, whereas the ores are closely related to post-lithification faulting and associated rock deformation and not to the sedimentary events or features.

Schultz points out that a contemporaneous emplacement of the sulfide-barite ores and of the ironstones would imply that the solutions emanating from the sea-floor springs were capable of depositing Fe^{+3} away from the fault but precipitated Fe^{+2} only a few tens of meters away from the area in which the ironstones (with their ferric ion content) were being laid down. Schultz sees no sign of a transitional phase between these two components of the imaginary system proposed to deposit both types of metal-rich material contemporaneously.

Riedel (1980) is in general agreement with Boast and his colleagues as to the manner of the formation of the Tynagh ores, and he relies on much the same items of evidence. The major exception is that Riedel postulates a N-S fracture as the channelway through which the metal-bearing solutions entered the area. This is somewhat the same concept for this entry as I propose later in this discussion, but Riedel considers that all ore-metal bearing solutions came through this N-S fracture into the area over much the same time span as the sedimentary formation of the various carbonate beds were being deposited. This means that, if Riedel is correct, both the Iron formation and the ores were syngenetic or early diagenetic in relation to the carbonate, whereas, I think that both the iron-rich and non-ferrous metal-rich ore fluids came from much the same source, but, coincident with the change in character of the ore fluids from iron- to iron- and non-ferrous-metal rich, the Tynagh fault became available to channel the upward movement of these fluids. Obviously, both of these concepts are irreconcilable, and a decision as to which is correct can be made only after a thorough study of the evidence presented in the various papers cited in the Tynagh bibliography at the head of this discussion.

Riedel considers that the following features point to the correctness of his hypothesis: (1) the location of the center of the Zn-Pb ore body in the center of Zone II (a zone directly related to the Tynagh fault, with its long dimension perpendicular to Riedel's supposed N-S fault but with the long dimension of the ore body being parallel to the Tynagh fault); (2) both the abundant copper mineralization and a complex Pb-Fe-Cu-As-Ba-Zn association are centered where the supposed N-S fault cuts the Zone II ore body;

(3) the highest Hg content in the sphalerite concentrates is similarly centered; (4) the maximum thickness of the Iron formation also is similarly centered; (5) a depression valley formed at this center during the development of the Iron formation; and (6) the maximum Mn content in the limestones also is so located.

Of course, there must have been a line of hydrothermal springs during the development of the Iron formation, a line that oriented the Iron formation in a generally N-S direction and channeled manganese locally into that formation. But these springs were superceded as a channel for rising hydrothermal fluid with the development of the Tynagh fault, explaining the E-W orientation of the ore bodies. With the exception of the non-ferrous metals, the ore fluid that came up along the Tynagh fault was much the same in composition as that that reached the sea floor through the line of earlier hot springs as is indicated by the presence of much the same accessory and trace elements in both Iron formation and ore.

Riedel dismisses the obviously epigenetic characteristics of much of the ore as the results of ore fluids moving (percolating) through the unlithified sediments thereby destroying primary sedimentary structures. Surely, it is equally possible that these textures were destroyed by metasomatism caused in the lithified carbonate rocks by the later ore fluids. Riedel fails to realize that ore fluids of high-enough temperatures to deposit the minerals of the primary Tynagh ore would have been so hot that; (1) if they did reach the sea floor in liquid form, would have killed bacteria capable of reducing S^{+6} and (2) probably would not have reached the shallow Tynagh sea without boiling and dumping their loads within the older solid rocks through which they passed (Ridge, 1973). In short, the mineral suite of the Tynagh deposit could not have formed if the ore fluids had brought the non-ferrous metals to the sea bottom during the period of carbonate sedimentation. Instead, the rock volume being metasomatically mineralized must have been covered with a considerable thickness of lithified carbonates (see Ridge, 1973) if the ore fluid was to have reached the rock volume where the ores were deposited without having suddenly lost their ore-metal loads at an appreciable depth below the sea floor.

A further argument for the epigenetic introduction of the sulfide ores is given by Williams and McArdle (1978) who point out that the close association of metallic sulfides with the large and varied suite of fossils found in the Reef limestones demonstrates that the ores could not have been introduced at the same as that at which the fossils were living. Had the sulfur of the ore sulfides been formed by bacterial reduction the large quantity of metals and sulfur that must have been present during ore deposition certainly would have had a strongly deleterious effect on the organisms producing the fossils, probably killing most of them off before they could have developed on the large scale in which their fossil remains occur in the Reef limestone and certainly badly stunting their growth, a condition of which no evidence is known.

Boast and his colleagues (1981), on the contrary, suppose four stages in the formation of the Tynagh ores: (1) growth of colloform and granular pyrite clots during early diagenesis of the micritic host; (2) rapid geopetal precipitation of microcrystalline sphalerite, galena, barite, and dolomite within a dilatant fracture system that developed as a consequence of the forceful injection of metalliferous fluids into the coherent (partially lithified) Waulsortian mudbanks adjacent to the Tynagh fault; (3) veining and replacement of the host rocks and previously precipitated phases by tennantite, chalcopyrite, bornite, coarsely crystalline galena, and coarse sparry barite, and (4) precipitation of calcite within veinlets and cavities and dolomitization of large bodies of previously unmineralized Waulsortian limestones.

Russell (1975) draws the following conclusions from his study of litho-geochemical environment of the Tynagh deposit: (1) major-element-geochemistry of the Tynagh Iron formation is similar to that of iron deposits of exhalative origin; (2) high trace-element concentrations in the Iron formation, particularly Ba, Pb, and Zn contents of the distal, manganiferous layer, suggest that an exhalative source supplied these metals; (3) the manganese aureole in the Waulsortian limestone is so extensive that an epigenetic dispersion must be ruled out - Russell holds that this aureole reflects manganese enrichment in the sea water of lower Carboniferous time around the Tynagh emanative center; (4) the hydrothermal system was active throughout the time of Waulsortian limestone deposition; (5) the Iron formation, the syngenetic enrichments of manganese and zinc in the Waulsortian limestone, and the epigenetic Pb-Zn-Cu-Ag-$BaSO_4$ deposit were derived from the same mineralizing solutions; (6) the east-west fault zone was in existence in mid-lower Carboniferous time - the fault scarp provided a topographic barrier, largely allowing trace-element dispersion into the sea water only to the north of the fault.

Probably Russell's most important conclusion is that the base-metal deposit is essentially post-lithification. He thinks, however, that this lithification may have been achieved within a few years after the carbonate mud was deposited in the Tynagh basin.

Boast and his colleagues (1981), in an effort to determine where the sulfur in the sulfides and in the barite came from, isotopically analyzed 193 sulfur samples with their greatest emphasis on the sulfides of stage 2 and 3 in their sequence (see above). Stage 2 was, according to Boast, geopetal precipitation of microcrystalline sphalerite, galena, barite, and dolomite. This contrasts with Schultz' separation of what seem to be the same sulfides into: (a) sphalerite, (b) barite, (c) copper minerals, and (d) galena (with very broad overlap with barite). It would seem that Schultz can see a more ordered sequence of precipitation than did Boast, and that Boast considers that the copper minerals definitely followed the geopetal precipitation of his stage (2) but with galena present in that precipitation, whereas Schultz does not admit the formation of galena until after that of the copper minerals. Short of another detailed study of the paragenesis of the Tynagh sulfides (and perhaps not then) can the differences between the paragenetic work of Schultz and of Boast and his colleagues be reconciled. But the results of both the single author and the author group do agree that the sulfides were formed in a narrow time-connected sequence, so that sulfur-isotope work on these minerals (and on barite) may provide the basis on which to judge the ultimate source of the sulfur in these minerals. Isotope ratios that they determined from the stage (2) sulfides closely approximate normal distribution, that is -4.1 to -26.0 per mil. These authors can find no correlation between these isotopic ratios and host lithology, elevation, distance normal to the fault, or distance along strike of the Tynagh fault. Coprecipitated sphalerite and galena (according to their determination) have such different ratios that they cannot be considered to be in isotopic equilibrium. If Schultz' sequence of precipitation is considered to be correct, and coprecipitation is not of major importance, the differences in isotopic ratios from one mineral to the other would mean far less than they do if they were coprecipitated. But either way, there is no denying the wide range of isotopic ratios in the sulfurs in the main stage of sulfide precipitation at Tynagh. Barites deposited in this same stage of main sulfide precipitation have a much narrower range of delta-^{34}S values. This convinces Boast and his colleagues that the sulfur in the barium sulfate came from the water of the sea. Therefore, the barite either was deposited directly on the sea floor or sea water circulated down and mingled with that water that rose from below and carried the barium ion as well as the metal ions of the sulfides. The wide range of delta-^{34}S values in the sulfides suggests to Boast and his co-workers that they resulted from kinetic fractionation of sea-water sulfate and did not derive from sulfide ion brought along with the metal ions. Yet any sulfur brought with the metal ions must have undergone reaction with its surroundings and well may have been oxidized in its upward journey from its source wherever that may have been. If the sulfide ions accompanied the metal ions in complex form, a concept for which I have argued for a long time (Ridge, 1968, 1980), the excess sulfurs must have been removed, and this removal must have been strongly aided, if not entirely controlled, by oxidation of sulfur and concomitant reduction of (probably) ferric ion.

As the Iron formation almost certainly was deposited at the same time as to surrounding Waulsortian (or Reef or Bank) limestone, this indicates that an iron-silica-rich solution was reaching the sea floor before the almost certainly epigenetically deposited sulfides. It is probable, however, that this iron-silica solution came from the same general source as the later sulfide-depositing ore fluid. This ore-forming fluid probably continued to carry abundant ferric iron, as it had done when the Iron formation was being deposited, but most of that ferric iron was reduced in the oxidation of S^{-2} [see equation (1) below] as is witnessed by the abundant iron di-sulfide in the deposit and appreciable iron in the epigenetic dolomite produced from the primary (and sedimentary) limestone. At certain times, the iron-silica solution was quite rich in manganese as is shown by the manganese-rich layer in the Iron formation. Such manganese, if present when the sulfides were being deposited, would have aided in the oxidation of sulfur.

A question yet finally to be decided is: was the Tynagh fault in existence at the time the Iron formation was being developed? Probably it was not or a more definite connection would exist between the Iron formation and the fault. That the iron-silica solutions that produced the Iron formation probably came from the same source as the later sulfide-depositing ore fluids is shown by the abundances of such trace elements as barium, lead, and zinc as were revealed by Russell's work (1975). Later, after much, or probably all, of the Waulsortian at least had been deposited and lithified, the character of these solutions changed and they moved toward the surface through the newly (?) created Tynagh fault to deposit the epigenetic sulfide ores of the Tynagh ore body. This ore body lacks direct spatial or temporal connection with the Iron formation, but it is highly probable that the iron-silica-rich fluids that reached the sea floor to deposit the Iron formation were an earlier expression of magmatic activity at depth that was later the parent of the ore fluids that produced the Tynagh deposit proper.

This new type of solution came up along the Tynagh fault and worked its way out into the rocks on the north side of that fault, depositing the sequence of sulfides, barite, and calcite, in the beds more favorably constituted, chemically or physically, for the reception of such mineralization. The presence of small, but significant amounts of the same mineralization in stratigraphically lower and higher limestones indicates that the actual primary mineralization at Tynagh was appreciably later than the lithification of the Reef limestone (Schultz, 1966; Russell, 1975).

The amount of sulfate ion in the sea water at the time of the deposition of the Tynagh sulfides, if sea water was able to mix with the rising hydrothermal solutions, may have provided the sulfate ion for some of the barite, but most of that ion almost certainly came from the oxidation of sulfide sulfur through its reaction with ferric ion as is shown in equation (1) that follows:

$$(1)\quad 8Fe^{+3} + S^{-2} + 4H_2O = 8Fe^{+2} + SO_4{}^{-2} + 8H^{+1}.$$

The opportunity for any appreciable amount of sea water to work its way down into the channelway provided by the Tynagh fault against the rising hydrothermal (sensu latu) fluid would not have been good even if the sulfate ion content of the sea water had been high enough to have made a significant contribution to the formation of barite could it have come in contact with barium ions.

The question remains, of course, can the fractionation of sulfide sulfur produce delta-^{34}S values in the sulfate sulfur comparable to those that are present in the Tynagh barites? The fact that all of the sulfur in the sea ultimately had a magmatic source shows that oxidation during hydrothermal processes is capable of having profound effects on the delta-^{34}S values, and these well may be sufficient to produce the fractionation necessary to provide the delta-^{34}S values of the Tynagh barite.

In many instances, hydrothermal ore deposits (in the strict sense) have delta-^{34}S values that center closely around the Diablo Canyon standard, but many others, almost equally as certainly deposited from hydrothermal fluids that came from great depths and presumably from magmatic sources, have delta-^{34}S values that range widely from that standard. Normally, these departures from the standard are thought to have resulted from the incorporation of sulfide ions produced by bacterial reduction of sea-water sulfate. In 1980, I attempted to show that the reactions that effect hydrothermal fluids on their way to the earth's surface can change appreciably the delta-^{34}S values with which the ore fluid is supposed to have left its parent magma chamber. It is universally accepted that the ore fluid reacts with the rocks through which it passes, with the minerals it has deposited previously, and within itself, yet it seldom has been suggested (McKinstry, 1957 is an exception) that we do not know enough about this process of fractionation to assume that the reactions that occur in hydrothermal fluids cannot affect it. In some instances, the effects essentailly balance each other, but, in others, the effects appear to be pronounced (Ridge, 1980). Thus, to assume that reactions in hydrothermal fluids did not affect the delta-^{34}S values of the sulfur in that fluid is to make an assumption so gross as to put all conclusions based on it into a realm of considerable uncertainty.

Similar problems obtain when the isotopes of oxygen and carbon are studied, although it is certain that the effects of fractionation during hydrothermal activity are appreciably less than for sulfur simply because sulfur valances in hydrothermal fluids range from 2- to 6+ and those of oxygen and carbon are essentially constant at 2- and 4+, respectively. Despite some suggestions that the oxidation of O^{-2} is carried out in hydrothermal fluids to a major degree, I have yet to see proof that this phenomenon occurs on any major scale in any specific hydrothermal solution unless it be by the dissociation of nitrogen-oxygen complex ions (Ridge, 1956). So far as carbon and oxygen in rocks and ores in a sulfide deposit in carbonate rocks are concerned, the chances for isotope fractionation of either of the elements is so great that any departure from the SMOW values well may have been achieved in the reactions within the hydrothermal fluids or by reactions with the carbonate ions already present in the wall rocks and breccia fragments. The analysis of data on isotope ratios of carbon or oxygen determined on specimens of ore or wall-rock carbonates has so many variables to consider that it is questionable if the mechanisms now exist for the solution of the problem of how the fractionations that appear to have occurred can be made to tell a coherent story as to the conditions under which the ore deposit was formed.

The last few paragraphs are, of course, a degression from the strict study of the Tynagh deposit, but they are pertinent in that they effect the interpretations made from the isotopic ratios obtained by spectrographic analysis of the isotopes of sulfur, oxygen, and carbon in any ore deposit in general or Tynagh in particular.

Boast and his colleagues (1981) conclude from their studies of carbon and oxygen isotopes at Tynagh that a depletion in ^{13}C can be charged to the mixing of host-derived carbon with bacterially oxidized carbon, plus, during their stage (4), the precipitation of calcite within cavities and in veinlets and the dolomitization of previously unmineralized Waulsortian limestone. The ^{18}O depletion of the mineralized host is a consequence of an isotopic exchange with the ore fluids. An oxygen-isotope-fractionation temperature of 200°C and a delta-^{34}O (SMOW) value of 10.5 per mil for the coexisting water is obtained from a single-stage (3) quartz-albite pair collected from the fault zone. If these values are accepted, they say, the delta-^{18}O values of the stage (3) carbonates can only be interpreted in terms of mixing of this isotopically heavy water introduced from the fault with isotopically lighter water already present in, or migrating through, the bank limestones. This well may provide the necessary explanations, but it does not have, I think, the certainty about it that Boast and his colleagues assert.

Nothing has been said so far in this discussion about the manner in which the residual ore that provided the initial discovery of the Tynagh deposit was formed. Fortunately, the study on this subject by Morrissey and Whitehead (1970) presents a well-reasoned and certainly sound approach to the problem. These authors believe that the stratigraphy of the transported portion of the residual deposit suggests that the deposit commenced to develop when mineralization in the Waulsortian limestone (the top of the primary mineralization) was exposed when the land surface was 6 to 20 m higher than the present surface. This exposure of sulfide mineralization probably took place in the late Tertiary when the climate was quite humid and the air temperature was somewhat higher than it is now. Seventy-five per cent of the residuum by volume is composed of barite, quartz, and clays; the most abundant minerals containing the ore elements are sulfides and carbonates of lead, zinc, and copper. Altogether 38 supergene minerals have been identified, 21 containing essential lead, zinc, copper, or silver. The age of the residuum may be as late as Plio-Pleistocene but may be as old as mid Tertiary. The weathering to produce a sulfide-rich residual body probably was made possible by a near-surface water table, the concentrated character of much of the mineralization, the high purity of the host limestone, and the fractured nature of the ore body. These authors consider that the shape of the residual body was controlled by the differences in rock chemistry from one part of the ore body to another.

Most of the residuum was produced below the water table where acid ground water dissolved an elongate cavity in the limestone in which the western half of the residual ore body later was contained. Somewhat later, solution and subsidence of limestones in the eastern part of the area produced a second large cavity. The two bodies were partially separated by a particularly resistant body of massive ore; the upper part of this body eventually was converted to gossan.

The main depressions were deepened by selective decomposition of the mineralized limestones, and disaggregated primary ore fragments were drawn into them by gravity. Some of these fragments were boulders weighing several tons. At the same time, considerable amounts of mud from the overlying Calp slid down into the depressions, mainly from their north side, and were mixed in various proportions with the primary ore fragments. In places, the Calp mud, sulfide rubble, and clay became interstratified and interfingered. Some of the finer-grained materials that were water transported settled in pools of quiet waters; the materials, after deposition underwent subsidence.

As the main depressions became filled with residuum, this material was differentially compacted and, as the underlying limestone continued to be dissolved, local depressions were produced on the surface of the residuum.

These, in turn, were filled with new and reworked residuum to form structural complexities that the authors say defy brief description. Glacial melt water may have contributed to the physical reworking on the residual material. At least two ice sheets crossed the area, but they seem to have had little effect on the deposit except to remove a little material that was redeposited as an extensive dispersion fan.

The main directions of groundwater movement were downward and northward with the water table apparently being gradually lowered. The downward moving ground water reacted continuously with the residuum and changed its character considerably by decomposing residual sulfides and moving large quantities of metal toward the margins of the residual body. Several species of supergene sulfides were formed during such movement. Local enrichment in copper and cadmium was so produced, and Calp mud bordering the heavily oxidized center of the deposit was enriched in silver, probably with the deposition of acanthite. Supergene galena was widely distributed in the residuum as microscopic-sized replacement coatings on the sphalerite. Numerous fragments of timber were replaced by zinc and cadmium sulfide (but not by lead sulfide).

With the passage of time, the formation of supergene sulfides declined, and the supergene sulfides already present became unstable; most of the surviving galena of this type is corroded and partly replaced by cerussite, and both cerussite and smithsonite are widely present in the dark residual material. Only, however, where the residual material has been most heavily oxidized are the primary sulfides completely oxidized; the most highly oxidized material was classified, for processing purposes, as oxidized ore.

Even after mining began, supergene sulfides of several metals apparently continued to form in the residual ore; pyrite is known to have formed in the 18 months after stripping and before mining because dead newts were found to have been pyritized.

Much secondarily dissolved metal was reprecipiated when it reached the limestone wall rocks. In pure limestone lead, zinc, and copper were reprecipiated as carbonates; in impure limestones, small quantities of such minerals as hemimorphite and dundasite $[PbAl_2(CO_3)_2(OH)_4 \cdot 2H_2O]$ were formed. About 250,000 tons of smithsonite ore (Zn content, + 20%) were developed in this way. Cerussite and malchite were deposited at the eastern end of the deposit; smithsonite formed at the western end. The envelope of metal-enriched wall rock ranges from less than 0.5 m to more than 15 m in width; its development probably continued for at least 2 million years.

As the deposit was raised above the water table, the body of massive primary ore between the two parts of the residual body was changed to gossan. Some gossan also was formed below the water table, particularly in the southeast quarter of the deposit. In such volumes of high ground-water flow, some metal carbonates became unstable and were converted to such minerals as pyromorphite and beudanite $[PbFe^3(ASO_4)(SO_4)(OH)_6]$. Lead, zinc, and copper were greatly leached from the gossans, but gossans formed from primary ore rich in galena and/or tennantite normally are high in silver, certainly as the native metal and perhaps as cerargyrite or argentojarosite; the native silver is alloyed with up to 30 per cent of mercury, the mercury probably having been derived from the tennantite.

The processing of this residuum was complicated by this material having been transported from one volume of the ore body to another after, and while, it was undergoing the various types of alteration already discussed. Various materials in various degrees of alteration occur mixed and interstratified with each other; accumulations of gossan, for example, acted as aquifers, thereby intensifying the subsequent attack on the materials that surrounded them.

If all papers describing geologic phenomena were as clear and well written as this one by Morrissey and Whitehead (1970) that has just been summarized here, the compilation of such a volume as this would be far less difficult than it usually is.

The residual ore would not have been developed, at least not in its present form and mineral character, had the primary deposit not been exposed to the humid and warm climate that was maintained over at least 2 million years. Its age almost certainly is late Tertiary and must be classified as Ground Water - B2.

The mineralogy of the primary ore is such as to suggest that it cannot have been formed under conditions more intense than low-intensity mesothermal; the moderately abundant amounts of tennantite-tetrahedrite associated with galena, plus the wide variety, but small amounts, of sulfosalts, indicate leptothermal conditions of deposition as well. The association of primary silver with both galena and tetrahedrite also confirm the leptothermal classification. The sphalerite and barite being earlier (according to Schultz, 1966) than galena and tetrahedrite-tennantite probably put that portion of the mineralization in the lower mesothermal range. The silver in the galena and its close association with tetrahedrite appear to bar any of the mineralization from the telethermal range. The primary ores, therefore are classified here at Mesothermal to Leptothermal. Obviously, if the ore was not deposited hydrothermally after the lithification of the host rocks, it would have to have been deposited by volcanic exhalations in which event the primary ore would be categorized as having been introduced by the introduction of fluid igneous emanations and water-rich fluids category IIA3.

GREAT BRITAIN

ENGLAND

CORNWALL AND DEVON

Late Paleozoic	Tin, Copper, Tungsten	Hypothermal-1 to Leptothermal

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Notes

The tin deposits of southwestern England are mainly in the county of Cornwall but with some being in western Devon. In general, the deposits are spatially associated with five major granite bodies that outcrop in the mineralized area. The deposits in Devon are grouped in the vicinity of the Dartmoor granite, but they are of minor importance in comparison with those of Cornwall proper. The four Cornish granites are (from northeast to southwest) those of Bodmin Moor, St. Austell Moor, Carnmenellis, and Land's End. It is difficult to estimate if more tin was recovered from lodes in the granite or from those in the surrounding metamorphic rocks (the Killas), but it seems fairly certain that essentially all the copper ores was mined from lodes in the killas.

Mining for metals has been carried out in Cornwall for probably 3000 years with much of the tin recovered (first from placers and later by surface mining of secondary deposits) being transported across France to the Mediterranean area where its use in the making of bronze appears first to have been discovered. Underground mining is thought not to have begun until the time of the Romans, but the evidence for this is not definitive. The first records of such mining activity began to be kept by the end of the 12th century with such operations having continued almost without interruption to the present day. The time of maximum production, of tin, copper, and arsenic, took place in the latter half of the 19th century. K. E. Beer (1978) estimates that the southwest of England has produced about 2.5 million tons of tin and some 2.0 million tons of copper. Tin production has been no more than a few thousand tons a year in the 1970s with the highest annual total coming in 1973 when over 10,000 tons of that metal were recovered. The wide scattering of Cornish miners to the corners of the earth began with the closing of the Cornish copper mines well before the turn of the century and continued as the tin mines became exhausted in the 20th century. Beer believes that the total value of Cornish mineral output to the late 1970s would be about US$35 billion at present day prices. Although China clay is not discussed here, it has added much to the total value of Cornish mineral production, at the rate of about US$100 million a year during the 1970s, and the industry is the largest employer in southwest England.

In 1975, the five mines then operating in Cornwall produced, as metal in concentrates, 3330 tons of tin, 562 tons of copper, 3092 tons of zinc, and 2600 kg of silver. This is a lower figure for tin than is given early in the discussion in the section of the Northwest Europe Volume on the United Kingdom. The two largest producers of tin (over the entire lives of the mines in question) were Dolcoath mine with 93,500 tons of tin (and 355,500 tons of copper) and East Pool and Agar mine 92,750 tons of tin (and 41,000 tons of copper).

It appears probable, if the Cornish district had been operated as a single entity over the duration of its life that: (1) production would bave been far higher than it was or (2) reserves would be much larger than they are now known to be. With more than 400 mines known to have produced tin during the history of the district, and almost all of these separately owned, it would be surprising if huge tonnages of ore had not been left in the ground, a large percentage of which probably never will be recovered.

The Cornish peninsula trends northeast-southwest; along the southeastern side three minor peninsulas are attached to the major one. These are the Lizard, Dodman, and Start, all of which are now thought to be the outcrops of a single overriding plate driven up from the southeast. These remnants of the thrust block do not contain tin and need not be considered in connection with the Cornish tin deposits. The Devonian and Carboniferous sediments that make up the basic framework of the Cornish peninsula are essentially non-carboniferous and normally are lacking in fossils and marker beds. Into these were introduced pre-granite mafic and ultramafic intrusives now metamorphosed to the greenstone facies. Into these sediments and mafic bodies was intruded, in Permo-Carboniferous times, a huge batholith of which the five batholiths mentioned above are outcropping cupolas. With this batholith are associated minor dikes and veins of granites and lamprophyres. Contact alteration has affected both the sediments and greenstones adjacent to the granites.

The Cornubian batholith is estimated to be about 370 km long, but 160 km of this are believed to be the southwestward extension that lies under the sea. Its exposed length (210 km) extends from the Scilly Isles to the southwest to Dartmoor in Devon on the northeast. This granite was not intruded in one huge mass but was polyphase in origin. Its surface is molded into ridges that locally extend upward into cusps or "high spots" that were formed by the lower surface patterns of the invaded rocks. The trend of these granite ridges in the southwest of the region is northeast-southwest, but, in the northeastern part, the trend has changed to approximately east-west. The cusps appear to have been produced where NE-SW and E-W granite-invaded antiforms met and possibly where the antiforms are cut by faults.

Small bodies of aplite and granite related to the main granite largely lack heavy minerals and have no content of tin. Opposed to these bodies, a large number of microgranite (usually porphyrytic) dikes are called "elvans" and were emplaced after the main granite. These bodies usually bear a close spatial association to tin lodes. The elvans are major structures that may be several kilometers in length and may be more than 30 m wide; normally, they strike about parallel to the adjacent granite ridges. The strike of the elvans also is essentially that of the tin-bearing lodes and veins.

The marine and deltaic lower Devonian to upper Carboniferous (Westphalian) slates, shales, sandstones, limestones, and volcanic rocks into which the granites were intruded retained their regional strike with only slight, if any, change. The rocks, regionally metamorphosed to the greenschist facies, were well folded and underwent drastic structural deformation. Three major fold phases have been recognized (Dearman, 1971) in the west and south of Cornwall but only two have been recorded in other parts of Cornubia. Dearman, through the latest age determinations, was able to find a limited southern zone of Bretonic folding (with a Caledonian trend) that was partly masked by a more broadly based east-west folding in Variscan time. The third type of folding was provided by the tecton thermal effects of batholith emplacement (Beer, 1978). Freshney (1965) believes that a slowly rising magma well may have been the cause of not only the flat-lying third-stage folds but also for low angle faults that cut all earlier structures. Stone (1971), however, thinks that the exposed granite bosses appear to have been introduced passively through large-scale stoping of the rocks through which they passed. Beer and his colleagues (1975), however, are of the opinion that Hosking's concept (1964) of ridges and cusps (see above) may be oversimplified in attempting to correlate the roof form of the batholiths with pre-existing folds in the country rock.

During the cooling of the granite magmas in the Cornish batholiths, however, Hosking suggests that the complex pattern of joints already produced in the invaded and metamorphosed Devonian to upper Carboniferous sediments by the earlier multiphase folding was reactivated to cut the granites as well as the older sediments and thereby provided channels of the emplacement of the silicic porphyry dikes (the elvans) and the ore-forming solutions that came in later. These fractures were opened and then filled several times so that the development of both elvans and veins was a multi-phase operation. This repeated dike and vein filling may have been in part controlled by the settlement of fractured blocks between the fracture or fault lines that strike in the general northeast-southwest trend of the Cornish structures.

The major development of ore lodes and veins took place in vein-swarms, all of which are bordered by greisen (that is, quartz-muscovite-or-lepidolite-topaz-tourmaline high-temperature alteration products of the granite). In many places, these greisen-bordered veins contain cassiterite and wolframite in sufficient abundance to be mineable as ore. As would be expected, these greisen-bordered veins are later than the granite intrusions in which they are contained or near which they have been developed in the adjoining killas. Nevertheless, the polyphase character of the granite intrusions makes possible the emplacement of such veins after one stage of granite intrusion and their attack by a later stage of intruded granitic magma. As at Castle-an-Dinas, a tungsten-rich lode that includes quartz and löllingite was cut by a tongue of granite that appears to have assimilated some of the lower portion of the lode, subsequently redepositing a heavy-mineral-rich granite halo in front of, and around, that portion of the original lode beyond the igneous tongue. Another example was found in the Great Wheal Fortune mine where a group of cassiterite-bearing veins was faulted by an elvan; other lodes either post-date or pre-date certain elvans.

Hosking (1970) thinks that ore structures associated with the apices of granite cusps were earlier than the lodes located on the flanks of the granite ridges. He is of the opinion that the ore fluids that filled the cusp-related vein swarms and other structures were developed from the last residuum of the crystallization of the granite magma from which the cusp was formed. On the contrary, the flank-lode ore fluids were produced at depth in the main mass of the granite batholith and migrated upward to deposit in the flank lodes. In both situations, the mineralogical character of the ore bodies formed depended on when, during residuum development, the ore fluid left that portion of the magma and when it arrived in the structure where it deposited its load. Hosking gives examples of cusp deposits that may consist of: simple pegmatites, feldspathic veins with wolframite and arsenopyrite as ore minerals; greisen-bordered veins that do not contain feldspar but may include cassiterite, wolframite, and various sulfides; and quartz veins without greisen borders.

He considers that the processes of ore formation may be complicated by a re-opening of greisen-bordered veins (in cusps?) during the period of flank-bordered vein-formation. As an example he mentions some of the greisen-bordered veins at Cligga Head (a minor granite boss on the north coast of Cornwall) in which sulfides, such as sphalerite, were deposited after a vein re-opening. In the South Crofty mine, he reports that early vein swarms (quartz-feldspar-wolframite-arsenopyrite) were intersected by faults of a later age that were filled by cassiterite-stannite-chalcopyrite-sphalerite mineralization, and some of this mineralization entered into the veins of the earlier dike swarm.

Vein swarms in killas, Hosking believes, are associated in most instances with late-phase granite cusps beneath the swarm area. An example of two-stage deposition in a dike swarm in killas is known from the South Crofty mine where a quartz-K-feldspar-wolframite-arsenopyrite body was later a location of cassiterite-chlorite-fluorite depositon. Hosking suggests that the abundance of chlorite indicates that the ore fluids in this case came from the same source as the chlorite-rich flank lodes. Hosking also points out that some of the killas vein swarms, although they have the typical character of greisen-bordered tin-veins, the vertical extent of their contained tin mineralization is only a 100 m or so; thus, they may well actually be equivalent to flank lodes.

The point should be emphasized, as Hosking does, that tin-bearing lodes and veins normally were re-opened several times with the mineralization deposited after each re-opening being appreciably different from that which preceded and followed it. The average lode is structurally more complex than a greisen-bordered vein swarm and was reopened more often. Further, the relative movement of wall rocks is common in lode ore bodies whereas in greisen-bordered veins such movement is rare.

Many of the mineral species in both types of tin ore bodies are common to both but an appreciably larger number has been reported from the tin lodes. For example, galena is most rare in greisen-bordered veins and antimony minerals have never been found in them. The order of deposition in both swarms and veins is normally much the same; both may exhibit primary zoning, but it is much more apparent in certain lodes than in any vein swarms. Both types of deposits show marked telescoping of their minerals, and wall-rock alteration is much more prominent and complicated in surroundings of lodes than in those of vein swarms.

Some of the minerals probably deposited at moderate (mesothermal) temperatures (low-temperature sphalerite, galena, siderite, and sulfosalts) may be present in the higher portions of tin lodes or may even be in independent veins with strikes parallel to those of the tin lodes with which they are associated. Most of these moderate-temperature lodes, however, are in lodes that strike at about right angles to the main lodes. Some cases are known where lodes of the two types intersect each other so that cassiterite and galena may be mined on the same level in a given mine. Another structure striking at about right angles to the tin lodes is the so-called "cross-courses" that, to a considerable extent at least, were caused by wrench faulting; these cross-courses in many places are filled with late quartz or clay (the latter may have been produced by the alteration of feldspar).

Garnett (1961) says that some of both the mineralized and barren faults that are at right angles to the tin lodes were developed before the tin lodes were formed and that later horizontal movement along these faults produced tension fractures in the ground between any pair of them. These tension fractures may have provided the loci for the initial deposition of the tin lodes.

Further, Hosking points out that the history of faulting in general and lode formation in particular may differ quite definitely from one of the Cornish ore fields to another. Compare, for example, that of the Geevor area (Garnett, 1961) with that of the St. Agnes district (Fern, 1920). All tin fields in Cornwall actually are markedly more complex than what has been said above would indicate; moreover, little is known about the structural geology of most Cornish mines because they were closed before this branch of geology was sufficiently developed.

The age of the tin deposits of Cornwall and Devon has long been thought to be late Paleozoic, and probably most metal ore in the district was emplaced during the time between the end of the emplacement of the Cornubian granites and the end of the Paleozic era. Darnley and others (1965) have recognized three stages in uraninite development (290 m.y., 225 m.y., and 50 m.y.); such uranium mineralization does not necessarily mean that the hypothermal tin-tungsten ore bodies were produced throughout this span of 240 m.y., but it may mean that the galena-sphalerite ores in faults at right angles to the tin lodes may have come in well into the Mesozoic; certainly in the neighboring ores of the Mendip Hills, the ores were introduced in the middle Mesozoic which may indicate an appreciably younger age for the "cross-course" ores in Cornwall than the end of the Paleozoic. Until more information on this matter becomes available, however, I think it best to retain a late Paleozoic age for all of the Cornish non-ferrous ores.

Hosking (1970) divides the economically valuable concentrations of tin (and tungsten) ores into five categories, the second of which is subdivided into four parts. These are: (1) feldspathic ore-bearing so-called "pegmatites"; (2) vein swarms and pipes poor in, or without feldspar and not in close association with lodes these are subdivided into: (a) greisen-bordered veins in granite; (b) greisen-bordered tin-bearing pipes in granite; (c) vein swarms in elvans; (d) vein swarms in killas; (3) early (?), mineralogically unusual lodes; (4) pyrometasomatic deposits, and (5) complex hydrothermal ores, including "normal" lodes, carbonas (irregularly shaped non-vein ore bodies, and replacement pipes and sheets), floors, and vein swarms associated with lodes.

The so-called ore-bearing "pegmatites" in Cornwall almost certainly are not pegmatites; a few actual genuine pegmatites do occur in the Cornish granites. Most of what have been called pegmatites are too low in feldspar or do not have the proper texture to be given this designation. These feldspar-quartz bodies in which wolframite and arsenopyrite normally are present apparently are simply high-temperature hydrothermal deposits that developed in the early stage of either vein-swarm (cusp) mineralization or flanking lode formation; these bodies have been little affected by later re-opening and re-mineralization or were not re-opened at all so that their initial minerals were not appreciably changed from their original character. Hosking gives as an example a lode in the South Crofty mine where a steep-dipping, lenticular body (locally as much as 4 m wide but of limited strike, and probably of dip, length) that originally contained quartz, pink K-feldspar, wolframite and arsenopyrite, the metallic minerals being in the upper part of the lode. Later, this body was re-opened and brecciated with cassiterite being deposited in a crude radiating pattern around the fragments of the original lode material and in fractures in the primary lode minerals. Still later, small amounts of chlorite were deposited and pink replacement rims of scheelite were developed around the wolframite, and some sulfides, including stannite, and siderite were the last minerals to be added. These "pegmatites" are most unimportant in the economic picture of the Cornish tin mining district.

Of the greisen-bordered veins in granite, most are located in the apices of granite cusps (as at St. Michael's Mount, Carn Brea, Cligga, Hemerdon, and Kit Hill). It is possible that a greisen-bordered swarm could develop above a cusp on a granite intrusion introduced into an older granite. Hosking considers that the South Crofty feldspathic swarm may be of this type even though the typical greisen-bordered swarm is essentially barren of feldspar. Where these veins cut into the surrounding killas, they die out over a short distance, and, in the granite itself, they do not extend more than a couple of hundred meters from the granite contact. Most of the veins in a cusp crest are sub-parallel, usually steep-dipping veins that may be up to 30 cm in width but rarely are more than 5 cm in width. Normally, they are bordered on both sides by greisen bands, but in places only one band is present. Again, it is usual for both bands to be of the same width but (locally where the veins are vertical) the bands may be very different in width, and the bands even may be discontinuous. These facts indicate to Hosking that the greisenization took place before the actual veins were developed. The granite between greisen-bordered veins is, in many places, appreciably kaolinized. In a few ore districts, the area may contain more than one set of greisen-bordered veins. The geometry of the veins suggests that they were emplaced largely by replacement and only slightly by the filling of (mechanically generated) open space. The mineral content of a given vein may differ markedly from another not far away at the same level; the veins also may have their mineral content telescoped with locally early and late minerals in the cycle of these veins being within 2 m of each other vertically. The common minerals in these veins are: (early) sericite, quartz, tourmaline, apatite, topaz (less common), and cassiterite and (late) chalcopyrite, sphalerite, and fluorite. Chlorite, hematite, wolframite, arsenopyrite, stannite, and pyrite are minor in amount and are intermediate in age. Obviously variations in these abundances occur as locally wolframite and arsenopyrite are important. Some crude zoning does occur since tourmaline needles are common at depth to be followed at higher elevations by erratically placed aggregates of wolframite, arsenopyrite, and cassiterite. Stannite, chalcopyrite, and sphalerite may be involved in the telescoping. Greisen may include appreciable cassiterite or may be essentially lacking in that mineral.

Hosking considers these veins quite similar to the ore-containing veinlets in porphyry-copper deposits so he says they could be thought of as porphyry tin-tungsten deposits.

In these greisen-bordered deposits, grade always is low; perhaps 3 kg of tin and wolframite together per metric ton. Where tests have been made, it appears that only about 55 per cent of the two metals was recovered.

In the Carnmenellis granite only, tin-bearing pipes have been reported. A major pipe at the East Wheal Lovell mine was worked from about 80 m below the surface about 220 m down. The pipe had a 1.3 cm-wide central fissure filled with Fe-bearing quartz and clay; this was surrounded by greisenized granite composed of quartz, a clay mineral, cassiterite, and minor fluorite, pyrite, chalcopyrite, bornite, chalcocite, and siderite that grade gradually into the surrounding granite. This is an example of a deposit showing by replacement development and telescoping. It is considered possible by Hosking that this pipe was a variant of the greisen-bordered vein-type, but this is not certain as the possible granite cusp has been much changed by erosion.

Certain of the Cornish vein swarms were produced by the splitting of a vein into a series of stringers on entering into the elvan-dike (Wherry and Wheal Vor mines). Numerous swarms that are tin-bearing and not associated with lodes are present in elvans. These are simple in structure and mineralogy and are thought by Hosking to have been formed by ore-forming fluids that leaked away from a ridge zone by easiest upward paths, that is joints in the elvan-dikes. In the Parbola mine (Gwinnear), a tourmaline-bearing elvan (10 to 30+ m wide) contains angular zenoliths of granite; the whole is contained in killas. Where it has been mined it has been found to be cut by near-vertical fractures that strike about at right angles to the elvan, these range in thickness from about 1 cm to 20 cm and usually stop at the elvan-killas contact. Some of these veins contain cassiterite, and, where the veins are narrowest, the host elvan is impregnated with cassiterite. The veins are nearly sub-parallel, but, where they intersect, small bunches, of cassiterite are found. Locally, a few veins enter the killas, and where they do, the country rock is tourmalinized and impregnated with cassiterite.

The vein-swarms in killas normally have structural and mineralogical characteristics much like the veins of the greisen-bordered swarms. These will have tin zones that reach vertically for perhaps 150 to 200 m. They are believed to overlie cusps in the granite beneath. The wallrock next to these veins is highly tourmalinized and contains, in places, much sericite. If the veins intersect lime-rich rocks, axinite is present instead of tourmaline and is associated with other typical skarn minerals. The veins may locally be quite telescoped, but zoning is not well-marked. The mineral content may be simple (cassiterite-tourmaline-quartz) or much more complex (sericite, arsenopyrite, molybdenite, wolframite, stannite, chalcopyrite, and sphalerite also being present). The grade of the ore in these veins is about as low as that in the greisen-bordered swarms. A content of 3 kg per metric ton is average; most such deposits were mined by open pits. These swarms probably are genetically related to flank lodes. Where veins in such swarms cut elvans, branch veins follow both walls of the elvan, and rich concentrations of cassiterite are common in these branch veins. At least one vein swarm in killas is much richer than the average and may have formed during a period of flank-lode formation in the neighborhood. Most cassiterite in this swarm was found in mineralized lenses that follow the bedding planes in the killas. One such lens of this rich variety was 20 by 3 by 80 m (the last being down dip) and ran about 50 kg of tin to the metric ton of ore.

Hosking has set up the category of early, mineralogically unusual lodes to make a place for lodes that are different enough from his other categories as to require a separate designation. One deposit of this type was in granite and consisted of a shear zone a few meters wide. This sheared rock had been replaced on several occasions; green biotite and red garnet were the major minerals. Locally microscopic masses of cassiterite, quartz, fluorite, chalcopyrite, and Fe-rich sphalerite were present. Hosking is uncertain why this peculiar mineral assemblage should have been developed; perhaps greenstone was assimilated into the granite, but he found no evidence of this. Such mineralization also has been noted on the dumps of the Trevoole mine and in drill core from it.

All of the major granite outcrops in Cornwall are bordered by altered greenstones and/or impure limestones. High-temperature tin deposits in these rocks, however, are quite rare and of little economic importance. The mineralogy of what ores there are is much the same as in those in Hosking's categories (2a-d) and (5). A tin lode in the greenstone near a granite contact in the Penlee quarry in the Newlin area is composed of blocks of an earlier cassiterite-quartz-chlorite lode that was broken by earth movements and cemented by quartz. Somewhat differently at the Magdalen mine (Posanooth area) four tin lodes occur that, where they cut greenstone, contain an unusual mineral group - magnetite-hornblende-cassiterite-quartz-chlorite with minor amounts of arsenopyrite and pyrite and traces of bismuthinite and scheelite. The magnetite crystals have cores of hornblende or chlorite and are older than the cassiterite. Magnetite-cassiterite layers alternate with ones that are silicate-rich.

In the Mulberry mine area, calc-flintas (calc-silcate rocks) in bands as much as 20 m thick are interbedded with slightly indurated and folded slates. Fractures in the slates contain tourmaline, mica, cassiterite, and sulfides. In the calcium-rich rocks, however, fissures are filled with the species just listed, but the wallrocks locally are rich in axinite and quartz, cassiterite, and chlorite are present in irregularly dispersed replacement-aggregates. In some of the axinite, spherulites of radial cassiterite have been noted. No mining has been done in this area.

A rather different type of high-temperature tin deposit in calcium-rich rocks is present in a Culm limestone inlier along the northern edge of the Dartmoor granite. The ore is in an anticlinal crest, and the tin is present mainly as malayaite [$CaSnSiO_5$] but also is in andradite and grossularite; no cassiterite or tin-bearing sulfides are known. Although the workings here were on a copper prospect, samples have been collected that run 6 per cent SnO_2; this zone is located within a much larger volume of iron-rich metasomatism. Granite to the south of this occurrence contains mines that are mineralized by cassiterite, specularite, quartz, and tourmaline, and these mines are located in a belt that, in other places, is cut by lodes in which are specularite, quartz, and tourmaline with some local pyrite and siderite.

The fifth of Hosking's categories of Cornish ores contains various complex hydrothermal deposits to which the general term lodes is most applicable.

In southwest England, these lodes are in fault zones that were subjected to important earth movements more than once, and, in many places, several times. The usual lode was started by the formation of a fracture, the dip and strike of which differed from one place on its length to another. During all, or most, of these periods of movement along fractures, ascending ore fluids were passing upward through them. Most of these faults were essentially normal (despite local cases of reverse and tear faulting), thus producing open spaces along the steeper portions of these fractures. The ore fluids traversing these open spaces not only filled these but also were able to replace appreciably the wall rocks enclosing them.

These normal lodes generally have strike lengths of several kilometers and, they may be mineralized down dip for a thousand meters or more. The width of these loads may be as much as 15 m but more usually is 1.5 m. As would be expected, these various dimensions change in dip and strike lengths and in widths, and they may have sharp walls or grade gradually into the wall rock.

These loads never were born in the final size; repeated faulting increased the degree to which wall rocks were broken and room made for further solution passage and mineral emplacement, both in the open spaces created and in the wall rocks to which this brecciation gave them access. Thus, the average load is a combination of brecciated bands, filled fissures, and wall-rock partings with the walls containing various veins and replacement masses. In many instances, the mineralization of the walls made the entire vein-wall rock arrangement profitable to mine. These walls may be either granite or killas or a combination of both with cleavage planes in the slates permitting the access of ore fluids into such rocks and providing centers from which replacement reactions might penetrate out from the cleavage planes. The most productive lodes in killas appear to have been developed where the cleavage joints dipped essentially vertically on the hanging wall. Veins and stringers in such situations were so numerous, though never individually of much width, and so well mineralized that they permitted stoping over widths of 15 m to as much as 30 m.

Another variety of this type of lodes, particularly on the Land's End peninsula, was made up of pipe-like and irregularly shaped bodies of highly altered, tin-bearing granite. These are called carbonas and generally are connected to normal lodes by one or more narrow feeder veins. One vertical carbona, 10 m by 5 m by 1 to 12 m, consisted of quartz, feldspar, tourmaline, and cassiterite and was connected to a normal lode 50 m below the surface by narrow veinlets and pipes. Another carbona (the great St. Ives Consols carbona) lacked clearly defined walls and contained tourmaline, cassiterite, fluorite, pyrite, and chalcopyrite and was localized at the junction of a north-south crosscourse and east-west veins. Its trend was north-south, unlike normal tin-bearing lodes in the immediate neighborhood.

What are called tin floors are much like carbonas and consist of one or more nearly horizontal ore bodies, irregular in outline and from 1 to 4 m in thickness and from 3 to 13 m across. Their mineral content was abundant cassiterite, associated with tourmaline and surrounded by much tourmalinized wall rock. These floors probably were connected by one or more normal lodes.

The lodes have much in common with veins swarms, the major difference being in the greater width of these structures. In many places, lodes change character markedly when they pass from one rock type to another. For example, the lodes may die out entirely on so doing or their structure, width, and mineral content may increase drastically, and their dips and strikes may be appreciably refracted.

The lodes may show primary zoning and also may be considerably telescoped, although Hosking says that this latter phenomenon has not been given sufficient prominence in the literature.

Although the paragenesis of the Cornish tin deposits has been much studied, the relationships of the various minerals are complicated by such possible processes as ore-fluid resurgence, two ore fluids from different sources, later ore fluids reaching areas in which earlier ore fluids had deposited their loads, and possible remobilization, zoning, and telescoping that results based on the examination of a single lode or even deposit may be at variance with the pattern established for detailed study of the entire district. No one ore geologist has studied more and seen more of the geology of the Cornish ores than Hosking. It, therefore, is most reasonable to follow his presentation of the paragenesis in a discussion such as this designed to introduce the reader to the area rather than to present an exhaustive study of it.

The earliest minerals (ore and gangue) deposited in the district (and only in scattered parts of it) are the feldspar-mica bodies known as "pegmatites" (see above). In addition, these "pegmatite" bodies contain arsenopyrite and wolframite as their major metallic minerals, plus some cassiterite and molybdenite. These pegmatites are followed by the early veins swarms (see above) that are in veins in, and above, granite cusps; they contain principally arsenopyrite, wolframite, and cassiterite, plus less abundant chalcopyrite-sphalerite, stannite, and molybdenite. Unusual alteration minerals associated with these swarms in places are garnet, beryl and euclase [$BeAlSiO_4(OH)$]. With both these stages, tourmaline is important and continues to be deposited through the early stages of the hypothermal lodes.

These early veins swarms are followed by the hypothermal lodes that generally are arranged in parallel to the granite ridges (that trend NE-SW in the west and E-W in the east) and their associated dikes. The major minerals, in general order of introduction are: cassiterite (plus wood-tin, see below), molybdenite (?), wolframite (plus minor scheelite), arsenopyrite (plus löllingite), sphalerite that is iron-rich and contains ex-solution bodies (mainly chalcopyrite), stannite [Cu_2FeSnS_4] and chalcopyrite. After the deposition of these hydrothermal lodes, an hiatus developed in ore-solution flow or in deposition from ore solutions for the mesothermal and leptothermal lodes that follow are general in structures at right angles to the granite ridges (although some small fraction of them may strike parallel to those ridges). In places, the ore fluids responsible for these later lodes have had access to some of the hypothermal bodies and have added these mesothermal and leptothermal minerals to the older lodes. The order of deposition of ore minerals in these later lodes is: minor bismuth and bismuthinite (?), pitchblende, niccolite, smallite, and cobalite, then galena (silver-rich), argentite, and iron-poor sphalerite (no ex-solution bodies), pyrite (plus some marcasite), siderite, tetrahedrite, bournonite, pyrargyrite (?), stibnite, jamesonite, and hematite.

During the deposition of the hypothermal and mesothermal-leptothermal lodes, the gangue minerals were the last of the tourmaline [plus minor danalite, $Fe_4^2Be_3\ (SiO_4)_3S$ and phenakite, Be_2SiO_4], then, in order of beginning of formation, chlorite, hematite, fluorite, barite and chalcedony, dolomite, and calcite. Chalcedony and minor chlorite are the last of the gangue minerals to have been deposited. The final metallic mineral was pyrite. In addition to copper, tin, and tungsten, metals of economic importance are (and were) arsenic, bismuth, uranium, nickel, cobalt, silver, lead, zinc, antimony, and iron. Obviously, not all of these metals are recovered from any one lode, but, in the five mines operating in Cornwall in 1975, 3300 tons of tin, 562 tons of copper, 392 tons of zinc, and 2600 kg of silver were produced, presumably at a profit.

Wall-rock alteration in the Cornish ore bodies; there are several types. Silicification may be early, intermediate, or late and may be present with any ore type. Kaolinization may be early (although this is not certain). Sericitization (the term Hosking uses for the mica in greisen) tourmalinization, chloritization, and hematization normally appear in that order. Sericitization is the dominant form of wall-rock alteration in the vicinity of veins in the granite cusps; the sericite so formed may be converted to various extents to tourmaline; gilbertite (a silky form of muscovite) may take the place of sericite in some greisen zones. In certain areas (Cligga, for example), Hosking is certain that kaolin is early.

Simple vein systems in the killas normally have intensely tourmalized wall-rocks, although various forms of mica may be present also as vein selvages. Such wall rocks are hematitized or chloritized in only a few places. In granites and killas immediately adjacent to the major lodes, however, may be, at any given horizon, changed to one or more of the types of wall-rock alteration just mentioned; where more than one type occurs, the youngest type has its major development farthest from the vein and so on back to the vein. Greisen is a minor feature of these major lode deposits. Tourmalinization usually is most important near those tin lodes that have been little disturbed after the deposition of the cassiterite that they contain. The deeper tin lodes normally are within tourmalinized walls even though they later were disturbed and later alteration developed higher up. Chloritization and hematitization normally follow the deposition of cassiterite. All of the wall-rock alteration minerals also are present as gangue minerals in the ore.

Not all Cornish mines, by any means, show vertical zoning, but it is so common as definitely to be an established fact. Many mines have been mined for tin at depth after having been copper mines above, and many tin lodes or veins have been mined for copper outward along strike from the center of the tin mineralization. It seems certain, however, that the mineralization of the average lode is so complex through repeated periods of mineralization that no simple zoning pattern can reasonably be established. Some of these later minerals that complicate any attempt to elucidate a zoning pattern also make the tin ore difficult to beneficiate. The close association of tin and copper ore minerals in the Geevor mine is due to the later introduction of copper and zinc minerals into a volume once occupied by tin ore only. Certainly, the Geevor silver was not present in the tin ore but was added with the copper and zinc sulfides. Thus, it would appear that, at the time that these copper and zinc sulfides were being deposited at Geevor, the temperatures of the ore fluids were lower than they were when the tin ore was emplaced. Hosking suggests that the better tin ores are located in rock volumes where the flow of the ore fluids was retarded, giving more time for the reactions needed to precipitate the tin minerals to take place; this situation also can apply, of course, to the minerals of other ore metals. The development of rich ores at fracture intersections also probably was the results of solution flow being slowed by the fault detritus, aided by the development of smaller fragments with fresh surfaces that could be readily attacked by the ore fluids. Normally, stannite is found as a replacement of cassiterite, with the tin content of the stannite very possibly having been obtained entirely from the cassiterite and not from the ore fluid. Hosking suggests that some of the abnormal paragenetic relationships in the lodes may have been due to remobilization, but this is certainly not established beyond doubt.

The erratic distribution of cassiterite in the tin lodes so that the conditions necessary for its precipitation are quite limited and are attained only sporadically in the average tin lode. Further, what may have been a tin lode that could have readily been mined at a profit underwent so much brecciation and dilution by later, essentially valueless, minerals that, when found, it could not be mined or was mineably only at a less profitable rate than would initially have been possible.

One of the interesting facets of Cornish tin lode mineralogy is the presence of wood-tin. Hosking quotes Newhouse and Berger's definition of wood-tin: "wood-tin displays, in general, a colloform appearance, usually botryoidal or globular, but occasionally occurs as gently undulating slabs apparently deposited on flat vein-walls. Microscopically, the wood-tin structure is found to owe its banding to variability in the color of the cassiterite, which varies from deep, almost opaque, red to practically colorless. This color banding is concentric with the external surfaces of the nodules and the individual cassiterite crystals are elongated transverse, and frequently normal to, this structure". Wood-tin is found in lodes in granite and in the killas but is restricted to a few areas in the Cornish district (in the St. Agnes area for example) and is less likely to be present in a given lode than normal cassiterite. At any given horizon, however, it can be found in considerable abundance. In the St. Agnes area, the wood-tin always is present in lodes developed by faulting in the killas. In the development of such lodes, the first minerals to be introduced were acicular tourmaline needles and quartz. These were followed by massive twined cassiterite. All these minerals were emplaced by open-space deposition or replacement of the killas. The products of this first stage of mineralization then were brecciated and, after this faulting, the wood-tin was deposited along with acicular cassiterite. Next came mild fracturing and then the deposition of pale cassiterite, partly by replacement of wood-tin and partly by filling open space. Further re-opening was followed by the deposition of curious forms of pale chlorite plates in bands in a quartz matrix. Still additional movement permitted more open-space quartz deposition and quartz replacement of wood and acicular tin. More rock movement was then followed by sulfide deposition, including high-temperature chalcopyrite and sphalerite that contain exsolution bodies, the one of the other; to a minor extent the sulfides may have replaced wood-tin, but no stannite was developed.

Wood-tin definitely is made up of a large number of bands of different colors. In places, spherical masses of silica were deposited with the wood-tin; the silica later becomes a fine mosaic of quartz. Whatever, the character of wood-tin at the time of its deposition, that now seen is composed of radiating needles of cassiterite that locally can be traced across two or more color bands.

At any event, wood-tin is not a product of secondary processes but is developed as part of the formation of the primary ore. It does bear a complex relation with quartz but certainly is a product of the same hypothermal solutions that usually produce normal cassiterite; its intimate relation with quartz is what distinguishes it from normal cassiterite. The fact that wood-tin in Cornwall, locally at least, is followed by high-temperature chalcopyrite and sphalerite definitely shows it to be a mineral formed as a member of a normal hypothermal sequence.

Stannite is erratically distributed in the Cornish lodes and veins but never in sufficient quantity to make its recovery worth attempting. Stannite normally is introduced in association with chalcopyrite and sphalerite (these sulfides being of the high-temperature variety); some to much of the stannite was developed by the replacement of cassiterite, but enough of it is not associated with cassiterite directly to make certain that the ore fluids carried tin in a different form than that from which cassiterite was precipitated.

The classification of the Cornish ores does not seem to present any difficulty. The higher-temperature minerals certainly were deposited in the hypothermal range. The minerals in the cross-courses, however, nearly as certainly were deposited in the mesothermal to leptothermal range with the pitchblende, bismuth, and the arsenides being at the high end of the mesothermal range or even at the very low end of the hypothermal but with the remaining minerals meso- to leptothermal.

BRITISH LEAD-ZINC

Middle Paleozoic to Middle Mesozoic	Lead, Zinc, Fluorite Barite, Silver	Mesothermal to Telethermal

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Robinson, B. W., 1980, Isotopic evidence on the origin of sulfur in Mississippi-Valley-type deposits, particularly in the British Isles: 5th IAGOD Symp. Pr., v. 1, p. 487-493 (Snow Bird)

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____ 1954-1955, Concealed lead-zinc fields in England: Inst. Min. and Met. Tr., v. 64, p. 477-485 (general); disc., p. 617-636, 655-656; 1955-1956, vo. 65, p. 33-36, 110-111, 386-390

Schnellmann, G. A. and Scott, B., 1970, Lead-zinc mining areas of Great Britain, in Jones, M. J., Editor, Mining and Petroleum Geology: Ninth Commonwealth Min. and Met. Cong. 1969, Pr., v. 2, p. 325-356; (disc.) p. 357-365

Scott, B., 1976, Zinc and lead mineralization along the margins of the Caledonian orogen: Inst. Min. and Met. Tr., v. 85, Sec. B, p. B200-B204

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Notes

Of the eight areas of lead-zinc mineralization considered in this volume, six are in England, one in Scotland, and one in Wales. The six in England are, from north to south, (1) the Northern Pennines in which the ore is contained in two blocks, the more southern of which is the Askrigg block in the western part of the North Riding of Yorkshire (Askrigg town, 54°19'N, 2°04'E) and the more northern of which is the Alston block located around the junction of the counties of Durham, Northumberland, Cumberland, and Westmoreland; Alston town (54°49'N, 2°26'E) is in Cumberland; (2) the Lake District of northwestern England centers between Keswick (54°37'N, 3°08'W) in Cumberland and Coniston (54°22'N, 3°05'W) in Lancashire; some mining also was done in Westmoreland, where the westward extension of that county lies between Lancashire and Cumberland; (3) the North Derbyshire district that occupies a belt some 40 kilometers long in the northern part of that county and at the southern end of the Pennine Chain in the Peak district where the Carboniferous limestone has been elevated in a great dome; this limestone massif contains a large number of now abandoned lead mines, the entries to some of which lie outside the crop line of the limestone but the workings of which are within that rock; (4) the West Shropshire district that mainly is in the western part of the county of that name but also overlaps to the south into what used to be Montgomeryshire in Wales - this district has two centers, one within a 10 kilometer radius of the town of Shelve that, in turn is about four kilometers south-southeast of the town of Worthen (52°38'N, 3°00'W) and the other within an area of about 2.5 square kilometers southeast of Breidden Hill that is, in turn, about 18 kilometers west of Shrewsbury (52°43'N, 2°45'W); (5) the Mendip Hills in northeastern Somerset on the southwestern side of which are the town of Cheddar (51°17'N, 2°46'W) and Wells (51°13'N, 2°39'W), famed for their cheese and their cathedral, respectively; (6) Southwestern England that includes mines in five areas; the first four listed here are in Cornwall and the fifth is in Devon. These are: (a) The Perranzabuloe-Newlyn East about 10 kilometers northwest of Truro; this area is about 16 kilometers northeast-southwest by five kilometers northwest-southeast; (b) the Perran Iron Lode about 11 kilometers north-northwest of Truro; this area is about five kilometers long (northeast-southwest) and of about the same width as the Perranzabuloe area; (c) the Menheniot that extends about 11 kilometers from south of Liskeard to due north of Menheniot; the area is about five kilometers wide (northwest-southeast) and nearly 10 kilometers long (northwest-southwest); (d) the Tamar Valley on the border between Cornwall and Devon, slightly west of a line between Tavistock and Plymouth and about equally distant from the two towns; its long dimension runs about north-south for about 6.5 kilometers and is 1.5 kilometers wide; (e) the Teign Valley, an area that extends about north-northwest-south-southeast several kilometers - it is somewhat more than three kilometers wide; (7) Leadhills (55°25'N, 3°47'W) in Lanarkshire in Scotland; the field extends across the border into Dumfrieshire as far as the town of Wanlockhead; and (8) the Mid-Wales ore field centers around the Plynlimon Dome in Wales; the mid-point of which structure is located about 25 kilometers east of the port of Aberstwyth (52°25'N, 4°05 W); the longest of the lodes appears to be the Castell that extends westward from the Castell to the Rheidol Valley; it has a total length of about 10 kilometers.

The ages of the various lead-zinc deposits in Britain range from Middle Paleozoic to Middle Mesozoic in the following order:

Middle Paleozoic	Late Paleozoic	Early Mesozoic	Middle Mesozoic
West Shropshire	Southwestern England	North Derbyshire	Mendip Hills
Lake District	Leadhills	Northern Pennines	
Mid-Wales			

This age classification, if correct, indicates that lead-zinc deposition continued over a long portion of geologic time, some 450 million years. Obviously, all of this widely separated series of deposits of such diverse ages cannot have obtained their metal-sulfur content from the same source and possibly not by the same general process. This problem will be examined in the discussions in this volume on the various deposits included.

In four of the eight deposits duscussed in this volume, the silver recovered from the various deposits, mainly from the galena, is much larger than is present in the standard examples of the Mississippi-Valley-type ores. That such standard MV-type ores were developed in Britain, the deposits in the Pennines, West Shropshire, and the Mendips testify, but even most of these produced small amounts of silver from the galena they contain whereas such deposits as those in the Mississippi-Valley or Poland or the Eastern Alps contain so little as to make it almost always not worth recovering. Why this unusually large amount of silver should be present in half of the British lead-zinc ores does not appear to have received the study it merits. Such considerable amounts of silver suggest that the British ores were deposited at somewhat higher temperatures than the lead-zinc deposits of the normal Mississippi-Valley-type. The acceptability of silver sulfide in its argentite form in the similarly structured galena and the great difference between the structure of acanthite and galena suggests that a considerable portion of the galena of the British lead-zinc deposits was precipitated above the temperature of argentite inversion - an event that takes place at 173°C. If this reasoning is correct, a major proportion of British galena must have formed in the leptothermal rather than in the telethermal range. The exception to this statement is in the galena of the deposits formed along the length of the Pennine Chain and in West Shropshire and the Mendips where silver is so low, but not completely lacking, that most of the galena probably was precipitated below the Ag_2S inversion temperature. On the contrary, however, the silver recovered from Leadhills, the Lake District, Mid-Wales, and Southwest England (northern Cornwall and west Devon) is so much larger than telethermal conditions would have permitted the galena to receive in solid solution that these deposits must be categorized, at least in considerable part, as leptothermal. Some genetic value must lie in the fact that the four silver-bearing British lead-zinc deposits lie to the west of the four in which silver is much less abundant or essentially lacking. I have never seen this point discussed.

The age of the four deposits that contain abundant silver is Paleozoic, two middle Paleozoic and the other two late. Of the four deposits largely lacking in silver, three (the two Pennine districts and the Mendips) are Mesozoic but West Shropshire is middle Paleozoic. All of these ores are classed in this volume as Telethermal, although the West Shropshire ores may be, in the lower reaches of the veins, leptothermal even though no silver worth mentioning is present in the galena.

As Schnellmann and Scott point out (1970), production of lead in Britain has amounted to between 6,100,000 and 7,000,000 tons of lead from 8,710,000 tons of concentrates, whereas that of zinc probably has been between 500,000 and 700,000 tons of metal from some 1,250,000 tons of concentrates. The largest portion of this production was obtained in the 18th and 19th centuries, although most of the zinc must have been recovered in the last two-thirds of the 19th because methods of recovering it and its usefulness were not understood until about that time. The last major lead-zinc mine is reported to have been closed in 1961. Obviously, not all lead-zinc mines closed in 1961 for, in the six years from 1960 through 1975, the production of lead in Britain averaged about 6,500 tons and that of zinc some 7800 tons, both of metal. The zinc, however, is stated to have been derived from a mixed copper-zinc concentrate which almost certainly was not recovered from any British lead-zinc deposit, but where else it came from is not immediately clear.

Of lead production through 1961 these same authors believe that 75 per cent came from the Northern Pennines and 52 per cent of the zinc. Rather surprisingly, 12 per cent (some 1,900,000 tons of lead) came from North Wales, a district not discussed in this volume because of the paucity and considerable age of the literature available for study. The North Derbyshire field is reported to have produced eight per cent of the total through 1961. Nearly all the operations were small; only six mines are known to have recovered over 200,000 tons of concentrates (probably about 500,000 tons of lead metal). In most fields, the mining was dominated by one operation, for example, 70 per cent of the concentrates from North Derbyshire came from one mine (Mill Close), but Dunham thinks this percentage unrealistically large. They (Schnellmann and Scott) also believe that 90 per cent of concentrates from the Lake district was from a single mine (Greensides). Two sub-areas in North Wales provided most of the production from that district; the mines of the Minera district and of the Halkyn district each appear to have accounted for more than 200,000 tons of lead concentrates. The mines of the Minera area also recovered 150,000 tons or more of zinc concentrates. The Westminster lode in the southern part of the Halkyn-Minera field probably produced about 100,000 tons of lead concentrates. The only reference I have been able to find that could be expected to tell anything about the North Wales is that by Schnellmann on the Halkyn district that covers 40 pages (585-624) in the Transactions of the Institute of Mining and Metallurgy for 1938-1939 and that is not enough on which to base a discussion of the scope I expect to put into this volume.

In addition to North Wales, there are two other districts noted by Schnellmann and Scott, Coombe Martin in North Devon and the rather large, but exhausted deposits in the Isle of Man. Both of these, as is true of Halkyn-Minerva, are high in silver and fit into the pattern already described which shows silver-rich deposits consistently to lie to the west of those that are low to poor in silver. This information does not go very far in helping explain why this fact is so. It is interesting, however, to note that a line running roughly northwest-southeast across the Polish zinc-lead district of Upper Silesia, divides galena with an appreciable trace content of silver to the southwest from lead sulfide to the northeast from which silver is almost lacking (Ridge and Smolarska, 1972). Factors bearing on the genesis of the Silesian-Cracovian lead-zinc deposits in southern Poland: 24th Int. Geol. Con., Sec. 6, p. 216-229).

NORTHERN PENNINES, YORKSHIRE, NORTHUMBERLAND, DURHAM, WESTMORELAND

Early Mesozoic
Fluorite, Barite, Lead, Zinc
Telethermal

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Notes

As is true of most British deposits of lead and zinc, the Northern Pennine mines were worked by the Romans, and mining has continued, with numerous interruptions and changes in the mineral materials sought, until the present time. Within the last 50 years (Ineson, 1976) the emphasis has shifted from the two metals to the nonmetallics - fluorite and barite. The Northern Pennine field is divided into two distinct structural blocks, the northern, the Alston block, that is bounded by faults on three sides, Stublick to the north, Pennine to the west, and the Lunedale-Stainmore fault system and the Stainmore trough to the south. To the east, the sedimentary rocks pass beneath the Coal Measures of the Durham coal field; the Coal Measures in turn are covered farther to the east by Permian rocks.

The southern block of the Northern Pennine field is the Askrigg and it is confined to the county of Yorkshire, whereas the Alston ores occur in four counties, Northumberland, Durham, Westmorland, and Yorkshire. The Askrigg block also is surrounded by faults and fold systems, the Lunedale-Stainmore trough to the north, the Dent fault to the west, and the Craven fault system to the south. The Craven system marks the line of change from block conditions to the north and basinal to the south. As is true of the Alston block, the ore-bearing Carboniferous sediments dip under Permo-Triassic and Jurassic sediments, successively farther east.

The two blocks can most satisfactorily be considered as separate entities and are so discussed here, with the Alston block comming first.

The mineral deposits in the Alston block are in rocks that range from Ordovician to Silurian to the Permian Magnesian limestone, but only the three limestone groups of Visean age contain workable deposits. The basement, the Skiddaw slates, made up of slates and flags, is of unknown thickness. It is unconformably intruded by the Devonian Weardale granite; unconformably over the granite is the Carboniferous Basement group, +300 meters of sandstones and conglomerates, followed conformably by the three divisions (upper, middle, and lower) or the Carboniferous Limestone series that has a total thickness of 540 meters and that consists of a rhythmic sequence of limestone, calcareous shale, non-marine shale, sandy shale, sandstone, ganister, and coal. This sequence is present in all three divisions but is best expressed in the middle Limestone. The combination of alternations of brittle limestones, compact limestones, and soft shales has (Ineson, 1976) influenced the form and disposition of the ore shoots. These Carboniferous beds normally are flat-lying, and they are involved in only one major tectonic event, the Burtreeford disturbance. The Burtreeford is a sinuous monocline that faces east; it has an amplitude of 160 meters, and cuts the ore field into two almost equal parts.

Above the upper Limestone series beds is the Carboniferous Millstone Grit, sandy deltaic beds that contain thin limestone layers. The Millstone grit was followed by the Coal Measures (also Carboniferous) that are divided into three sections, the lower of thin, sandy coals, the middle of rhythmic alternations of limestones, shales, sandstones, and coals, and the upper of grits, sandstones, ganisters, coals, limestones, and shales. The Millstone grit is 370 meters thick, the Coal Measures 750 meters.

The Carboniferous beds were intruded by a late Carboniferous - early Permian series of interconnected phacoliths (or sills) with associated dikes; the intrusive is a quartz dolerite and is known as the Great Whin Sill. The Whin sill is present at numerous horizons, but its greatest and deepest expression is in Viséan rocks in the Teesdale area (toward the south-central part of the Alston ore district). The quartz dolerite is in Westfalian rocks in the northwest of the Alston block. The sill thins and cuts upward through the formations as it is followed toward the edges of the ore district. The intrusion was introduced in the late Carboniferous or early Permian and has metamorphosed the neighboring rocks for as much as 18 meters from the intrusive. The soft shales of the wall rocks have been silicified, converting the shales into what appear to be competent sandstones; in places this silicified shale is a host to ore. At the same time as the sill was intruded, a tectonic event produced a domal structure over the ore district and was accompanied by faulting and folding.

In the middle Permian, the district was cut down to a low desert plain, on this sands and breccias were deposited. The deposition of these materials was followed by the transgression of the Zechstein Sea and the depositon of upper Permian Marl slate and Basal Yellow sandstones, after which were deposited the upper Marl and Magnesian limestone, and finally the Keuper marl and the Bunter Sandstone. The total thickness of this upper Permian sequence is 1025 meters. The Triassic proper consists of thin evaporites, red sandstones, and mudstones. Rocks of Jurassic and Cretaceous age may once have covered all or most of the Alston block ore district, but, if present, they were removed by Tertiary uplift and erosion. This Tertiary uplift resulted in a regional dip to the east and was accompanied by the introduction of an east-southeast series of tholeiitic dikes.

Masses of Caledonian granite are known in the basements of both the Alston and the Askrigg blocks; the granite obviously is older than the ore, epigenetic traces of which are found in the rocks of the Coal Measures; most of the ore, however, is in the Visean limestones.

In the Alston block, the ore was deposited in near-vertical veins; these are filled fault fissures with which minor shear folds were associated. The main fissure directions were N65°E, N80°W, and N10°-25°W; these probably were formed during the doming in late Carboniferous through early Permian time; the intrusion of the Whin Sill may have been responsible, in part at any event, for the doming and fracturing. The principal ore bodies are in faults that have one to 1.5 meters of downthrow; the widespread and persistent joints of this trend (N65°E) are not, however, mineralized. In general, the veins in which galena is dominant are those with N65°E trends; the N10°-25°W fractures seldom are mineralized even though they probably formed at the same time as the other two sets. Fluorite is the main mineral (but with byproduct galena) in the N80°W fractures, which now are the principal sources of both fluorite and galena.

These fractures are essentially vertical in limestone and sandstone but may be as low as 45° in soft shales. Open space is developed (Dunham, 1976) only where the fractures are steep; they tend to close up in shales. Thus, the ore shoots are usually thin in longitudinal sections and have much greater horizontal than vertical dimensions. In the major deposits, several ore shoots may occur one above the other and are separated from each other by barren ground. In the N80°W (the fluorite) veins, stratigraphic control is less pronounced, and the ore shoots formed in east-west openings of limited extent, openings that resulted from left-hand strike slip movements. Where veins pass from limestone into shale, the flow of the ore fluids appears to have been slowed down, so that these fluids, in considerable part at least, could work their way outward through minor fractures to form replacement ore bodies (flats, wings, or mantos) that in many places have been economically profitable to work. Such replacement deposits may extend up to 12 meters into the surrounding limestone. Such replacement mineralization includes the production of an ankerite-rich material that also contains various amounts of siderite and silica. Some of the replacement has been of the volume-for-volume type, in which bedding planes, joints, and fossils have been preserved, but, in other parts of various veins, solution (or karst) cavities have been formed; these may be lined with secondary minerals. In some instances, the cavities developed have grown so large, through thinning of beds and collapse of the beams so created, that the lower portions of the caverns were partly filled with rubble breccia. Considerable mineralization appears to have taken place by filling voids in the breccia and by replacement of breccia fragments. Ineson (1976) suggests that the paragenetic sequence seems to have been ankerite, siderite, and silica (deposited prior to the introduction of sulfides) then the sulfides, fluorite, and barite, and finally calcite.

The different strength characteristics of the host rocks produced by the cyclic system of deposition caused them to yield differently when affected by tectonic forces, so that the veins normally are refracted as they pass from one rock type to another. The greater the depth at which a particular vein is observed, however, the less this refraction appears to be. This may account for the greater concentration of ore minerals in the rocks of the relatively near-surface Carboniferous limestones. About 75 per cent of the ore taken from the rocks of the Alston block has come from the Limestone series.

Although galena was the principal mineral mined from the time of the Romans until the beginning of the 20th Century, it now is only a byproduct of fluorite mining. Sphalerite was present in mineable quantities only in a few areas in the block (e.g., Nenthead-Alston and Greenhurth). Barite was once mined throughout the block, but reserves of this mineral now appear to be nearly exhausted. In 1976, barite was mined only in the Closehouse mine. In the central part of the ore area of the Alston block, fluorite is present as a matrix for the sulfides and occurs in purple, mauve, yellow, and green colors. Minor sulfide minerals, which are present in small and localized amounts, include chalcopyrite, pyrite, marcasite, and pyrrhotite.

As far as 100 meters beneath the surface, the ores have been appreciably affected by surface processes. The major supergene secondary alteration is to such minerals as cerussite, smithsonite, hydrozincite $[Zn_5(CO_3)_2(OH)_6]$, malachite, and limonite.

The veins of the Alston block contain from eight to 25 per cent SiO_2; this condition is not duplicated in the Askrigg block, where silica is only present in the veins in minor amounts, if at all. Considerable quantities of ankerite and siderite are produced in the Alston block with ankerite being most abundant in places in which replacement of host limestone has occurred. The siderite is more common in Weardale where it is mined as an ore of iron. Dunham (1948) suggested that the materials in the ore veins had been leached from the Whin sill; the results of recent exploration, however, have shown ankerite-siderite-quartz to be present in the lowest limestone well beneath the Whin sill and suggest a deeper source than the sill for the three minerals just mentioned and probably for the sulfides and other non-metallics as well. The only possible way to avoid this conclusion is to assume that migration of epigenetic materials was both upward and downward from the sill, probably not very likely in late Paleozoic - early Mesozoic time.

An unusual feature in the Whin sill was an economically viable replacement ore body of witherite $(BaCO_3)$ that is now exhausted.

Dunham (1948) considered that the field could be divided into concentric zones with the fluorite-dominant veins forming the central zone. The intermediate zone veins contain fluorite but no barite, and the fluorite is not nearly as abundant as in the central zone. In the outer zone, barite is the major non-metallic and fluorite is not present. Quartz is ubiquitous in the Alston ore field but is most abundant in the central zone. Sphalerite and chalcopyrite appear to have been confined to the central zone, but their location there is spotty. Both evidence from drill holes and from deeply eroded veins suggests that the Alston field may be zoned vertically as well as laterally with quartz-chalcopyrite-pyrrhotite ores at depth grading upward into, first, galena-sphalerite, then fluorite-barite, and finally barite. Galena is present all through the fluorite zone, but it is most common near the transition from fluorite to barite. Sphalerite, where it exists, is in the same border zone. In the outer (barite) zone, sulfides gradually die out.

Sawkins (1966) demonstrated that the zoning is not only one of minerals but also of temperatures of formation. His work showed that the temperature of mineral formation in the central zone was about 220°C and that it dropped steadily to 100° at the outer edges of that zone. The barium zone may have formed at even lower temperatures, although Sawkins' evidence on this point is not fully substantiated.

The ore veins appear to have been re-opened at least more than once, producing recementation of the ores and crustification banding on the freshly developed walls. Earth movements that occurred after the mineralization (or after one phase of it) produced slickensides where they cut through the ores.

The presence of granite beneath an ore field was suggested by geophysical work and demonstrated by the Rookhope drill hole. As the granite is covered by Carboniferous beds that rest unconformably on it, it must be older than the Carboniferous sediments. The presence of granite in the center of the zoned area of mineralization, however, makes it seem possible that the granite provided channelways for ore fluids to move upward toward the limestones from an igneous source below the known granite.

Wall-rock alteration has affected both the sediments and the Whin sill. Under the microscope, it can be seen that this alteration results in the progressive loss of original and diagenetic textures. So far as alteration minerals are concerned, the limestone was changed to ankerite-rich rocks with varied quantities of siderite and silica. Volume-for-volume replacement, although locally occurring, (particularly in cavern-filling fragments) was not universal, so the usual thickness of a limestone bed grows less as the vein is approached. Metasomatism causes a complete loss of calcium, an increase in silica and iron sulfides with alumina being incorporated in new sericite. In the wall rocks, K_2O, Na_2O, TiO_2, Rb, F., Pb, Mn, Zr, Zn, Cu, Ni, and Fe all show increases over background as the veins are approached, whereas only Sr decreases the nearer to the vein the sample is taken.

So far as the sandstones are concerned, the main alteration effect is recrystallization of silica, plus minor replacement by galena and pyrite. Zr, Sr, and Rb decrease away from the veins. In shale, the main alteration is silicification.

The alteration of the quartz dolerite of the Whin sill is normal, the main effect is the formation of a white clay-carbonate rock containing kaolinite, illite, anastase, ankerite, siderite, muscovite, calcite, and pyrite. CaO, MgO, Na_2O,, and Fe have been removed and CO_2, K_2O, and H_2O have been added. The alteration usually is no more than one meter wide around each vein in sill rock.

Geologically, the main evidences for the age of the ore are that the veins were developed after the Whin sill was intruded and that they were after the deposition and lithification of the upper Permian Magnesian limestone. These two relationships appear to put the arrival of the mineralization in the ore area of the Alston block just before or just after the end of the Permian.

Geochronological evidence is very confusing, dates between 284 ± 40 million years and 170 million years have been put forward by reputable experimenters. Dunham, and others, (1968) think that they can see, through ^{39}Ar and ^{40}Ar dating methods, that three major geochemical events affected the area at about 284, 230, and 170 million years. The best guess I can make from the evidence (remembering the apparently early Mesozoic age of the North Derbyshire ores) is that the Alston block ores came in about the same time, but certainly the final verdict is not in.

Almost as many genetic models have been put forward for the Northern Pennine ores as there are authors who have studied the problem. An early worker thought the ore fluids were surface waters percolating downward. Later, it was suggested that the solutions came from great depths, almost certainly meaning a magmatic source for them. Still later, Dunham (1948) thought that a proportion of the ore and gangue minerals came from the wall rocks, but later Dunham statistically demonstrated that such an hypothesis could not account for more than a fraction of the lead that had been produced from the whole Northern Pennine field (both blocks). Again, Dunham (1934) suggested that the ore fluids came from hidden granite at depth, these fluids finding their way to the Carboniferous limestones through fracture channels in Weardale granite cupolas beneath the district.

Sawkins' work on fluid inclusions from the Northern Pennine minerals discovered their high saline content which he felt he must incorporate into any satisfactory hypothesis for the genesis of the ores. He envisioned the base metals and the fluorine coming upward in juvenile hydrothermal fluids through the Weardale granite (possibly from the mantle itself). Contrarily, the barium was thought to have come from connate waters that originated in the adjacent sedimentary beds and migrated up dip to arrive, at essentially the same time as the ore fluids from deep in the crust. The barite-bearing connate waters, however, would deposit their loads in an oval-shaped area outside the area under the influence of the fluorine-bearing juvenile (?) waters, thus separating, as is largely true in the field, the fluorine and barium zones. To obtain the chlorides in the fluid inclusions (and by inference in the ore-forming fluids) Davidson (1966) appealed to surface waters to dissolve chlorides from Permian evaporites; these would go down into the crust to depths sufficient to be heated enough to remove metals and fluorine from the wall rocks of fractures (presumably in the granite) and then finally rise to the proper formations in which to deposit the ores. To acquire the necessary heat, the ore fluids would have to penetrate 15 kilometers into the crust unless they made contact with a still molten (granite?) magma at a lesser depth. Solomon (1971) concluded, from a study of the sulfur-isotope ratios, that the sulfate in barite came from lower Carboniferous connate waters. He did not, however, think that the delta-^{34}S values of the sulfide sulfur in the ores could have been obtained from a magma or, for that matter, from connate sulfate either. He appealed, therefore, to a process by which the connate waters descended to considerable (or great) depths and underwent differentiation (though what was meant by that word or how the process would operate is not clear). At any event, these rejuvenated connate waters, with their delta-^{34}S values now right for the Northern Pennine ores would meet the fluorine-lead-rich waters or have been mixed with them at such times and such amounts as to make possible the deposition of the ores in the places and abundances known to exist.

This version of the genetic hypothesis for the generation of the Northern Pennine ores has something for everyone in it. Or, if you do not like the present mix, wait a year or so and some further addition will be made to the complex picture already painted to solve (finally?) the problem of the genesis of such ores. If only we could return to Dunham's simple 1934 hypothesis.

Now to consider the Askrigg block that is located in the Yorkshire dales (the second home of James Herriot), Swaledale, Wensleydale, Wharfedale, and Nidderdale. This block also is bounded by faults or folds on three sides and by covering Permo-Triassic and Jurassic rocks to the east (see the second paragraph of these notes). The sedimentary sequence (Ineson, 1976) is comparable to that north of Stainmore. The rhythmic sequence in the Alston block is well developed in the northern part of the Askrigg; in the southern Askrigg, however, the upper part of the upper Limestone group and part of the middle Limestone group have been cut out by the transgression of the Millstone grit across the eroded surface of the limestones. As is true in the Alston block, the thick Basal (lower) limestone (Great Scar limestone) outcrops abundantly in the south of the Askrigg.

The ore deposits of the Askrigg are in ribbon veins and stratiform bodies; in Swaledale, Wensleydale, and the northern part of Wharfedale these ores occur in thin hard beds of the rhythmic sequence and in the Main limestone (probably the equivalent of the middle Limestone group of the Alston block). In the southern part of the Askrigg (Wharfedale and Greenhow), the Great Scar limestone is the host rock to the ore. On the contrary, in Grassington and Nidderdale, the mineable deposits are in the basal Millstone grit.

In the same manner as the Alston block, the Askrigg is underlain by granite, an elongate, east-west-trending mass of igneous rock. No equivalent of the Whin sill is known in the Askrigg block.

The ore bodies of the Askrigg are veins in faults that have only minor displacement. In the northern part, the fault-vein pattern is east-northeast and west-northwest; whereas in the central portion, north-northwest veins dominate. In the southern margin of the field, the preferred direction of the fault veins ranges through 40° to 20° on either side of an east-west line.

The average deposit is small, though they are numerous; any single deposit therefore is short-lived and the total output of each is minor in amount. It is possible that, in the Greenhow area, where exploration was being carried out in 1976 and before, large reserves may be found at depth.

In contrast to the broad zoning of fluorite and barite in the Alston block, fluorite in the Askrigg is confined to a number of small and separate areas; in these, fluorite usually does not amount to more than 10 per cent of the vein filling material, in which the main minerals are galena, barite, barytocalcite, and some witherite. Around each of these fluorite centers, the gangue of the lead veins contains no fluorite but instead is barium minerals or calcite. Silica is notably absent, either as quartz as in the Alston or as chert as in North Derbyshire. Notable parts of the mineralization are in crustified banding, and polyphase mineralization is common.

No such direct relationship appears to exist between the Wensleydale granite beneath the Askrigg ore area and the ore areas as is true in the Alston block. Ineson (1976) says than an interpretation of the gravity and magnetic data (Bott, 1967), as was done for the Weardale granite under the Alston block, suggests that coincidence between zones of fluorite mineralization and granite cupolas cannot be established as it was in the Alston block. Ineson, however, considers that the fluorite centers do indicate feeder zones.

Little work has been done in the Askrigg block on wall-rock alteration, fluid inclusion temperatures of filling, and isotopic-age determinations. Data on isotopic analyses of sulfur and galena from the Greenhow area are given by Mitchell and Krouse (1971). They believe that these data fit a theory that the sulfides were deposited from an Na-Ca-Cl brine, relatively homogeneous, and that its source was in the lower crust and related to the granite basement. They thought that the "granitic basement" had been produced by anatexis of earlier rocks during the Caledonian orogeny. This implies that the ultimate source of the ore and gangue elements was from Paleozoic and Precambrian sediments and volcanics. They proposed two ideas for the manner in which remobilization of the elements of the Askrigg ore bodies was achieved.

The first of these mechanisms was assumed to be similar to that by which the Cornish deposits were formed, that is, fluids of deep-seated genesis moved upward into the Askrigg block through channels opened (or reopened) in Variscan time in the granite underlying the block. These fluids ultimately reached the actual sites of Askrigg block ore deposition and there formed the Askrigg ores. The second suggestion was that the crust below the Askrigg block was partially melted and produced ore fluids that contained the anomalous leads found in the Askrigg ores. They base their idea of a partial melting of the crust on the presence in Derbyshire of basaltic lavas and tuffs and of the Whin sill and its associated dikes in the Alston block, the presence of these igneous rocks being obvious proof that some melting had taken place at depth. These ideas are all very well, but they are based on rather tenuous reasoning from a rather uncertain data base. I suspect that there is appreciable truth in what they surmise, but their ideas hardly mesh with those that have been put forward most recently and with most acceptance in the Pennine ore districts. It is almost certain (to me at any rate) that ores so similar to each other as are those of the three Pennine districts could not have been formed by waters of compaction in North Derbyshire and the Alston block and by ore fluids derived from igneous magmas for the Askrigg block ores.

My decision as to how to classify these deposits is easy. They certainly were formed by solutions at low temperatures and pressures from solutions of high salinity. These deposits, therefore, are telethermal no matter from where they came or how they were produced and are so categorized here.

In 1971, Mitchell and Krouse published a study on the ores of the Greenhow-Skyreholme area in the Askrigg block of Yorkshire. They consider that the ores in this area were emplaced epigenetically and are strata-bound only in the sense that they are localized in favorable horizons. Most of the deposits are in fault fissures of small displacement. The fissures contain appreciably amounts of breccia, and the breccia fragments were fully or partially cemented during the mineralization process. Much open-space filling by the mineral material also took place, and much of this shows rhythmic deposition. The ores in this area are confined entirely to the Visean limestones (zones S_2, D_1, D_2 - see the notes of the North Derbyshire deposits).

Part of the ores of the Greenhow-Skyreholme district are located to the north of the major North Craven fault and part to the south. On the north side of this fault, the ore veins are related to domes on the east-north-east-striking Greenhow anticline. The anticline is cut by a few large faults that run in various directions and numerous small fractures in which the ores are contained that strike mainly in an east-west direction. South of the fault, the ores are in fractures that cross the axis of the Skyreholme anticline and strike, on the average, about east-southeast. Whether or not the two anticlines were once one structure before the developemnt of the North Craven fault apparently is uncertain. The mineralogy of the ores is typical of much of the district; galena is the only sulfide and fluorite the major non-metallic ore mineral. The other non-metallics are calcite and barite. These authors say that a crude mineral zoning obtains in the area with fluorite-rich centers being located at Greenhow, Gill Head, and Burhill. The veins at Greenhow (and Black Hill) are abundantly mineralized with galena, fluorite, mainly a pale violet, and calcite. The veins at Gill Head and Burhill are almost entirely filled with pale yellow fluorite; galena is rarely present. The areas are similar to the fluorite-rich zones in the Alston block (Dunham, 1934); Dunham considers that these zones may indicate the location of feeder channels through which ore fluids rose from great depths. Mitchell and Krouse point out that this conclusion was reinforced by Sawkins' fluid-inclusion studies (1966). From the Caledonian age of the granites under the Askrigg block, these could not have been the source from which the ore fluids came, but through-going fractures in this granite may have been the channelways through which ore fluids, originating at even greater depths, may have reached the Visean limestones in the Askrigg block. Against this argument is the fact that the granite beneath the Askrigg block does not have cupolas (as proposed by Bott, 1961) under the fluorite-rich areas as is the case in the Alston block. However, it can be argued that it is not unreasonable to assume that channelways may exist in parts of the granite not involved in the cupola structures.

The range of delta-^{34}S values in the ores from the Greenhow area of the Askrigg block is small, the standard deviation being $\pm 1.1^{\circ}/_{\circ\circ}$ and the mean delta-^{34}S values is $-2.2^{\circ}/_{\circ\circ}$. These are values to be expected from magmatic hydrothermal deposits. Most Mississippi-Valley-type deposits have a much wider range of delta-^{34}S values. Whatever the problems are in interpreting the wide-ranging delta-^{34}S values of most Mississippi-Valley-type ores, they need not concern us here. The sulfides in the Askrigg ores are slightly more negative than those of basalts that Mitchell and Krouse consider definitely to be from a mantle source. This can probably be explained by the fractionation of ^{32}S into the ores (and the resulting concentration of ^{34}S in the ore fluid that eventually escaped to the surface and was dissipated). Thus, the sulfur-isotope evidence definitely points to a magmatic source for the Askrigg sulfur (and, by inference, for that of both the Alston block and the North Derbyshire district).

Mitchell and Krouse have calculated the Holmes-Houterman model ages for the Pennine leads for which they have isotopic composition and found that all of them gave ages that are far to very far lower than the geologically determined ages. This means that the Pennine leads are anomalous or J-type (J for Joplin) in that they indicate, by the methods of mathematical analysis in present use, that the ores are far younger than the geologic relations of the rocks that contain them permit them to be.

Mitchell and Krouse also comment on the age of mineralization in the Pennines as determined by $^{40}Ar/^{39}Ar$ dating methods as used by Dunham and his colleagues (1968). Dunham's results indicated that Pennine mineralization took place in three stages, at 280, 230, and 170 million years ago. Dunham used Moorbath's lead ages, which Mitchell and Krouse think are invalid, particularly for Moorbath's 280 million year date. Secondly, they are not clear as to what the 230-million-year and 170-million-year ages refer. In addition to objecting to the manner in which Dunham interpreted his and his colleagues' results, Mitchell and Krouse point out that several other ways of interpreting his results would be valid, given what is known today. Dunham and his colleagues worked with very fine-grained material, and the validity of results from such material is uncertain. Further, they contend that mechanical crushing may liberate appreciable Ar and that certainly the actual physical size of the grains studied influences argon retentivity, the smaller the grains, the less well argon is retained. Thus, they conclude that, even if the White Whin (altered Whin sill) was produced during ore mineralization, the ages given by Dunham may represent no more than the differential loss of argon in the different-size fractions of the alteration zone. Mitchell and Krouse note that similar effects have been determined in deuterically altered Whin sill material in Teesdale where dolerite containing aggregates of clay minerals give an appreciable spread for ages for mineralization; they believe that the spread is due to sample inhomogeneity of grain size rather than to instrument differences.

Mitchell and Krouse think that, once an environment favorable to the formation of anomalous leads has been established, two processes exist by which lead may be extracted from this source to form ore deposits.

The first of these derives from data obtained from the Cornish ores. There ages as young as 50 million years have been obtained from the leads. They suggest that these different ages could be obtained by the escape of ore fluids over a long period of time (in geological terms) from a long-continued pool of still molten material deep beneath the surface. Such a process could also have obtained in the Pennines. Their second alternative is that the ore fluids could have been derived from the differentiation of masses of partly fused material in the lower crust of the Pennine basement. This differentiation could have produced fluids containing anomalous leads and these could have transported and deposited the leads in the Pennine districts. Either of these mechanisms well might produce several separate ore fluids that would deposit in several centers of mineralization, each of which would have its own isotopic lead composition. Mitchell and Krouse favor the lower crustal source for the ore fluids, but they admit that present-day data are not available to distinguish which, if either, of the hypotheses is applicable to the Pennine districts.

Whatever the source and composition of the Pennine ore fluids, they certainly deposited their mineral loads at low temperatures; the mineral suite is typically that of the telethermal range, a fact confirmed by the lack of recoverable silver in the Pennine ores. The deposits are, therefore, here categorized as Telethermal.

NORTH DERBYSHIRE

Early Mesozoic Fluorite, Barite, Lead, Zinc Telethermal

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Notes

Lead has been mined from the Peak district in northern Derbyshire since the time of the Romans. Mining has continued in the district, with various degrees of activity and changes in minerals sought. At present, fluorspar is the main material mined (some 150,000 tons annually) with important amounts of barite (20,000 tons per year) and lead concentrates (4000 tons annually). Zinc apparently is not now mined in any appreciable amount, but it was abundant in the Mill Close mine and essentially nowhere else in the district. Of the 680,000 tons of lead concentrates produced from the district through World War II, Dunham (in Schnellmann and Willson, 1946-1947) reports that two-thirds came from one mine, the Mill Close, and, of the

80,000 tons of zinc nearly all was mined at Mill Close. The Mill Close mine, although in the lower Carboniferous limestones is located under a cover of Millstone Grit in a salient that extends west from Matlock; the Mill Close vein strikes irregularly north-south from the center of the south boundary of this salient. This mine is discussed in more detail later in these notes. Although chalcopyrite and pyrite are common in trace amounts principally as inclusions in fluorite and calcite, these minerals, plus several other copper minerals were produced in quantity only from the Ecton mine, near the western border of the ore field. Silver is present in the galena, normally less than 0.005 per cent but locally as much as 0.05 per cent has been found.

The North Derbyshire (Peak) district is about 40 kilometers long from north to south and its greatest width is about 20 kilometers, but most of the working and worked deposits were in the eastern half of the area. The ore is confined to a gently undulating plateau, the highest elevations of which are 450 meters above sea level. The plateau is surrounded by outward dipping scarps of the coarse sandstones of the Millstone grit that rise to an elevation of 600 meters and is deeply dissected by narrow dales (valleys).

The mining of lead ended with the end of the 19th century, but fluorspar mining continues to be important. Exploration for additional deposits of fluorspar is being actively conducted.

The host rock of all the workable deposits is the Carboniferous, but some 30 areas of minor galena mineralization are known in the Triassic Bunter sandstones. The Carboniferous stratigraphy is most complicated by numerous changes in facies both laterally and vertically, and, as yet (Ford, 1976), no synthesis has been developed that encompasses the entire mass of Carbonate rocks in the Peak district. The sequence of limestone beds is about 500 meters thick on the average, and its base is not exposed anywhere in the district. Two holes have been drilled through the limestones, one showed 300 meters of limestone beneath the lowest outcropping at the surface; the hole bottomed in volcanics, the age of which is not known. The second went through 1800 meters of limestone and was stopped after reaching Ordovician slates. The hole penetrated anhydrite lenses near the base of the limestones.

The rocks that can be seen at the surface have been divided into paleontologic zones, known as C, S, and D (from bottom to top). Each lettered zone is divided into two parts on the basis of diagnostic index fossils, mainly corals and brachiopods. The validity of this scheme was under question (Ford, 1976), but he knew of nothing better to use in stratigraphic studies in the district. The main part of the outcrops in the ore district is composed of D_1 and D_2 subzones, the greater part of which is massive calcarenites; chert nodules are common in the upper beds. Around the northern and western borders of the ore field, reef complexes are impressively exhibited; they contain highly fossiliferous calcilutite bioherms. The material deposited out in the basin at the same time as the development of the reefs is poorly exposed and, only a short distance from its outcrops, dips beneath the surrounding Millstone grit. Within the D_2 subzone, shallow basins and swells were developed, and these in places are aligned with later folds.

Within the D subzone, numerous occurrences of basaltic volcanics, contemporaneous with the sediments, are known; usually they take the form of bedded lavas and tuffs with very few necks and sills and almost no dikes. The basalts are known as toadstones, either from the German "todt stein" or from their green and yellow color. The use of these volcanics as stratigraphic markers is limited because they thicken and thin so irregularly that it is difficult, if not impossible, to trace individual beds over any appreciable distance. The greatest thickness of the flows is about 30 meters, but a thickness of 20 meters is much more common. Some narrow tuff beds are altered to green clay known as "way boards".

The toadstones (Ford, 1976) are considered to have exerted considerable control over the upward passage of the ore-forming fluids because of their overlap and varied thickness. The toadstones also control the present distribution of groundwater and thereby had much influence over what ground could, and could not, be mined in the days before modern pumps were developed.

Ford (1976) says that the general paleogeographic picture was of a lagoonal area partly surrounded by reefs to the north, west, and south with deeper-water basins beyond them. Locally, within the lagoon, deeper water also occurred; mounds of crinoidal debris also were accumulated within the lagoon. Volcanic activity took place irregularly throughout the life of the lagoon to lay down ash falls and lava flows on the lagoonal floor and even down the reef-front.

The three fossil-divided stratigraphic sequences all are lower Carboniferous (Avonian); at the end of the Avonian, minor uplift and warping produced anticlinal crests that were considerably eroded, and the debris, in places very coarse boulder beds, covered the eroded reef front.

The rocks overlying the lower Carboniferous are upper Carboniferous; the formations are (from bottom to top); (1) the Millstone Grit series, which is a cyclic sequence of deltaic sandstones that alternate with shales of Namurian age and (2) the Coal Measures, also a cyclic sequence that has fewer sandstones than the Grit series and contains many coal seams - all of which rocks are of Westphalian age. Both the Millstone Grit series and the Coal Measures are each about 1500 meters thick at the north end of the ore field and about half that at the southern. The lowest formation of the Grits (Edale shales) ranges considerably in thickness, depending on the structure over which it was laid down. For example, the Edale is some 270 meters thick at the north end of the district, but it is no more than 50 meters at the farthest east tip of the ore field. The thickness of the Edale, however, is such that this formation probably provided a nearly complete seal over the limestones in the area during mineralization. The impervious character of this seal is shown by the presence of no more than traces of mineralization in the Millstone Grit and in the Coal Measures. In Carboniferous times, the lower Carboniferous limestones must have been buried under between 1500 meters (south) and 3000 meters (north) of Millstone Grit.

The area of mineralized limestone usually is called the Derbyshire "Dome". The structure, however, is much more complex than a dome. The eastern portion of the ore field is composed of a series of east-west plunging anticlines with east-west axes. The up-folding of these structures began before sedimentation ceased, as is demonstrated by the changes in facies between anticlinal crests and synclinal troughs. In some areas, these east-west folds have been so distorted as to produce south-facing monoclines at two places along the eastern margin of the ore area. Where the east-west folds are followed east from the ore area, they appear to curve to the southeast where the Derbyshire coalfield is reached. Locally, high points on these folds form two inliers in the Millstone Grit east of the ore field. Both of these structures contain many veins, the mineralization in which is dominated by fluorite. On their western ends, the folds merge into the usually north-south crestal area. In fact, in the west and southwest of the ore district, both folding and faulting trend strongly north-south. These trends date from the post-Triassic as they also are seen in the Staffordshire coalfield even farther west. The dips on the folds normally are less than 20°, generally between five degrees to 10°. Only in the marginal reef complexes are the dips as high as 40° to 50°. The deformation of the area was further increased in late Permian, post-Triassic and Tertiary times as can be inferred from the stratigraphy of the surrounding areas.

All faults in the lower Carboniferous limestone have been mineralized along at least parts of their lengths, and these are not distinguished from veins in mapping. The northwest-southeast set of veins, although they are mineralized, show little displacement; the monoclinal folds, however, grade into major east-west faults in the south-central part of the interior of the mineralized area. Even more important northwest-southeast faults, in the same general area, produce graben troughs as much as one kilometer wide. Most of the east-west faults and some of the northwest-southeast faults exhibit horizontal slickensiding; they are almost vertical. If the beds they cut dip gently, the displacement appears to be normal, but actually, these are wrench faults with small east-west displacement. This desplacement is, in places, right-handed and, in others, left. The minerals filling the veins locally are brecciated, refaulted, and re-opened by lateral stress and tension. In the west of the district, some north-south faulting is present, though little mineralization is known in any of these structures.

The age of the faulting is not certainly known, although it is post lower Carboniferous and, from the re-opening during mineralization, the faulting must have been at least partly contemporaneous with the mineralization.

Ford and Ineson (1971) believe that the stress field changed considerably with time; first a north-south compression caused folding, then east-west compression produced both folding and faulting in the west of the district, then, through lateral (northwest-southeast?) stress, wrench faulting occurred, then north-south tension re-opened the wrench faults to permit later mineral deposition. But, even this rather complex sequence of events does not account for the richest mine of all, the Mill Close, where the ore is in a series of en echelon north-south fractures that cross an east-west syncline that is outside (east of) the area of the outcrop of lower Carboniferous limestone.

After the deposition and lithification of the upper Carboniferous Edale shale and the higher rocks of the Millstone Grit, the rocks of the district were upwarped and this permitted erosion to remove much of the cover rocks from over the lower Carboniferous limestones in the present area of their outcrop. This removal was mainly in the southern part of the ore field. This period of erosion was closed by the transgression of the late Permian Zechstein sea. Ford believes that this advent of the sea allowed magnesium-rich brines to penetrate where cover was thin or lacking and to convert portions of the uppermost lower Carboniferous limestones to dolomite. After the deposition and lithification of Zechstein rocks, the rocks of the cover were upwarped in the early Triassic and much of the Permian cover was eroded away. This erosion was followed by the transgressive deposition of the Bunter pebble bed and sandstone as alluvial fans from the south; later a cover of Triassic Keuper marl was laid down. During the Jurassic and Cretaceous a thin layer of late Mesozoic rocks was deposited. These last rocks were removed by erosion as the result of an early Tertiary uplift; a more impressive late Tertiary uplift permitted the removal of the Triassic and remaining upper Carboniferous cover. The present land surface in the district was a result of Pleistocene glaciation and the accompanying periglacial cutting of the dales (narrow valleys) of Yorkshire.

The southern Pennines are flanked to the east, south, and west by nearly horizontal Triassic beds; these include the basal Bunter sandstone and pebble beds, then the Keuper sandstone in the middle, and above that thick Keuper marls that contain beds of gypsum. Ford (1976), Ford and King (1968) report that some 30 localities in the Bunter sandstones contain galena barite, and dolomite in voids between pebbles and sand grains. Most of these mineralized volumes are small lenses in coarse sediments in the vicinity of a clay layer and, in many places, near a fault. The ore fluids may have been remnants of the major ore-fluid volume that produced the major ore veins in the lower Carboniferous rocks and to have been channeled into their sites of deposition by faults and clay layers. The largest of these Triassic mineralized areas (Alderley Edge) is in northeast Cheshire, just east of the border with Derbyshire. The Alderley deposit is at least 10 km from the nearest lower Carboniferous-rock outcrop; this Carboniferous outcrop is a small, unmineralized inlier, some distance removed from the main lower Carboniferous rocks. The Alderley ores are largely malachite-cemented sandstone with galena impregnations and much barite cement; they are some 1000 m above the base of the Triassic beds and are noted for a considerable content of cobalt, nickel, and vanadium minerals. These ores may or may not have any genetic connection with the lead-zinc mineralization of the main North Derbyshire ore field that is contained in carbonate rocks; the probabilities seem to be against such a connection. Warrington (1965) believes that the Alderley ores resemble the Red-Bed copper deposits of western North America; he also reports that age determinations on Alderley galenas suggest a time connection between the two mineralizations (North Derbyshire and Alderley Edge) at least. The Alderley ores, although they are largely restricted to definite horizons and lithologies in the Triassic beds, are so closely related in space to the fault system of the area that they probably were introduced by epigenetic processes.

Although the differences between the Alderley and North Derbyshire ores are great enough that both cannot have been developed by highly similar ore fluids, their generating fluids almost certainly were of the same class and time of introduction into their respective host rocks. The youngest North Derbyshire rocks to contain at least modest amounts of ore minerals are those of the Triassic Bunter. The rocks immediately above the Bunter, the Keuper marl, probably were impervious enough to have prevented the ore fluids from moving higher into the stratigraphic oolumn in appreciable quantities; it also is possible, however, that the ores were brought in before the Keuper was deposited. In either event, the age of the North Derbyshire (Peak district) mineralization probably was Triassic rather than Jurassic, although this possibility cannot be dismissed. It is, nevertheless, apparently better to classify the North Derbyshire ores (and those at Alderley Edge) as Early Mesozoic rather than Middle.

Galena is the only sulfide present in abundance in the North Derbyshire ores. Lesser quantities of sphalerite are present, but, as has been mentioned, the only economic source of the zinc sulfide was in the Mill Close mine, off the east edge of the carbonate-rock outcrop area. Sphalerite (except at Mill Close), along with pyrite and chalcopyrite, occur as traces, mainly as inclusions in fluorite and calcite. Bravoite is widespread as inclusions in fluorite, but it is almost unknown as a separate mineral. The only mineable mass of chalcopyrite was in the Ecton mine, some dozen kilometers due south of Buxton; but probably no more than 10,000 tons of copper in all were produced there. No definite genetic connection has been established between the Ecton ores and the lead ores of the district, although such may exist.

The major minerals in the North Derbyshire gangue are fluorite, barite, and calcite, whereas quartz and dolomite are much less in amount. Cerussite, smithsonite, and rosasite $[(Cu, Zn)_2(CO_3)(OH)_2]$ are present as the result of oxidation of primary sulfides. About 100 minerals in all have been identified from the district (Ford and Sarjeant, 1964). Fluorite appears to dominate in the east, calcite, in the west, and barite in between, this certainly is not a consistent rule, many variations from it having been noted. Fluorite occurs in several varieties: hydrocarbon-rich Blue John, cloudy yellow Crich spar that has drained or gas-filled inclusions, and pyritic fluorite; it has been suggested that these fluorite varieties are also zoned, but this arrangement also suffers from many exceptions. The barite is cream-colored and is the earthy-fibrous variety locally called cawk.

The paragenesis of the North Derbyshire ores has not been thoroughly studied, but it appears that several stages of mineralization have taken place with repeated faulting and brecciation occurring between successive generations of fluorite and barite; galena is mainly confined to the early stages. Ixer (1974), however, has argued that the mineralization (at least in the Masson-Hill flat) was a single, long-drawn-out event with fluorite beginning precipitation early on; barite was appreciably later and, Ixer thought, ceasing to deposit before fluorite stopped. He believed that galena and sphalerite deposited over only a short period in the latter half of the mineralization cycle. Calcite began early but deposited heavily only near the end of the cycle. From the diagrams that Ford gives, it appears that the order of ore and gangue mineral crystallization was determined by the relative concentrations of the mineral constituents in the ore fluid, which must have differed widely from place to place. Nor does it seem that the relative concentration of the mineral materials was constant in any one ore fluid. Cinnabar and greenockite are trace minerals.

In many of the deposits, hydrocarbon compounds are present, some 30 having been identified. These mostly are paraffins that resemble closely organic materials contained in the country rocks and similar compounds in the oil of the East Midland fields. It seems probable that these hydrocarbons had no genetic connection with the ore-forming fluids but that they were caught up in these fluids as the latter travelled upward and outward from the source area.

The North Derbyshire ore bodies are locally divided into four types: (1) Rakes - These were produced by the filling of major faults with ore and gangue minerals; originally the fractures were wrench faults, of usually near vertical dip, that were later opened by tension. In places, slight departures from a single direction of strike, plus some horizontal movement, brought convexities on a right- (or left-) hand wall into opposition to a similar structure on the other wall. This would result in a narrowing of the vein at that contact, whereas, the opposing concavities further along the vein would form open spaces of appreciable dimensions. This resulted in the common phenomenon of pinching and swelling on any rake-vein where the proper conditions of initial vein shape and later horizontal movement produced such confrontations. Most of the rake-veins were faulted after having been filled or partly filled by ore and gangue minerals, resulting in new open space that later could be filled by a new generation of ore minerals with fractured blocks of the older mineralization being cemented by minerals of the younger. Not only are slickensided walls common in rakes, but slickensides also may cut across the brecciated and cemented vein filling; the slickenside grooves normally are nearly horizontal. The rakes are strong structures, having been followed in many instances for more than six kilometers. The rake widths may be as great as 30 meters although the average is between five and seven meters. Despite some rakes having been followed downward for 300 meters, no rake was bottomed because the ore ran out - instead haulage or drainage problems forced cessation of mining.

(2) Scrins - These consist of open joints on which movement was negligible, if it took place at all. These tiny veins were nearly always essentially vertical with the minerals that filled or partially filled them growing in from the walls. Scrins are much shorter than rakes, seldom more than a few hundred meters in length, and few were mined below 100 meters from the surface. Scrins may branch out from rakes, and, where they intersect rakes, both structures may be considerably enriched. Just how narrow a rake may be or how wide a scrin is not defined, but it is agreed that there

are hundreds of named rakes but that scrins number in the thousands. Most rakes trend east-west, but scrins can differ in strike from northwest-southeast to northeast-southwest; probably the nearer a wide scrin is to being called a rake depends on its strike.

(3) Pipes and Flats - Just as it is difficult to say where rakes end and scrins begin, so it is difficult to distinguish between pipes and flats. A flat is defined by Ford as a strata-bound ore body in which width and length are much the same; flats more or less follow the bedding in the Carboniferous country rock in which they are contained with important exceptions. A pipe is a strata-bound deposit in which length greatly exceeds width. Pipes are elongate flats much longer than wide; again, the problem is to decide at what disproportion of length to width is the ore body a pipe and not a flat. Actually, this is no real problem as a quick description of the ore body will tell the average ore geologist which term he ought to use; only a few will be borderline cases. Although Ford (1969) calls the flats and pipes stratiform, he really means strata-bound since, both flats and pipes can meander through one or more formations as his diagrams show. In fact, some pipes are fillings of karst features of various stages of development, but these pipes do not appear to cross from one formation to another. Flats, on the contrary, are not formed in karst features but, instead, appear to follow bedding planes and low-dip fractures that may cross-bedding planes to connect one favorable bedding plane with another. Further, flats may be flat (or essentially so) when following one bedding plane, steepen somewhat to travel upward to the next bedding plane or to be ponded beneath a toadstone or shale bed where they flatten again. It appears, from Ford's diagrams (1969) that some flats actually cross from one formation to another (within the carbonate sequence) so not all of them, at least, are strata-bound. Thus, pipes seem to be connected with fissures and karst cavities, whereas flats are governed by such features as bedding planes and impervious beds. Despite these differences, an elongate flat that has no connection with karst features may nevertheless be classed as a pipe. The entire problem is complicated by the fact that almost all mines in pipe and flat ore bodies have been abandoned and can no longer be seen to make certain if they obey the rules just laid down. An exception to this last statement appears to be the flat at Masson Hill, near Matlock that is described by Ixer (1974). This ore area will be discussed later in these notes, after the Mill Close Mine.

Ford (1969) mentions three types of pipe settings that may help clarify the problem. The first of these is the Blue John Fluorspar fillings of caves and voids between large boulders (up to 10 meters in diameter) in the basal Namurian conglomerate. Such a pipe appears to have been fed from an adjacent rake, but the rake also seems to have provided the material that fills cavities in a karsted reef complex in the Carbonaceous beds immediately under the conglomerate. Although Ford's diagram is in two dimensions only, it would appear to me that these ores are flats rather than pipes.

The second is a typical series of bedding plane and inclined channelways that connect one bedding plane with another, and all are filled with ore. They cross at least two, probably three, formations so are, in the strict sense, not strata-bound at all.

The third shows cavities growing upward by means of hydrothermal karst activity. These cavities later were filled probably not much later and probably through activity of the same cycle of solutions that produced the karst cavities. The limestone between the upper and lower bounding planes of toadstone appears to be of one formation, and the ore-bearing structures probably are correctly designated as strata-bound pipes.

The fourth is the ore structures produced by the filling of solution cavities developed alongside of rakes. These are nearly vertical structures and probably cannot be called anything but pipes even though they are hardly pipes in any classical sense. Of interest is that their filling involves a certain amount of replacement, a process not as important in typical Peak district pipes as it must be in bedding-plane flats.

The two large copper pipes at Ecton are quite different, both structurally and mineralogically, from those in the main eastern portion of the Peak district. These pipes are localized around the intersections of two veins to produce a more or less cylindrical and vertical ore body. These ore bodies, now abandoned and flooded, are known to have been mineable for at least 300 meters below the surface but what their relationship was to the tightly folded and faulted host limestones that the intersecting veins cut is impossible now to determine. The abundant chalcopyrite in these deposits, however, so markedly different from the mineralization in the remainder of the Peak district, places these deposits outside the category of North Derbyshire ores.

The pipe deposit at Masson Hill, classed by Ford as a pipe and by Ixer (1974) as a flat, is an elongated combination of void filling and replacement in a junction zone between dolomite (above) and limestone (below). The ore body appears to depart only slightly from the horizontal and to be confined in its rather narrow length by parallel sets of scrins on its sides and by toadstone layers (locally no more than wayboards) above and below. The main mineralization results from replacement of sedimentary calcite and dolomite and quartz. This mineralization, however, is typical of the fluorspar deposits of the Peak district, consisting mainly of fluorspar and quartz with lesser barite, calcite, and galena. Locally, the Masson and Jughole deposits contain a layer of mixed sulfides - galena, pyrite, bravoite, pyrite, marcasite with traces of chalcopyrite and sphalerite. It typifies the problem of properly naming the more or less horizontal ore bodies; it certainly is flattish, but it also is much more elongated than equidimensional. Therefore, Ford refers to it as a pipe even though it does not have the vertical character normally thought of as being characteristic of pipes.

The huge (by Peak standards) deposit of Mill Close is unusual in that, for its entire length it is covered by Edale shales and does not outcrop. It was discovered by following surface outcrops of minor veins on the limestone outcrop south of the contact of limestone with Edale shales. Farther north, these veins made contact with others of far greater widths and depths than had been shown by the veins followed from the outcropping limestone. At Mill Close, the ore has been mined down from the overlying shale by following veins and vertical pipes until a huge flat (or manto) was reached under the seventh (counting down from the shale contact) lava-pyroclastic toadstone layer. The combined vein (Rake-fault-joint system) proceeds (starting at the south end of the structure) along the Old Mill Close Joint (essentially of north-south strike) to a junction with the northwest-striking Mill Close Fault; this, in turn, intersects the Main Joint (actually a fault of minor displacement) and follows that structure to about 2100 m north of the zero coordinate where it is connected through rather indeterminate fractures, to the north-northwest-striking Pilhough fault where the mineralization appears to die out, probably cut off against the west-southwest-striking Long Rake. To the south, at least 3650 m south of the Long Rake, the Coast Rake (also a west-southwest rake) cuts off the ore to the south. The maximum length of the mineralized zone appears to be about 4.5 km.

The dominant structure of the Mill Close area is a broad anticline, the axis of which strikes roughly northwest-southeast. The northwest system of gash veins (rakes) developed as a reaction to this folding. The north-south system of joints probably was due to a later stage of this folding (Traill, 1939). After a quiet interval, a minor north-south crumpling along the eastern flank of this anticlinal dome formed a series of broad folds (with an east-west strike) that died away gradually to the west (towards the main north-south anticlinal axis). The structures that localized the Mill Close ore bodies are on the southern limb of one of the minor east-west synclines and are limited to the north and south, respectively, by the Coast and Long rakes. Thus, the dominant east-west structure is a trough that pitches gently eastward; it has an east-west length of about six kilometers and a width, between the two rakes, of about four kilometers. The syncline is not symmetrical as the dip of the southern limb is quite gentle and that on the north flank is quite steep.

The folding caused the opening of a large number of joints that fall into three systems: (1) Northwest-southeast gash veins that range in strike between north-northwest and west-northwest; (2) North-south joints, and (3) generally east-northeast-west-southwest fractures, although individual fractures depart somewhat from this direction. System (3) provides by far the greatest number of veins, and these range in size and continuity from the great rakes (Long and Coast) to small scrins (or stringers) that intersect the members of the other two systems. All of these three sets of fractures potentially may contain ore, and the main line of ore bodies shifts from one system to another along the generally north-south direction they follow. Most of the fractures are true joints and, where these are not enlarged by solution, they are the thinnest cracks. The few true faults have vertical displacements, in most instances, up to 2.5 meters, with the upper limit being five meters or less. Most of the actual faults belong to the north-south system and a few to the northwest-southeast system; no fault is of the rake system. Although the faults are minor enough in displacement to create essentially no problems in mining, they certainly provided channelways through which the upward-moving ore fluids could travel. Because the toadstones break less cleanly than the limestones and because they are less soluble than the limestones, the ore fluids normally had to find a way around the lens-like toadstones if they were to reach higher into the formation sequence; but some ore-passage faults cut through toadstone.

The ore-body size and shape at Mill Close was determined by the position of the toadstones in the sequence, the degree to which the limestones were fractured, and the continuity of the fractures. Since not all the beds are flat in the Mill Close area, ore fluids, in many places, followed steeply dipping beds upward and usually bypassed the toadstones only when the beds became flatter and the toadstones lensed out. On the lowest level of the mine, where the ore fluids apparently were much blocked off by the lowest toadstone layer, they not only rose upward through the formation, utilizing fractures that bypassed the toadstones but also took advantage of karst features developed in the lowermost portion of the rock volumes to fill open space and to replace debris on the floors of the karst cavities. Other quantities of ore fluids worked their way out into the walls of the major and minor fractures and replaced solid rock to add to the total size of the manto.

Traill (1939) outlines the history of the deposit as follows:

(1) folding along a northwest-southeast axis to produce the northwest-striking joints in limestone; the fold plunges northwest at a low angle; later this same tectonic episode caused the opening of north-south joints and east-northeast-west-southwest fractures

(2) development of karst cavities on either side of these fractures; cavities are confined between over- and underlying toadstones; solutions moved upward from northwest-southeast

(3) cavities along joints eventually produced more or less circular inclined pipes, average diameter about six meters

(4) further increase in cavern sizes, eventually removing enough support from toadstone roofs that they collapsed, allowing blocks of toadstone and limestone to fall to the cavern floor

(5) with the essential completion of cavern formation, the entire area later to be invaded by ore-forming fluids (as opposed to cavern-forming ones that probably were closely related in both genesis and time) would resemble a huge tunnel to a greater of lesser extent filled with blocks and finer rubble of limestone and toadstone

(6) invasion of the karst tunnel by ore-forming fluids that not only deposited to a minor extent in open space but metasomatically attacked the limestone (but not the toadstone blocks) by the introduction of very fine-grained fluorite and sphalerite

(7) simultaneous bleaching and conversion to clay minerals of the toadstone blocks, in which clay matrix galena was deposited

(8) finally, the crystallization of brown sphalerite and calcite in the remaining open space with probably additional galena.

This outline probably does not do justice to the complexity of the mineralization, which almost certainly was introduced in more than one phase with partial or complete duplication of the mineral paragenesis in each successive phase.

Traill points out that certain beds definitely were more favorable to karst attack and to the metasomatic emplacement of the ore and gangue minerals. All of the favorable beds were massive, light-colored, more or less coarse-grained limestones. The dark, thin-bedded, cherty, or shelly and coralline limestones were appreciably less favorable but, in places nevertheless, were rather well mineralized but in small and irregular bodies.

The principal primary minerals in the ores of Mill Close (Traill, 1939) are galena, sphalerite, pyrite, calcite, fluorite, and barite. The galena occurs as crystals or masses in calcite gangue or largely enclosed in toadstone-derived clay. The silver in the galena is only one to 1.5 ounces per ton of galena, not worth recovering. The much less abundant sphalerite has, however, a much wider distribution since it is found along joints far beyond the areas where galena was economically worth mining. In the uppermost mineralized limestone, sphalerite was the earliest ore mineral and was followed by fluorspar, galena, barite, calcite, and pyrite. At greater depths, at least a considerable portion of the sphalerite was later than the galena. At these depths, the usual paragenetic order was calcite, galena, more calcite, sphalerite, and still more calcite. Sphalerite was deposited in at least two different stages, one older and one younger than the galena; the earlier sphalerite was normally fine-grained and black and was a replacement of fluorite-containing limestone. The later sphalerite was in honey-colored to brown crystals. Both types of sphalerite contained minute specks of chalcopyrite (?). The sphalerite was quite pure, averaging about 65 per cent Zn; it also contained an average of one per cent Cd. Normally, sphalerite and galena were not intermixed in the ore, only on the 129 level did much such mixing occur. Pyrite, fluorite, and barite were much more abundant on the higher levels of the mine than on the lower. Some of the fluorite on the upper levels was covered with specks of marcasite.

Calamine (smithsonite) was the commonest secondary mineral and usually occurred as small crystals on calcite crystal faces. Cerussite was fairly common throughout the mine, indicating the considerable depths to which surface oxidation penetrated the ore. Greenockite was present as incrustations on various minerals, the cadmium, of course, having been derived from the sphalerite; greenockite did not show crystal faces. A few other rare secondary minerals have been found. Traill (1939) believed that the ore-forming fluids were hot-aqueous solutions forced up from great depths and possibly were derived from later phases of igneous activity.

Wall-rock alteration in the southern part of the district is quite widespread and consists of dolomitization and silicification. In the south, areas of wide-spread dolomitization are known around ore. In the north, dolomitization is minor and alteration is mainly progressive loss of original and diagenetic textures and an increase in minor fractures as the veins are approached. The more common the toadstone layers in a given area, the greater is the amount of silicification; this is particularly true in limestones close to basalt ash and lava. This relationship suggests to Ford (1976) that the silica was leached from the toadstone. Further, aureoles of trace- and ore-elements have been found irregularly distributed in the vicinity of several rakes. Pb, Ba, and Zn fall off within 10 meters of the major veins; the irregularity of distribution appears to be related to the intensity of the minor fracturing. Some regional soil sampling for metals and other trace elements has found areas of high background counts of Pb, Zn, Cu, and F, indicating that valuable mineralization might be found under them at shallow depths, particularly along the western margins of the limestone outcrop area.

Ford defines the Peak ores as strata-bound in the sense that they are confined to the Carboniferous limestones, but this is rather stretching the definition of "strata-bound" beyond that normally used. The stratiform pipes and flats (except the Ecton pipe) are truly such, but they are not syngenetic mineralizations but definitely were introduced into karst structures, developed by the early phase of the ore fluids, that are more or less parallel to the bedding.

Several theories have been put forward to account for the genesis of the ores, no one of which completely satisfies all ore geologists who have studied the problem. These theories are: (1) residual fluids from a granite magma at depth. No evidence of necessary granite beneath the Peak district has been detected by geophysical means - the youngest are quite appreciably older than the ores; (2) a buried granite as a channelway. A Silurian granite has been detected beneath Mount Sorrel 50 kilometers south of the Peak district; it is cut by a few mineralized veins. But this is a long way for ore fluids to have travelled, and this idea can be dismissed; (3) interstratal brines derived from evaporites. (a) Sub-Carboniferous evaporites: The high salinity of the fluid inclusions in the North Derbyshire ores suggests to Ford (1976) that the relatively thin evaporite beds beneath the Carboniferous limestones (70 meters) may thicken to the east. He believes that these thicker evaporites are in the proper position for compaction and diagenesis to have resulted in an upward migration of minor elements in solution. He adds that the patchy dolomitization of the lower Limestone beds in the two deep boreholes in the district suggests the presence of Mg-rich brines trapped beneath the unaltered limestones and toadstones of these areas, but the veins in these areas contain very little dolomite, probably effectively eliminating this concept. Such dolomites as are known in the Matlock area (near the Mill Close mine) probably were produced by downward-percolating brines of sub-Zechstein age. From these thickened evaporites, it is possible to conceive of heated, saline solutions having been produced under the North Sea and that these might have migrated into the North Derbyshire lower Limestones. It also is as easy, I think, to imagine that the ores are one of Merlin's miracles. (b) Post-Carboniferous evaporites. Ford also suggests that the evaporites in Triassic rocks of the English Midlands might possibly have been the source of mineralizing brines. He adds, however, that these evaporites are shielded from the lower Carboniferous limestones by their enclosure in thick impervious Keuper marl and by the Millstone Grit and Coal Measure cyclothems that contain thick shales that lie between the evaporites and the ores. Certainly this concept is invalid; (4) Local ground waters. These waters are sulfate-rich, but Ford asks: what possible relationship can such Recent waters have to mineral fluids in Permo-Triassic times? This question is all that is necessary to destroy this concept; (5) Derivation from host rocks. Nothing in the present condition of the host rocks suggests that the huge amounts of ore elements in the North Derbyshire ores can have been provided by local leaching; (6) Migration of deep formation water. Ford would move waters driven out of the lower Carboniferous limestones (apparently) under the North Sea in Triassic times when they were being compacted by some six kilometers of cover. Granted the appropriate thermal gradient between the rocks under the North Sea and North Derbyshire and granted the ability of the waters of compaction to dissolve the elements needed, if they were available, and transport them to the North Derbyshire of Triassic times, perhaps this concept is possible. There are, however, many unknowns, such as the actual bulk of the limestones below the North Sea, the concentrations that must have been present in these limestones to provide the necessary ore elements, the possible aid provided to the process by brines from the underlying evaporites, the assistance to migration that might have been provided by wrench faults that may have extended beneath the North Sea, and fractures that might have allowed some migration upward of something valuable to the ore deposits to come from the sub-Carboniferous basement. It seems almost impossible to believe in all these uncertainly needed events having occurred during the same space of geologic time and having joined together in sending a proper ore-forming fluid to the North Derbyshire ore area. Surely such a conjunction of unrelated processes to produce so huge an area of ore mineralization strains the credulity of any ore geologist not desperate to explain a phenomenon that does not fit into any preconceived mold.

Even granting that this lastly outlined process could bring ore elements to the North Derbyshire of the Triassic, where did the sulfur come from? Ford appeals to pyrite disseminated in the dark basinal limestones, to sulfurous hydrocarbons, or to bacterial reduction of anhydrite sulfate at the base of the lower Carboniferous limestone. Only the tooth-fairy remains to be employed. Ford also considers the problem of where the spent ore fluids went, but it seems that more needs to be certain about the ore fluids themselves before it is necessary to consider their spent residues.

If none of these suggestions of Ford's are valid, what can have been the source from which the ore fluids came and how did they obtain the ore elements they needed to contain? After the wild conjectures just summarized, the return to the old idea of an igneous source at depth seems to be too simple to be worthy of consideration. But, perhaps, that is all that is left. Of course, neither of the two deep holes drilled in North Derbyshire has found any granite of Permo-Triassic age under the area. But surely not enough holes have been drilled or enough geophysical work done to guarantee that no such igneous source exists.

At any event, the ore fluids were at a low temperature when they deposited the ores in North Derbyshire, and the mineralogy is typical of telethermal, Mississippi Valley-type ores the world over. Therefore, even without a firm conclusion being possible as to where they came from or how they were transported and deposited, it seems safe to classify them as telethermal in the broad sense, and this is done here.

WEST SHROPSHIRE, SHROPSHIRE, MONTGOMERYSHIRE

Middle Paleozoic	Lead, Zinc, Barite, Witherite	Telethermal (?)

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Notes

The ore mineralization along the border between England and Wales on the west edge of the English Midlands is divided into two areas: (1) the Shelve-Habberley area which encloses a circle with a radius of about 10 kilometers centered at Shelve village, 33 km southwest of Shrewsbury; most of this area is in Shropshire, but a small portion of the southern part is in Montgomeryshire; and (2) the Breidden Hills area that measures not more than 2.6 km^2. These hills are about 18 kilometers west of Shrewsbury.

In the Shelve area, the mineralized veins are in pre-Cambrian and Ordovician rocks; they are overlain unconformably in all directions but east by Silurian and Carboniferous rocks that there contain no mineralization. In the Breidden Hills, the ore bodies are in Ordovician rocks in the main but some of the veins reach into the Silurian beds.

Lead was mined in the Shelve area by the Romans; pigs of lead stamped with the seal of the Emperor Hadrian have been found, and the names of some of the mines at least suggest a Roman heritage - Roman Gravels and East Roman Gravels, but others - Sawpit vein and Second North vein - were worked from the Roman mine in those days.

Records of lead production were not really kept prior to 1845; zinc was not recovered before 1858 and barite before 1860. Between 1845 and 1913, lead ore totalled 235,650 tons, and the largest yearly production was nearly 8000 tons in 1875. Only about 19,000 tons of zinc ore were mined between 1858 and 1913, and barite was produced between 1860 and 1913 and nearly 1000 tons was the highest amount in one year (1882). Barite amounted to about 271,500 tons between 1860 and 1913, and the maximum annual barite production was nearly 14,000 tons in 1913. Except for the Huglith mine, barite production began shortly after 1913; in that mine nearly 300,000 tons were recovered between 1914 and 1940. This was the largest amount of ore (of any type) produced from a West Shropshire mine. The district as a whole probably produced about 500,000 tons of barite.

The division of the Ordovician rocks into formations was first done by Lapworth in 1880 and does not appear to have differed much from the stratigraphy that Dines (1958) presents. These beds, as they are locally known, are (from bottom to top): (1) Stiperstones quartzite, 77 meters thick, (2) Mytton flags and shales, 460 meters thick (the main host to ore veins), (3) Hope shales 245 meters thick (soft black shales), (4) Stapeley volcanic group, 275 meters (andesitic lavas interbedded with ash falls, grits, and shales, (5) Stapeley shales, thickness included with that of the Stapeley volcanics, (6) lower Weston grit, 61 meters thick, (7) Weston shales, 31 meters (8) upper Weston grit, 61 meters, (9) Betton shales and flags, 61 meters, (10) Meadowtown shales, 245 meters (with flags and limestones), (11) Rorrington shales, 122 meters (black mudstones), (12) Spy Wood grit, 53 meters, (13) Aldress shales, 215 meters, (14) Hagley volcanic group, thickness included with Hagley shales (ash falls), (15) Hagley shales, 245 meters, and (16) Whittery volcanic group, thickness included with Hagley shales (ash falls), (17) Whittery shales, 275 meters (some grits included). The thicknesses given here appear to be of approximate accuracy only.

The general strike of the beds is north-northeast. The ore area is cut into two parts by the Stiperstones quartzite in both the topographic and geologic senses. To the east of the Stiperstones are only Cambrian beds (the Habberley or Shineton shales); beneath the Habberley formation are (from top down) Precambrian Longmyndian Bridges group and Bayston group. The Pontesford baryte lode is contained in even older Precambrian rocks.

The Ordovician beds west of the Stiperstones Ridge dip gently in that direction until the synclinal axis (with a north-northeast trend) is reached that runs through Ritton Castle, about 2.5 kilometers west of the Stiperstones Ridge. The next anticline is reached 1.6 kilometers west of the synclinal axis; it has essentially the same trend and passes through the town of Shelve. The flags and shales of the Mytton beds come to the surface on the crest of this Shelve anticline and form an inlier in that area; the Stiperstones do not outcrop on this ridge. West of the Shelve anticline, the various beds dip more or less uniformly in that direction. The entire ore area (except to the east) is surrounded by Carboniferous beds. The area contains contemporaneous andesite and andesite tuffs, and, in the Silurian, laccoliths, sills, and dikes, of diabase and dolerite were intruded into the Ordovician formations.

The major faults in these Ordovician beds trend mainly northwest or north-northeast and minor faults (that have little or no displacement) trend east-northeast. The faults provided volumes of space into which both the igneous rocks could be introduced and ore minerals deposited.

The earlier of the two groups into which the igneous rocks can be included are contained in the north-northeast-trending fault fissures, and these rocks probably were introduced in Silurian time. The later dikes are in the fissures of northwest and east-northeast trend; these structures are more regular than those that trend north-northeast, dip more steeply, and are appreciably narrower. The fissures of these latter two trends (northwest and east-northeast) carry the mineral veins. In some places, both dikes and veins occupy the same fault structure. Where the vein material is in contact with the igneous rock, the dike rocks are altered to sericite; the friable white rock, so developed, is called locally White Rock or Whitestone. This relationship dates the dike rocks as older than the ores, but how much older is uncertain. Dikes and ores may have come from molten or fluid materials given off from the same magma chamber but were so given off at successively different times.

The fault fissures in most places are not a single fault plane, but they normally consist of a number of parallel fractures. The younger ones (northwest or east-northeast trends) in general dip southwest or south.

Locally, two faults may dip toward each other, in which case, the rock between them normally is shattered as is true in veins of the Roman and East Roman Gravels mines. These veins converge with depth, and between these veins are several smaller veins.

The faults are developed best in the harder formations such as the Mytton beds and the ash rocks and flags of the Stapely volcanic group. On the contrary, in the softer shaly rocks, the faults die out entirely or become so indistinct as to be most difficult to trace. If faults are present in these soft beds, they become filled with gouge and are, thereby, essentially barred to the entry of ore-forming fluids and are barren of vein material, or nearly so. Some faults may be traced for long distances, but, in these, economic ore occurs in those portions of the faults that pass between hard walls. As a major exception to the normal behavior of faults in hard and soft rocks, the Stiperstones quartzite contains essentially no ore mineralization even though it is a quite hard rock. In the flaggy top of the Stiperstones, some ore is known where bedding planes intersect faults.

Most of the workable ore is located in northwest and east-northeast faults where these structures, particularly the latter, intersect each other. Because few of the ore bodies are faulted, Dines believes that the mineralization took place after essentially all of the period of fault development was passed. He also thinks that only one period of mineralization occurred because, where ore has been brecciated by later (perhaps much later) fault movements, the breccia fragments never have been known to have been recemented by later primary minerals, sulfides or non-metallics.

Dines is convinced that the mineralization differs with the depth at which it was formed beneath the surface. The major concentrations of zinc are at the lowest levels in the veins in which they have been found, but there is considerable overlap between the zinc zone and the lead one over it. Although some barite and witherite are found above the lead ore, appreciable amounts of these two barium minerals are intermixed with the lead. In places a little chalcopyrite may be associated with sphalerite and galena, but most zinc-lead veins seem to be barren of the copper-iron sulfide.

One reason that the Mytton beds are so well mineralized is that they are capped by the largely impermeable Hope shales that acted as a shield against the upward movement of ore fluids. Where veins, which are strong and well mineralized in the Mytton, enter the shales of the Hope formation or even the shaly upper portion of the Mytton itself, they split into small, poorly mineralized stringers or die out altogether. Although the shaly portions of the Mytton are paleontologically correctly assigned to the Mytton beds, from the way they behave toward the ore fluids they are more reasonably considered the base of the Hope formation. The explanation for the veins in formations well above the Mytton is simply that there are flaws in the cover provided by the Hope beds, and some ore fluids are able to escape upward into the favorable environment provided by the strongly fractured Stapeley volcanics, the upper Weston grit, or the Hagley volcanics. But the ores in these higher localities and, therefore, well west of the Mytton beds, are mineralized with barite alone; only traces of galena and sphalerite are to be found in them.

Dines believes, with reason, that the only hope of finding more ore in the Shelve-Habberley district is in deep-level prospecting in the Mytton beds, where, for example, they lie at depth in the Ritton Castle syncline between the main outcrop of the Mytton and the Shelve inlier. To do this, however, would require shaft sinking or diamond drilling to slightly over 300 meters to reach the Mytton and 450 meters more to penetrate it completely. The central area between Ritton Castle (south) and Round Hill (north) is considered the most likely place for such prospecting.

The principal vein of the Roman Gravels mine in the Shelve area is the Roman vein that probably was the site of the main workings in Roman times. This vein is a fault-fissure with a northwest trend that crosses the entire width of the northeast-striking Mytton beds (that dip 25° to 40° northwest) in the Shelve inlier. The northeast side of the Roman vein is downthrown, and it dips toward the northeast. At its southern end, it bears gradually more westwardly until, where it leaves the Mytton beds in the South Roman Gravels mine, it dips vertically and strikes about N70°W. At the northwest end of the vein, the dip is 20°NE. The fault fissure offsets a sill of mafic igneous rock so that the north side of the sill has moved west relative to the south side. About one-third of the vein (at its eastern end) has not been explored.

The vein is filled by calcite with appreciable amounts of sphalerite, galena, and fragments of country enclosed in the calcite. The Romans began work on this vein as an open pit but eventually went underground to a depth of about 30 to 35 meters. In addition to the Roman vein, four other veins are associated with it to make up the Roman Gravels group; these four are: Caunter (or First North), the Second North, the Sawpit, and the Spring. The first two are northeast of the Roman vein and the other two southwest of it; the four are roughly parallel to the Roman vein. The Second North and the Sawpit also were worked by the Romans as open pits. The strikes of the four veins range from east-west to west-northwest; three dip east of northeast, and the fourth (the Spring vein) dips southwest.

The Snailbeach mine in the Ritton Castle area immediately east of the Shelve area consists of a group of about 10 more or less parallel veins that, in a few cases, actually cross or merge. They all cut through the outcrop of Mytton beds and barely penetrate the Hope shales to the west and the Stiperstones quartzite to the east; these veins are about two kilometers southeast of Habberley. The strike of the veins of the group is roughly east-west. The Mytton beds here strike west-northwest and consist of hard flags or grits with which are interbedded soft shales. The ore shoots largely are confined to the harder rocks and pitch to the west, following the dip of the Mytton beds. The veins all appear to dip south, but at various angles and normally range up to 2.5 meters in width but may reach a maximum of seven meters. Where veins join, the width at the junctions may be as much as 12.5 meters. In the wide parts of the veins, crystalline and granular galena occurs in small bunches in calcite; in the narrow parts, the ore minerals are present as ribs, mostly near the hanging wall. The sphalerite is mainly in compact brown and fibrous yellow varieties; sphalerite was essentially always less abundant than galena. Barite was confined to the upper levels of the mine and was practically free of sulfides. Witherite was present but was not abundant. Different veins were worked at different levels for different ores. Since 1845, Snailbeach produced 132,000 tons of lead ore, nearly 4000 tons of zinc ore, and over 42,000 tons of barite. From 1845 to 1920 (when Snailbeach closed), the lead ore mined was 56 per cent of the total output of West Shropshire and the zinc ore 23 per cent.

Directly to the east of the Stiperstones Ridge are the Cambrian (Tremadoc) beds; these lie beneath a slight unconformity with the Stiperstones beds above them. The Cambrian belt is about 1.2 kilometers wide; to the east these rocks abut against the Habberley fault, another north-northeast-trending fault. East of this fault, the country rocks are Precambrian (Longmyndian) red and green grits, conglomerates, sandy shales, and mudstones of various Precambrian formations. Some 10 kilometers northeast of Shelve are even older Precambrian beds (Uriconian). The Habberley fault is pre-Carboniferous; it does not cut such late Paleozoic rocks. The Habberley fault also is later than the east-west faults in the Longmyndian beds, these east-west faults are mineralized in some places with barite and traces of copper ore developed secondarily. The Cambrian rocks between the Stiperstones and the Habberley fault are mainly shales and include no ore veins.

The copper in the Longmyndian rocks may have been derived from primary copper minerals and might suggest a richer copper ore lies at depth. To reach its present position, this copper ore must have been uplifted to the level of the lead-zinc ores to the west by the upward movement of the east side of the Habberley fault.

East of the Stiperstones, no mineable veins of the lead-zinc ores ever have been found, but barite has been mined in important amounts. The barite in this area may simply have been moved laterally from the mineralized area west of the Stiperstones, but this seems less likely than that it is the top of a separate zone of mineralization that should be explored.

Although the area east of the Stiperstones and south of the latitude of Shelve contains no mineable ore, the same belt of rocks north of the Shelve latitude as far north as Pontesford Hill, some 11 kilometers, and east of the Habberley fault contains several moderately important mines and one major mine. The most important mines are in the portion of the belt that extends from Habberley south for some three kilometers. Five of the mines are in the youngest Precambrian Oakswood group (including the Huglith) and two are in the next older Bridges group. Although it is difficult to speak of the width of a belt that contains only seven mines, a good average is about 2.5 kilometers. On the other (west) side of the Habberley fault, the Cambrian formation is the Habberley or Shineton shale. The shale dips west beneath the Stiperstones quartzite and strikes north-northeast. These rocks are quite shaly and, therefore, are poor hosts to fault-filling ore bodies.

The Precambrian beds strike more or less north-south and dip to the east. The Oakswood and Bridges groups already have been mentioned. Under the Bridges is the Baystown group, into which the more or less east-west striking veins of the Cothercott mine appear to extend. Otherwise, the Baystown seems to be barren of ore or, at least, any ore that was of any economic importance.

These Precambrian formations consist of red and purple grits, sandstones, sandy mudstones, and shales. They are cut by several major faults that strike a few degrees on either side of east-west; they displace the rocks westward on their south sides. Both the major faults and some of the minor ones are mineralized, with the main ore mineral being barite. Dines believes (1958), as seems quite probable, that these ores, which contain small amounts of secondary copper minerals, may have had chalcopyrite as the primary copper mineral. Why Dines says that the barite in these deposits must have been contemporary with the barite to the west of the Habberley fault is not clear; perhaps it is because both types of deposits contain barite. At any event, no lead and zinc are associated with the ores in Precambrian rocks (as is true of ore in Ordovician rocks) but the strikes of the mineralized faults in the area east of the Habberley fault are at almost the same angles as those west of that fault. But does this prove that both types of ore came from the same source at much the same time? I think not.

The country rocks of the Huglith mine in the Habberley area are coarse grits of the Longmyndian Oakswoods group; they strike about north-south, dip steeply to the east and are cut by numerous fissures and faults. The veins or vein groups strike about east-west. The Westcott group on the south consisted, when being worked, of some five barite veins that trend between E12°S and S20°E; they dip generally to the north. The barite died out at shallow depths.

The next vein, going north, is the Huglith Main vein that trends a bit north of east and displaces the outcrops of country rock about 300 meters to the west on the south side of the fracture; it dips 22°S. The ore shoots on this vein have been as much as 750 meters in length. The vein is cut off to the east by passing into brecciated country rock mixed with clayey material;

in this mixture the walls become lost and the vein filling degenerates into irregular stringers of barite. To the west, the vein is cut off below a fault, probably the Habberley, and by a boulder clay in which it dies out.

The third and fourth veins are the Riddleswood and Mud (from south to north). The Riddlesworth dips 15°N and the Mud 12°S. The ore shoots on these veins are much less extensive than on the other two. The barite died out in the bottoms of the known ore shoots; the ore shoots tend to pitch to the west; the Main vein continues beyond the bottom of the barite as a breccia of country rock cemented by quartz, the Riddlesworth as a calcite vein.

Only the Westcott mine in the Habberley area is said to have produced some copper ore (in 1880), and the total output of the area can almost certainly never have amounted to much. The Huglith mine produced, as has been mentioned, nearly 300,000 tons of barite; the Cothercott, Gatten, and Wrenthall were responsible for from 20,000 to 10,000 tons of barite ore each.

The Breidden-Hills area is about 18 kilometers west from Shrewsbury. These hills are made up of an outcrop of Ordovician rocks about seven kilometers long from northeast to southwest and 2.4 kilometers wide. It is bounded on the north and west by alluvium of the Severn River and on the east and south by Silurian and Carboniferous beds. The Ordovician rocks were intruded by a dolerite laccolith that makes up the main Breidden Hill and were folded into a northeast-trending anticline. The conglomerates and tuffs on the southeastern flank of the anticline have been eroded into a series of hills. Both the Ordovician and Silurian rocks of the Bredden Hills have been cut by several northwest- and north-trending faults; only one mine, however, was ever developed in the Breidden Hills (the Bulthy) it is located on the county boundary. The rocks of Ordovician age are not well correlated with those of the Shelve-Habberley area, but they definitely appear to belong to the upper Ordovician; the Silurian and Carboniferous rocks are undifferentiated in the mapping of the hills. A north-south fault in Silurian shales, about 2.0 kilometers from the center of the Breidden Hill, contains mineable barite. About the center of the mineralized stretch on this fault, a branching fissure that strikes northwest and offsets the Silurian-Ordovician contact joins (or comes near to joining) the north-south fault; the latter fault does not offset the contact. The northwest fault also contains a mineable barite. Another small fault of northwest trend cuts the north-south fault farther north near the contact between conglomerate (south) and shale (north). Its mineralization has been traced for only about 100 meters. Except where stained brown by surface weathering, the barite is white; it contains a little galena and considerable calcite. From 1879 (when the mine was begun) to 1918 (when the mine was closed) it produced about 8500 tons of barite and one ton of lead concentrates.

The igneous rocks in the Shelve-Habberley area were introduced in two stages, both of which probably are of Silurian age. Both also were introduced before the mineralization. Ellis says (1978) that the ores are post the Caledonian revolution; how long after the end of the igneous activity the mineralization was introduced is geologically uncertain, but it appears from work by Ineson and Mitchell (1975) that K/Ar ages for the altered wall rock from the veins gives an age of about 355 million years. Thus, the ores probably should be classified here as Middle Paleozoic.

Some primary zoning of the ores appears to have taken place, sphalerite is the principal mineral in the deeper mines; as the veins are followed upward, lead-zinc ores become dominant, and still higher barite and calcite are the main, and almost sole, minerals. In ore veins near the surface, the lateral distribution of barite-calcite ore is much wider than that of the lead-zinc ores into which the barite-calcite grades downward and inward. The veins appear normally to be undisturbed by important later faulting nor do the veins seem to have been altered secondarily, except in the barite-calcite-secondary copper ores in Precambrian rocks. Remobilization has not been appealed to by Dines (or anyone else of whom I know) to account for any of the mineral relationships that have been observed in the ores.

As has been suggested earlier in these notes, I think there is considerable question as to the relationship of the barite-calcite-minor secondary copper ores to the lead-zinc ores west of the Habberley fault. I suspect that, if enough ever is known about the district, it will be found that the barite-calcite-minor secondary copper ores actually were formed at much greater depths and were brought much nearer the surface by upward movement on the east side of the Habberley fault than they were when they were emplaced.

The lead-zinc ores, from which most of the production of the district has come, probably contained only small amounts of silver, normally not worth recovering or reporting. For some reason or reasons, now unknown, records were kept of the silver in lead only in 1882 and 1883. The results of these assays were reported in ounces per ton of lead; whether this really meant per tons of lead or of lead concentrates is uncertain. Because of the large proportion of lead in lead sulfide (86.6% Pb), it does not greatly matter which was meant. Certainly the barite mines did not produce silver, and the lead-zinc mines for which no silver production is listed either did not operate in 1882-1883 or did not have their lead assayed for silver or the lead contained none that could be detected. The reported amounts of silver in the ores ranged from 2.23 ounces per ton of "lead" to 0.5 ounces per ton, almost certainly not enough to qualify the ores for a class higher than telethermal. Nor are any sulfosalt minerals reported that suggest that leptothermal conditions may have obtained in the district. Under what temperature-pressure range the deep sphalerite ore may have been deposited is not known. Microscopic examination of polished sections from such sphalerite might give clues through the presence or absence of exsolution blebs of chalcopyrite or other minerals.

On the basis of this rather limited evidence, which does not appear to include any work on fluid inclusions, it seems best to categorize the West Shropshire deposits as Telethermal.

Official records of production were not kept in the district until the middle of the nineteenth century, so it is difficult to say which were the most productive mines over the centuries from Roman times to the end of mining activity in 1945 (at the Huglith mine). It appears, however, that the largest amounts of lead were recovered from the Roman Gravels and Snailbeach mines and of barite from the Huglith. Although only 50,000 tons of lead ore is recorded for the Roman Gravels mine, this probably was only a small fraction of the total; perhaps the total was somewhere between 250,000 and 500,000 tons of lead concentrates. The lead concentrates mined from Snailbeach between 1945 and 1913 were about 132,000 tons; this was 56 per cent of the lead concentrates produced in Shropshire in that period. Between 1958 and 1912, Snailbeach also recovered 4400 tons of zinc (concentrates?), about 23 per cent of the total from Shropshire. Since a pig of Roman lead was found in the mine, it probably operated spasmodically from that time to 1913, so the total production of lead, at least, must have been far greater that 132,000, and it possibly may have been great enough to have equalled or surpassed Roman Gravels. If the 1882-1883 figures mean anything, the ore at Roman Gravels (2.23 ounces of silver per ton) was nearly five times as rich as that from Snailbeach (0.50 ounces of silver per ton).

As for barite, the most productive mine was Huglith with about 300,000 tons of run-of-the-mine barite ore for the entire period of operation (1910 to 1945).

MENDIP HILLS, SOMERSET

Middle Mesozoic (Primary Ore) Late Tertiary (Oxidized Ore)	Lead, Zinc	Telethermal (Primary) Ground Water-B2 (Oxidized)

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Notes

The lead-zinc deposits of the Mendip Hills, worked since the Romans ruled in Britain, are located in the County of Somerset in southwestern England (Ford, 1976). This range of hills is about 40 kilometers long and eight kilometers wide and trends generally west-northwest-east-southeast. The highest elevations reached in the district are some 300 meters; the tops of the hills are quite flat; the slopes dip steeply away from the ridge line and locally are dissected by narrow gorges such as the famous one at Cheddar. The towns of Wells (51°31'N, 2°39'W) and Cheddar (51°17'N, 2°46 W) lie along the southwest margin of the belt of hills. To the east, the Mendips merge with the south end of the Jurassic-limestone Cotswold Hills.

Although lead was mined from the days of the Romans, the zinc content was valueless to them as they did not even know of that metal. During the latter part of the 19th century and up to 1908, Roman slags were reprocessed for lead.

In the period of modern mining, both lead and zinc (mainly as smithsonite-calamine) were obtained from the ores. Barite and fluorite are so sparsely distributed that they never were mined for themselves but appear to have been recovered in places as byproducts of lead-zinc mining. At present, mining for the metals has ceased but much limestone is quarried in the district.

The dominant rock of the Mendips is limestone, that, at its maximum, has a thickness of 1050 to 11,150 meters, though normally the thickness in any one location is much less than the larger figure. In the folded rocks of the hills, the Devonian Old Red sandstone (the Portishead beds) forms the cores of the Carboniferous limestone anticlines and usually constitutes high ground. The Devonian beds are some 450 meters thick as a maximum and are fluvial-fan facies of the Old Red; they include conglomerates, sandstones, and thin marl beds. The lowermost beds of the (lower) Carboniferous limestones are a basal lower Limestone shale, which is a shale with thin limestone beds intercalated in it; it is about 150 meters thick as a maximum.

Above the Limestone shale is 900 to 1000 meters of massive limestone that is divided into five beds; these are, from bottom to top: (1) the Black Rock limestone, a fine-grained crinoidal rock 225 to 300 meters thick; (2) the Vallis limestone, a massive crinoidal limestone zero to 90 meters thick: (3) the Burlington oolite that passes laterally in part into the Vallis - the Burlington is 45 to 225 meters thick; (4) the Clifdon Down limestone that is light and dark gray calcite mudstones - it is 160 meters thick; (5) the Hotwells limestone that is a massive gray crinoidal limestone with thin shales and quartzites at the top. Above the Carboniferous limestone are the Millstone grit (below) and the Coal Measures (above). After deposition and lithification of the Carboniferous beds, the rocks were uplifted and folded and eroded until, by late Triassic time, their geomorphic forms approximated those they exhibit today. By the late-Triassic, the area was an archipelago of limestone islands that stood above the Triassic lowlands and sabkhas. Erosion on these islands developed a series of ravines and channels in the upper Triassic rocks, that became filled with a dolomitic conglomerate; this formation is known as the Dolomitic conglomerate; it is from zero to 110 meters thick and is made up of small to large boulders of Carboniferous limestone. In places, these boulders have been converted to hematite or dolomite, and all are cemented by a matrix of sand, gravel, or marl. Away from the Mendip Hills, these fans of conglomerate grade out into sandy or marly beds that merge into the upper Triassic Keuper marl. This relationship effectively dates the formation of the Dolomitic conglomerate as upper Triassic. Ford emphasizes the point that these beds, beneath the Somerset levels to the south, contain halite and gypsum deposits (that he later uses in his concept of the manner in which the Mendip Hills deposits were formed).

The Dolomitic conglomerate reached sufficient depth that it buried the Paleozoic rocks of the central Mendips; to the east; however, lower Jurassic (Liassic) beds overlap the conglomerate. These Liassic beds are normally clay but have marginal littoral limestone facies. Still farther east, middle Jurassic beds of shelly oolites overlap the older rocks; both Jurassic rock types are beyond the eastern boundary of the area that has been mineralized. The Triassic and Jurassic beds have been folded very slightly, but enough so that their contacts with the strongly folded Paleozoic beds present a most impressive unconformity. Groundwater-produced karstic features under this unconformity are either late Triassic or early Jurassic and contain sediment, and, locally, such cave sediments are mineralized.

The rocks that make up the Mendip Hills were folded into a series of anticlines (periclines) and synclines, the axes of which run about east-west and which are shifted to the east as the ore area is crossed from north to south, thereby producing an en echelon pattern. The three major periclines are (from northwest to southeast) the Blackdown, the North Hill, and the Pen Hill. The folding was completed before the earliest Mesozoic (Keuper) rocks were laid down, and, by that time, the relief in the area was much greater than now.

Green (1958) suggests that the structure of the Mendip Hills be considered by taking up the periclines in the order in which they best display the controls that govern the mineralization. The first of these is the North Hill pericline, the axis of which trends from east to west and which lies in the south-central part of the mineralized belt. The northern side of the pericline dips steeply north or is even overturned; the dips on the south side are much more gentle and are in a southerly direction. To the west, the structure pitches quite gently but, to the east, the pitch is appreciably steeper. This pericline is cut by two more or less north-south normal faults, the Stock Hill to the west that trends northwest-southeast and the Biddle that (in the North Hill area) is about north-south. In the southern portion, the Biddle fault cuts pre-Triassic rocks, whereas, farther north, it displaces rocks as young as Liassic. Green believes that the southern part of the fault is pre-Triassic and the northern at least post-Liassic. Actually, the situation may be that there are post-Carboniferous rocks in the southern part of the North Hill pericline and that these have been covered by spoil from the old mine workings or they too would show that they had been displaced by the southern portion of the Biddle fault. Thrust belts cut across the Castle of Comfort area (north of North Hill), and others are located farther south between the North Hill and Pen Hill periclines.

The central part of the North Hill pericline was deeply eroded during post-Carboniferous time, and the basin so created (in the center of which Old Red sandstone was exposed) was filled with a major thickness of Dolomitic conglomerate. This conglomerate was then overlain by Mesozoic sediments that range in age from Keuper Marl to lower Lias. These Mesozoic strata have been folded into gently elongated domes and basins, the long axes of which generally trend east-southeast.

Major breaks in the lower Limestone shale were provided by the Stock Hill fault, the eroded core of the pericline, and the southern part of the Biddle fault. The scale of mineralization in the rocks of the North Hill area, does not appear to have been influenced by the dip of the beds, except that, where they stand nearly vertically, the bedding planes may have served as channelways for ore-forming fluids.

To the east of the Stock Hill fault, the Old Red sandstone has been displaced to the southeast and now stands as Stock Hill that contains a core of Old Red sandstone. North of Stock Hill, the area known as Chewton (Rabbit) Warren, contains numerous northwest-striking veins that contain much of the mineralization of the North Hill pericline. Another large area of Old Red sandstone, known as Eaker Hill, straddles the Biddle fault. A large area of Dolomitic conglomerate lies to the west of the fault and west of the northern part of the Eaker Hill Old Red; this area of conglomerate does not appear to have been appreciably mineralized, probably because it lacks the northwest-trending fractures of Chewton Warren. Another area of ore-filled fractures in the Dolomitic conglomerate is located east of the Biddle fault in the angle between the Biddle fault and the east-west striking Emborough thrust; these fractures appear to be continuations of those of Chewton Warren.

Both Green and Ford consider that the entire area of the North Hill pericline (and of the Central Mendips in general) eventually was covered by the fine-grained early to middle Mesozoic argillaceous rocks (Keuper, Rhaetic clay, lower Liassic limestone and clay). Although the hills (North, Stock, and Eaker) probably all initially stood as islands above the level of the Mesozoic sediments, eventually they were drowned by the sediments of that era. Apparently, the low domes and shallow basins developed in the Mesozoic rocks were related to loci of ore deposition in that most of the ore lies under such domes and most areas under basins are ore-free. Thus, not only did the Mesozoic rocks act as a cover that trapped the ore-forming fluids, but these fluids were guided up into the late domal structures developed in the Mesozoic rocks and overprinted on the pre-Mesozoic ones. Thus, the ore-forming fluids took the paths of least resistance which lead them into the Paleozoic and Triassic rocks beneath the domes where they then deposited their mineral load. It appears probable to Green and Ford that the ore mineralization took place only after the end of the mid-Mesozoic folding and certainly after the deposition of the Liassic rocks. Of course, it is possible that the lack of ore in the post-Dolomitic conglomerate rocks indicates that these were not present until after mineralization was completed, but the direct locational connection between the mid-Mesozoic structures and the major ore concentrations argues against this.

For the purposes of structural discussion, Green divides the Blackdown pericline into two sections, the easterly between the Blackdown and Lamb Leer (and the Lamb Leer fault) and the westerly between Blackdown and Shipham. In the nose of the pericline pointing east toward Lamb Leer, Old Red sandstone is exposed; the structure is a simple fold pitching east at 13° to 14°. Following this anticlinal structure back (west) from the nose, the crest, so narrow to the east, broadens out and becomes a much wider, gently folded structure that has a general east-southeast pitch of 7° to 10°. The flanks of the pericline dip quite steeply away from the crest (at angles of 25° or more) as pitching monoclinal folds. Three km east of the Old Red sandstone outcrop, the monocline on the north dies out and is replaced, about 500 meters to the south, by a similar fold, near the crest of which is located the east-west Lamb Leer fault. The flanks of the central area of the Blackdown pericline plunge beneath the cover of Mesozoic rocks to the east and southeast with dips of up to 30°. The subsidiary folds on the Blackdown crest, which strike southwest at Charterhouse, become essentially east-west near Lamb Leer. The Lamb Leer appears to be a normal fault in which the downthrown side is to the south. A fault that runs southwest from the vicinity of Charterhouse appears to be a continuation of the Stock Hill fault. Both the Lamb Leer and Stock Hill faults seem to have been first formed before the Triassic but later movement may have occurred on both of them.

The faults probably exerted control over the location of the ores, as the areas of mineralized ground are never far from these structures. The erosion surface of the area of mining is remarkably flat and is at an elevation of about 250 to 275 meters and cuts across all of the Carboniferous strata and the Dolomitic conglomerate as well.

From Blackdown to Shipham, the pericline is overturned on the north but has appreciably less steep dips on its south flank. No major faults cut through the rocks. Along the anticlinal axis, its core of Old Red sandstone is exposed; this core has been subjected to major erosion and was later filled in by Dolomitic conglomerate. Around Shipham, the Dolomitic conglomerate has been considerably fractured with the fissures striking west-northwest to east-southeast and being closely spaced. Some mineralization also was located in the nearly vertical Carboniferous beds along the north flank of this portion of the Blackdown pericline with most mineralization being in the Burrington oolite. The strike of these veins is, of course, essentially east-west. As no major faults are associated with the fissures, Green believes that the ore fluids must have risen up through the appreciably porous Old Red sandstone. Because of the major amount of erosion from the anticlinal core, the total thickness of the ore zone never will be known, but the dips of the Mesozoic rocks surrounding the pericline suggest that they were appreciably domed over the missing Dolomitic conglomerate.

The Pen Hill pericline rocks are compressed into isoclinal folds that are overturned on the northern flank; this limb is bounded by thrust faults. The generally east-west structures are cut by both the Biddle and Stock Hill cross faults. Mineralization has been located along the trace of the Biddle fault. Additional mineralization is contained in the rocks of a rather corrugated area on the northwest edge of the pericline. In this area, the veins are lined up with the plunge of the folds. In other places on the pericline, the veins follow the strike of the Carboniferous rocks or are nearly parallel to the north-south cross faults. Although the area may have been blanketed by Mesozoic formations, the evidence for this is so sparse as to prevent any great confidence being placed in this conjecture.

Ford considers that the main time of folding was after the Carboniferous but before the Triassic. After the folding, the relief in the Mendips area was considerable, but, as has been described above, the relief was worn down and the ravines and channels filled with Dolomitic conglomerate and the entire area covered by a thin veneer of Keuper and Jurassic beds. Ford considers that the area must have had at least a sparse cover of Cretaceous chalk. Except for a few small remnants of these later Mesozoic formations, uplift and erosion have removed all of the unfolded post-Triassic rocks. The lack of Tertiary rocks and a lack of interest in the history of that epoch means that little is known of the geologic events in the Mendip region in Tertiary time. Tertiary groundwater activity probably produced the oxidized zinc (calamine mainly) and lead (cerussite mainly though in much less amount than the calamine) ores in the karst features. These karsts were developed in the late Triassic to early Jurassic and probably had their primary mineralization introduced at the same time as the emplacement of the lead-zinc ores not in karst features. Although other suggestions have been made, it appears probable that the primary ore was emplaced all at one time. Since the lower Liassic rocks contain some ore, it appears most likely that the age of the primary mineralization was middle Jurassic and, therefore, is here classified as middle Mesozoic. The oxidation of these karst ores and the development of gossans probably took place in late Tertiary time and these features are so categorized here.

The deposits that have been of economic interest are almost entirely concentrated in the central portion of the Mendip Hills. The ore veins are on the flanks of the anticlines, Pen Hill, North Hill and Blackdown and on the unnamed structure (possibly Stock Hill) between the later two structures. Although no plans of the mines or records of production have survived, Green (1958) calculated the lengths of veins in the various formations. These are:

Dolomitic conglomerate	42,000 m	
Carboniferous limestone	71,300 m	
Burrington oolite		33,100 m
Black Rock limestone		25,100 m
Clifton Down limestone		12,400 m
Hotwells limestone		700 m
Totals	113,300 m	71,300 m

Almost two-thirds (64%) of the ore is in the Dolomitic conglomerate and the Burrington oolite, with 37 per cent being in the Dolomitic conglomerate. The amount of mineralization rises progressively from the Black Rock limestone (22%) to the Dolomitic conglomerate. Thus, the formations almost certainly were not uniformly favorable loci for ore deposition.

On the North Hill pericline, the ore mineralization is concentrated on either side of a patch of Triassic rocks between the two cross faults (Biddle and Stock Hill). The veins in this area range in dip from subparallel to the steep north dip of the rocks or to the bedding, strongly suggesting an impressive structural control of ore localization. Some of these veins can be traced into the overlying Dolomitic conglomerate. In fact, in the Chewton Warren area, the strongest veins of the Mendip area follow west-northwest-east-southeast fractures in this conglomerate. Farther south, where the Dolomitic conglomerate unconformably overlies the Biddle fault, the former is heavily mineralized. In the North Hill pericline, therefore, the loci of ore deposition have been controlled by faulting; these faults have permitted the ore fluids to reach all the favorable beds in the Carboniferous stratigraphic section and to continue up into the Dolomitic conglomerate. In certain places, the Triassic Keuper marl and the Liassic clays have acted as a barrier to further upward movement of the ore fluids. No ore mineralization is present in the thrust faults, although Ford thinks that they locally may have acted as channelways for the ore-forming solutions.

The veins in the Blackdown pericline are concentrated in the Burrington oolite on the northern slope of that structure around the nose of that fold, in the Clifton Down limestone and in the Blackrock limestone near Charterhouse south of the eastern end of the pericline. The veins are located near to the northern end of the Stock Hill cross fault or in badly broken rocks near another cross fault (the Lamb Leer) that lies to the east of the Stock Hill. Near the Lamb Leer (three km east-southeast of Charterhouse) the trend of the fractures is east-west, but farther west they change to a northwest-southeast strike. In the vicinity of Charterhouse, the veins have two predominate directions, northwest-southeast and west-northwest-east-southeast; these veins are locally subparallel to highly mineralized faults. Some of these veins, particularly in the northeast part of the mineralized area, extend upward, without change, into the Dolomitic conglomerate. In places, the vein fillings have been highly oxidized and include calamine, manganese wad, and some secondary lead minerals. In the area of Shipham (5.5 km west-northwest of Charterhouse), in the core of the Blackdown pericline, the Dolomitic conglomerate was strongly mineralized; on the north side of the Blackdown structure, in the Burrington oolite that there stands almost vertical, numerous veins also are present. The veins extend east-west, parallel to the bedding in the limestone, but also occur in joints in the conglomerate. Ford reports that faults are essentially not exposed in the Dolomitic conglomerate beneath which, however, such may be concealed.

The Pen Hill area is isoclinally folded and this is overturned toward the north; it contains several thrust faults and is cut by both the Stock Hill and Biddle faults. Along the trace of the Biddle fault are numerous veins; these were highly mineralized and are either parallel to the bedding of the Black Rock limestone or are subparallel to these major faults. The potentially ore-bearing Dolomitic conglomerate has been eroded from the hill, but Ford reports the presence of some traces of the lower Lias clay.

The principal sulfide, now present on dumps or previously present in the ore, was galena; only minor sphalerite was to be found in the veins but much calamine was mined; it was derived, of course, from the once-abundant sphalerite. The principal gangue minerals are calcite and barite, with only minor traces of fluorite. Barite and calcite may be interbanded, and barite is most common in the center of the ore-bearing area. A considerable portion of the galena has been oxidized to such minerals as cerussite and the much rarer mendipite ($2PbOPbCl_2$) and hydrocerussite [$2PbCo_3Pb(OH)_2$]. Chloroxiphite [$2PbOPb(OH)_2CuCl_2$] and diabolite [$2Pb(OH)_2CuCl_2$] are very rare and are associated with mendipite. Crednerite ($CuOMn_2O_3$) also is rare as are mimetite [$(Pb_5AsO_4)_3Cl)$], wulfenite ($PbMoO_4$), and pyromorphite [$Pb_5(PO_4)_3Cl$]. Rhodocrosite and tetrahedrite are known. More usual minerals to be associated with telethermal deposits also are present but need not be mentioned individually. The secondary lead minerals generally are in cavities in porous masses of iron oxide, probably correctly called gossans; these gossans may contain quartz crystals as cavity linings. Galena is not present in the iron oxides. Iron and manganese oxides are, in places, intermixed to form wad.

The only fluid-inclusion study on a Mendip mineral (so far as Ford knows) was made on fluorite at a site outside the main ore area; it homogenized at 84.8°C. Green (1958) suggested that the ore area is vaguely concentrically zoned, with its center at Chewton Warren (that is due east of North Hill and north of Stock Hill), an area where barite is more common than anywhere else in the district. A few determinations of lead-isotope ratios have been made, and about all they prove is that the isotopic ratios from galenas in several formations are much the same.

Essential agreement now appears to exist that the ores of the Mendip Hills were emplaced at one time and are epigenetic to the rocks that contain them. They are typical Mississippi Valley-type ores so far as work done to now can suggest. The majority of geologists who have studied the ores agree that the ore-forming solutions rose from depth to invade the carbonate rocks of the Central Mendip Hills. Ford (1976), however, points out that no significant mineralization has been found in the Old Red sandstone or in the Silurian rocks beneath it and that the lower Limestone shales provide at least a partial barrier to the upward movement of ore fluids. Further, such geophysical studies as have been done in the area do not indicate any basement rock, granites or otherwise, that could have been a source of ore fluids in the Mid-Mesozoic. Certainly it is clear that, in Cornwall and Devon to the southwest, large quantities of ore fluids of hydrothermal character invaded large volumes of Paleozoic rocks (mainly of igneous but also of sedimentary origin) to deposit tin and associated ores. But about 100 million years must have elapsed between the deposition of the Cornish ores and those of the Mendip Hills. Dunham has suggested (1952) that pools of hydrothermal fluids may have been trapped at depth beneath the Mendip-Hills area and only escaped in the Mid-Mesozoic. This explanation seems rather farfetched and has been discarded by Ford. Ford himself suggests a radically different hypothesis.

Ford (1976) suggests that the formation of the ores of the Central Mendip Hills be explained by what he calls a "mechanism of basin margin migration and precipitation". He explains this as follows. The Carboniferous limestones of Mendips were deposited in a shallow sea, the coast line of which sea was some 50 to 60 kilometers to the north. He interprets the pattern of isopachs and facies changes to suggest deeper water basins to the east, south, and west during the lower Carboniferous. Ford believes that the Carboniferous limestones were buried by the thick sediments of the Upper Carboniferous, Permo-Triassic, and lower Jurassic to depths of two kilometers or more. He points out that the main Carboniferous geosyncline lay to the south of the present location of the Mendips. Concealed thrust faulting has forced the sedimentary rocks developed in this geosyncline closer to the Mendips by at least 15 kilometers, and the complex folding undergone by these rocks has reduced their length by a least 20 per cent. Ford conceives that compaction and compression in these thrusted and shortened beds forced out formation waters toward the basin margins to the north, the moving waters making use of the network of faults that had been produced during the tectonic activity of the late Carboniferous. Ford conceives of this migration of these waters toward the basin margin as taking place in early Jurassic times, long after the fracturing had taken place. Precipitation from these formation waters occurred in the "structural culminations" of the Mendip folds where they were covered by the Triassic Dolomitic conglomerate, all of these rocks being then capped by late Triassic and early Jurassic clays. Ford believes that some contribution to the formation waters (and ultimately to the ores) may have come from beneath the Somerset coal field to the north of the Mendips, but he does not consider this contribution important to the ore volume. Ford sees no reason, although little work has been done on the problem, to assume that the ore-forming fluids of the Mendip ores were different in any major respect from those that obtained in typical Mississippi Valley-type ore fluids. Ford's hypothesis has the advantage that it derives the ore-forming fluid from a process that is known to have taken place - compaction of the basin sediments, but he tells nothing about how the chemistry of the process worked. The lead and zinc, fluorine and barium ions, and sulfur in some form or other may have been available in the formations being compacted, but how the ore fluids dissolved the various components found in the Mendip ore, how they concentrated them sufficiently to reprecipitate them, and how the sulfide ions were formed, all are matters about which Ford says nothing. He does refer to a paper by Wolf in a Volume edited by Chillingarian and Wolf, but it does not answer the problem in detail sufficient to fit the concept to the production of the ores of the Mendips.

What does appear to be certain, however, is that the ore fluids rose from some depths beneath the surface, probably coming from the south, carrying the constituents of the Mendip ores, and depositing them in favorable loci in the carbonate rocks of late Paleozoic and early Mesozoic age. These events appear to have taken place under conditions of low temperature and confining pressure from solutions that cannot have been highly acid or they would have attacked the carbonate formations violently (which they did not). This indicates that is would not be unreasonable to categorize the ore-forming process as telethermal (in the broad sense). The secondary oxidation of the ores almost certainly was due to the action of groundwater. Apparently, iron sulfides, pyrite and marcasite, are almost unknown in the district. What iron was present in the primary ore well may have been there as one or both of the iron sulfides, but surface oxidation in late Tertiary time probably resulted not only in oxidation of the iron sulfides but also in the transportation of much of the iron into karst cavities where it accumulated in considerable amounts. The iron-oxide-bearing materials in the caverns cover a cavity lining of quartz, calcite, and barite. It would seem from the relationship that the iron-oxide material probably was the last material to deposit, but this does not mean that the parent iron sulfides were among the last materials to deposit, rather they may have been precipitated at much the same time as the ore sulfides but were much more readily affected by the ground water that reached the cavities at any appreciable time after the sulfides and the crystalline gangue minerals had been emplaced. Where the manganese associated with the iron oxides came from is uncertain; no traces of primary manganese minerals have been found nor is manganese normally an appreciable constitute of iron sulfides. The source of the manganese, therefore, can be added to the source of the ore-forming fluids as the two major mysteries of the Mendip ores.

LAKE DISTRICT, CUMBERLAND, WESTMORELAND

Middle Paleozoic	Lead, Zinc, Silver Iron as Hematite Tungsten, Copper	Hypothermal-2 (Fe as Hem) Hypothermal-1 - Mesothermal (Cu, Zn) (W) Mesothermal to Leptothermal (Pb, Zn, Bar, Ag)

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Notes

The Lake district is made up of some 2000 km^2 in northwestern England and lies mainly in the counties of Cumberland and Westmoreland. The Lake district has three separate areas of economically valuable mineralization, in all of which mining has now ceased. The last ore mined came from the Greenside mine located at the head of Ullswater; production ended there in 1967. The only proved reserves of ore in the district are in the Carrock tungsten mine that is nearly 32 km west of Penrith. At the Threlkeld mine, near Keswick, some reserves of zinc have been indicated. The total production from the district is, of course, unknown. Although mining was carried out in the district at least as far back as the 16th century, almost nothing is known as to the amounts of ore produced. Between 1850 and 1947, production from the Lake district mines amounted to 200,000 to 300,000 tons of lead concentrates (75%-80% Pb), and about 40,000 tons of zinc concentrates (about 45% Zn).

Because of differences in mineralization within the district, it is divided for discussion here into three parts. The first of these, Caldbeck Fells, is north of a line through Cockermouth and Penrith (about 54°40'N), the second (the Vale of Newlands) encompasses the area 13 kilometers in radius and centered around the town of Keswick, the third (Coniston) consists of a few deposits, randomly scattered in the southern part of the outcrop area of the Skiddaw Slate series and throughout the outcrop area of the Borrowdale volcanic series that is south of the contact line between the Skiddaw slates (north) and the Borrowdale volcanics (south).

The roughly circular area of Ordovician to Devonian rocks that make up the stratigraphic sequence of the Lake district are surrounded by a rim of Carboniferous limestones on all sides except that to the west where Permo-Carboniferous overlaps the Carboniferous beds and rests directly on the older rocks. The oldest rocks known in the Paleozoic sequence are the lower Ordovician Skiddaw slates, psammites and pelites; their contact with the basement never has been seen, so their thickness has not been ascertained. The Skiddaw group contains the Skiddaw slate below and the Eycott Volcanic group above; both are Llavirn-Arenig in age. The Skiddaw rocks are overlain unconformably by the 8000 m+ thick Borrowdale Volcanic group that is pre-Caradoc, post-Llanvirn in age; the Borrowdale rocks are mainly lavas and pyroclastics but include minor intrusions (see below). These Ordovician beds are highly folded and faulted, and ore deposition is confined to them except in the Carrock Fell area where tungsten has been sporadically recovered from the Carrock mine. The ore of the Carrock mine is spatially associated with a silicic intrusive that is pre-Carboniferous and possibly as young as early Devonian.

Above the Borrowdale are the Coniston Limestone group, 25 meters of limestones and shales of Caradoc age and the black shale beds that have various names in different parts of the district and average about 15 meters in thickness; they are of Ashgill age. The Silurian beds that overlie the Ashgill shales are the Llandoverian Skelgill/Bowgill beds of siltstone and shale. They are followed by the Wenlockian Brathy flags, that is, siltstones, and the Ludlovian that contains, in various part of the district, the Coniston grits, the Bannisdale slates, and Kirby Moor flags. The total thickness of the Silurian sequence is more than 5000 meters. The Devonian began with intrusions of the major silicic intrusives, the Caledonian granites, and these are overlain unconformably by varied thicknesses of Upper Mell Fell conglomerates. The overlying Carboniferous rocks include, from the base up, the Basement group of red sandstones and conglomerates, the Limestone series and the Millstone grits that have different names in different parts of the district, and the Coal Measures; the total thickness of Carboniferous rocks is about 1000 meters. Finally, the Permo-Triassic consists of more than 300 meters of red beds (predominantly sandstones, plus evaporites) that lie unconformably on the Carboniferous beds. Over another unconformity are Pleistocene and Recent beds, peat, alluvium, glacial moraines, and boulder clay (Ineson, 1976).

The Caledonian orogeny produced the main structural features of the district. The Skiddaw slates were folded, the folds having southwest and west-southwest trends. The Borrowdale volcanics have a similar trend. It may be possible to divide the Caledonian structural episodes into three parts: pre-Borrowdale volcanics, pre-Caradon, and post-Silurian phases. The first two phases have varied trends that generally are northerly. The last has elements of the former trends to which have been added northeast to east-northeast folding. During the Armorican orogeny, the Caledonian wrench faults were reactivated, and new structures were imprinted on the older ones thereby adding north-south and northeast-southwest elements to those already present. Even Alpine structures are evidenced by a north-south tilt to the area and reactivation of the faults of the area.

The Caledonian (Devonian) earth movements appear to have begun after the deposition of the Ludlovian Kirby flags. No Downtonian rocks are known in the district, nor does it contain Old Red sandstone unless some small patches of conglomerate near the border of the area may be of that formation (Mitchell, 1956). The major folds and fractures in the district were produced at this time. Considerable disagreement remains as to how these complex structures are to be interpreted, but, in general, it is agreed that the Skiddaw slates were sharply folded and crumpled along a generally southwestern or west-southwestern trend. The Borrowdale formation appears (Green, 1920) to have been separated from the Skiddaw by a series of faults; the folding of the Borrowdale probably was more broad, although the same trend-directions were followed. The Geological Survey mapped a major syncline through Sea Fell and an anticline to the southeast through Langdale.

Early workers from Cambridge, of whom Harker was one, suggested that a deep-seated thrust plane lay beneath the Lake district, being inclined upward toward the north. They characterized the faults between the Skiddaw and the Borrowdale as lag faults as they did the faults between the Borrowdale and Coniston formations. Green, in an earlier work, took the view point that the Skiddaw and Borrowdale pass gradually from one to the other but that there was an unconformity at the base of the Coniston. He also postulated a major anticlinal structure, with overfolds, in the Borrowdale and Skiddaw rocks and, to a lesser extent, found these structures in the Silurian rocks as well. A great deal of work has been done since Green expressed the opinions just quoted and many more details of the structure have been worked out, but this work has not changed the general concept of highly folded areas of Skiddaw and Borrowdale rocks throughout the district. Locally axial trends departed from the southwest trends, but these divergences do not change the general picture.

Mitchell (1956) considers that the Devonian earth movements were severe, and the pressures causing them probably came from the south. This Devonian folding was accompanied by much faulting, and the planes of the strike faults seem almost to coincide with the rock bedding planes. Against these near-bedding-plane faults, most of the tear faults end. Some of the strike faults are of major dimensions as is true of the great strike fault that cuts the Stockdale slates and against which dip- (tear-)faults end. A major strike fault in the Stockdale slates at Kentmere is a thrust, although its dip is from 45° to 90°S; it is paralleled to the south by many minor thrusts. All the thrusts have a west-southwest trend. Locally this great Kentmere thrust is offset by north-trending faults that may displace the Kentmere fault by as much as 400 meters. The tear faults might be later than the Devonian, but they probably were a part of the earth movements of that time. The much more complex folding in the Skiddaw probably is due to its being much less competent than the Borrowdale, which is much less drastically folded and faulted, but some complex folding does affect the Borrowdale. In the folding, the greatest relief of pressure probably was in the area of the Ordovician Ashgill shales and the Silurian Llandoverian shales immediately above them. In this rock sequence a great strike fault is located, either a thrust or a lag and probably compound with some blocks moving faster than others and thus being thrust over or lagging behind those immediately adjacent ot them. Mitchell believes that the final effect was packing with sliding along or close to the bedding planes. This compaction seems to have continued into the younger Silurian rocks, which are, however, less strongly folded.

The differentiation between true dip and cleavage dip appears to be quite difficult in the district; usually both features have essentially the same strike, but locally may diverge appreciably. In places, the axial planes of the folds and of the slaty cleavage are almost parallel. As cleavage is now known in the youngest Silurian beds present in the district, it is probable that the area was covered by even later Silurian rocks and by Devonian Old Red sandstone, rocks that have by now been completely eroded away.

After the cleavage had been effected, tension was set up along to two sets of axes, one directed west-southwest and the other north-northwest. The first stage of tension produced the low-angle-of-dip joints trending west-southwest and north-northwest; into these two sets dolerite dikes were introduced. Later tension formed high-angle-of-dip joints. Joints of these west-southwest and north-northwest trends are common in the Carboniferous of northern England, so may also have been developed or subjected to further tension in the late Paleozoic earth movements, but they probably were already in existence by the time of the Carboniferous earth movements.

The intrusive rocks of the district include three major bodies of granitic rocks, the Skiddaw granite, a small mass intrusive into the northeastern portion of the Skiddaw slates, The Eskdale granite along most of the western margin of the outcrop area of the Borrowdale volcanics, and the Shap granite, also a small body, at the southeast corner of the outcrop area of the Borrowdale volcanics. Near the western end of the contact between the Skiddaw slates (north) and the Borrowdale volcanics (south) is the Ennerdale granophyre; it has an outcrop area about one-half the area of that of the Eskdale granite, the largest of the igneous bodies. The Carrock Fell complex lies along the eastern part of the contact between the Borrowdale volcanics (north) and the Skiddaw slates (south); it has two to three times the outcrop area of the nearly neighboring Skiddaw granite. So far as geological relations go, only the Shap granite shows any definitive evidence of its stratigraphical age. It has metamorphosed the Silurian sediments (Coniston limestone and Ashgillian shales) that lie immediately south of it. Further, Carboniferous beds to the east of the Shap granite contain feldspar fragments that must have been derived from the Shap intrusion. All of the granites are late or even post-tectonic intrusions in an area that shows strong cleavage and folding but is not thermally metamorphosed. For the Shap granite (Brown and others, 1964) the mean age of determinations (by the Rb/Sr and K/Ar methods) is 393 m.y. For the Skiddaw granite, the mean age for the determinations (all by the K/Ar method) is 399 m.y. For the Eskdale granite, the mean of the age determinations (all by the K/Ar method) is 383 m.y. All these results suggest that the intrusions were rather early in the Devonian.

Brown and others (1964) report that both the Shap and the Skiddaw granites are surrounded by wide aureoles of thermal metamorphism. Apparently these aureoles indicate that the intrusions were emplaced under thick rock cover. This supposed deep-seated character of these two granites is confirmed by the absence of recognizable fragments of the granite or of fragments from the associated aureoles in the Mell Fell conglomerate, probably of Old Red sandstone age.

Brown and others (1964) apparently believe that the arguments are convincing in favor of the Lake district granites having risen isostetically at a similar rate under progressively shallower conditions due to erosion, but the Eskdale, being somewhat younger than the other two granites rose toward the surface under a lesser thickness of overlying rocks, and, this inference is given confirmation through the generally discordant character of the igneous (Eskdale) rock-intruded rock contact and through the appreciable presence of intrusion breccias, both of which indicate a more brittle environment for emplacement than obtained for the Skiddaw and Shap intrusions. Mitchell sees some direct relationship between the causes for igneous intrusion and for the late-orogenic isostatic uplift, an uplift that is not confined to the Lake district but extends throughout the Caledonides of Britain. What that causal connection may have been, Mitchell does not attempt to say.

Because the other two major intrusive rock masses in the Lake district have been highly altered, the Ennerdale granophyre and the Carrock Fell complex are not completely suitable to have their age determined by the K/Ar method. Results of whole-rock analyses by this method, however, should be reasonably accurate because not much of the argon escapes entirely from the rock volume in which it is generated, but, in large part, at least, remains in the whole rock, and a large enough sample will provide fairly satisfactory age determinations. The age determined by this method for the Carrock Fell complex is 356±20 m.y.; geologically, however, the Carrock Fell rock is older than the Skiddaw granite at 399±6 m.y. Thus, the age obtained by this method for the Carrock Fell complex can be regarded only as an absolute minimum, but it does show that the two rocks (Carrock Fell and Skiddaw) are both Devonian in age.

The mafic variety of the Ennerdale granophyre gives an age, by the whole rock K/Ar method, of 334±18 m.y.; again this must be looked on as a minimum age but shows that this rock also is early Devonian. The silicic type of the Ennerdale gives and age of 370±20 m.y., and the argon was extracted from perthite that Mitchell says definitely loses argon over a span of geologic time. This means that the Ennerdale is probably older than the date just given, but even so, the age still probably places it in the time frame of the Caledonian revolution.

The Shap granite normally is a pink rock with prominent pink feldspars. The entire intrusive mass is made up of several increasingly silicic rocks. These include lamprophyres and quartz felsites that are intruded not only into the Ordovician rocks but also into the Silurian ones as well.

The Eskdale granite is also a pink granite that contains many biotite-rich clots. It has been suggested that the Eskdale was derived from (granitization of) the volcanic country rocks. It does metamorphose the Borrowdale volcanics and a pebble of this rock was found in a pebble in a core of rock from the Millstone grit.

The Ennerdale granophyre is stock-like in form and consists of several types, the most common of which is a pinkish, medium- to fine-grained rock made up of quartz and feldspar; the ferro-magnesian mineral (or minerals) that it contained are now chloritic material.

The Carrock Fell complex contains gabbro, granophyre, diabase, and felsite. These rocks occur as a group of steeply dipping inclined lenses or sheets that parallel the east-west to east by south strikes of the host Borrowdale volcanics and Skiddaw slates.

The Skiddaw granite is a white rock with conspicuous black mica; it has three outcrops, two in Caldew Valley and a third in Sinen Gill. The extent of the metamorphic aureole that surrounds it indicates that it actually occupies a much larger area than that of the outcrops. Many dikes of this rock cut through the aureole, further suggesting that the concealed mass is much larger than the outcrops. The Skiddaw slates were folded in the Devonian and then were first intruded by dikes. The slates later were metamorphosed by the intruding granite, the intrusion having been preceded by the development of cleavage and by thrust faulting. Thus, the granite cannot be older than the Devonian earth movements. In addition to the granite, the Skiddaw intrusion also includes, in order, aplite, pegmatite, and hydrothermal action that produced both greisen and hydrothermal mineralization.

Other less-studied and smaller-sized intrusive masses also are known in the Lake district, including the Threlkeld microgranite, Haweswater dolerite, the Armboth granophyre dike, the Little Knott mafic picrite, and the Sale Fell minette.

The suggestion has been made that the extensive and major alteration of the Borrowdale volcanics may have been caused by a major intrusion underlying the Lake district. The dome-like uplift of the area in the Tertiary also may have been due to reactivation (?) of the intrusive activity that caused the alteration of the Borrowdale rocks.

At one time or another over the last four hundred years at least an appreciable amount of mining has been done in the Lake district. Probably quite accurate figures are available for production, beginning in 1848. Before that date, records are essentially non-existent, but the pre-1848 production cannot have been impressive by modern standards at least. The known amounts of lead and zinc mined and concentrated after 1850 (1848?) are given on the first page of these notes. In addition, the lead pigs probably contained about 10 ounces of silver per ton. Assuming that the total lead production was the 186,000 tons of about 80 per cent Pb concentrates were produced between 1950 and 1947 (as given by Dunham in 1976) and that this is an underestimate by 20 per cent, allowing for lead produced before and after 1850 and 1947, respectively, the total tonnage in terms of lead pigs would have been about 225,000 and the total silver, recoverable, if not recovered, would have been about 2,250,000 ounces, a respectable amount by any standards.

The only tungsten mine of any importance in the district is the Carrock (Fell) mine, and production figures from it are not available (to me, at least). Production, however, was sufficient that "good returns" are reported to have been made in the latter half of World War I and promising exploration was done during World War II, but the tungsten shortage eased sufficiently that work was stopped before the exploration program was completed; further, reported exploration and preparation for mining appeared in 1976 to be progressing favorably. No information has been found in my work as to the total production of copper from the Lake district, but, for the Coniston mine, the tonnage is reported to have totalled 52,000 tons of cobbed or dressed ore that ran from five to 13 per cent copper, at most 5500 to 6000 tons of recoverable copper metal. If any other mineral material was recovered from the Coniston mine, it has not been reported in any publication I have seen.

The mineralization in lead, zinc, and silver is contained in fissure veins and replacement ore bodies, and these are loosely concentrated in three areas, all within short distances of Keswick. Northeast of Keswick, in the Caldbeck Fells region, are the mines of Roughtongill, Sandbed, Driggith (the last two are on the same vein), Potts Ghyll (3.0 kilometers northeast of Roughtongill), Red Gill (one kilometer west-northwest of Roughtongill), and Drygill (one kilometer east of Roughtongill and perhaps on the same vein as Driggith). All of these mines, except Potts Ghyll and Ruthwaite are lead, copper, zinc, barite, and probably silver mines; the zinc might now be recovered from the dumps, but the sphalerite probably is so oxidized by now as not to be recoverable by flotation. Silver probably was recovered but I have seen no records that this was done. These mines are all within about three kilometers of Carrock Fell mine, but they appear not to be genetically related to the complex. The Roughtongill mine (apparently composed of two more of less parallel veins) was worked at least from the time of Queen Elizabeth I, mainly for copper. At the southwestern end of the two veins, the more easterly vein began in the Carrock Fell complex and then passed into the Borrowdale volcanics, and then on into the Carboniferous limestones where it may not have been mineralized and where the structure may have been opened after the mineralization was complete to the southwest. The more westerly vein appears to have begun at the contact between the complex and the Borrowdale volcanics; it joined the easterly vein southwest of the point at which the easterly vein enters the Carboniferous limestones. Driggith and Sandbed are reported to be on the same vein; if this is so, the Sandbed section has been displaced to the east by a more or less east-west fault by nearly one kilometer. For both the Driggith and Sandbed mines, the production was appreciable, for example, between 1850 and 1870, 5000 tons of lead concentrates were removed from the Driggith mine. The Sandbed appears to be confined to the Borrowdale rocks and the Driggith to the rocks of the complex. Potts Ghyll in 1956 was the only producing mine in the area, and it produced only barite in amounts worth mentioning.

Two major mines near, and southeast of Keswick are the Greenside mine, at the head of Ullswater and the Threlkeld mine about eight kilometers directly south of the center of the Carrock Fell complex and 1.5 kilometers southeast of Keswick. The Greenside mine was in intermittent operation for well over 200 years and produced more than 200,000 tons of lead concentrate and 65 tons of silver before its final closing in 1967.

The Greenside mine is located almost entirely in Borrowdale fine-grained andesites and some tuffs; in the northern part of the mine, at depth, some Skiddaw slates are known. The Threlkeld mine is contained in large part in dark Skiddaw slates but enters the widespread metamorphic aureole of the Skiddaw granite, first encountering chiastolite slate and farther north cordierite slate.

The veins in the Vale of Newlands area (west and southwest of Keswick) include the mines of Force Crag, Thornthwaite, Barrow, Stoneycroft, Yewthwaite, Brandlehow, Goldscope, and Castle Nook. The country rock of these veins is Skiddaw slate.

In the Coniston area, the ores are contained in east-west and west-northwest veins that cut the generally northeast-southwest-trending beds of the Borrowdale volcanics. Although overlain by Silurian beds, the mineralization is confined essentially to the Ordovician rocks.

The replacement bodies of hematite ores in the Carboniferous limestones of south and west Cumberland almost certainly are the youngest ores in the district. The major amounts of ore are in strata-bound deposits in various beds of the limestone sequence. These major replacement masses are normally linked by vein-filling deposits in fault systems. In west Cumberland, also present are veins filling northwest- or west-northwest-trending faults; they contain quartz, dolomite, chalcopyrite, and galena. In south Cumberland, the fault pattern is more complex and consists of minor northwest and east-west faults and major north-south and north-northwest cross faults; the minerals in these veins are hematite, quartz, and calcite. These iron-ore veins obviously are quite different than those of the other areas of the Lake district, and the ore fluids that deposited them almost certainly came from a different source than those responsible for the non-ferrous ore veins of the bulk of the Lake district. Further, they can be no older than Carboniferous, and their ore-forming fluids must have had a source appreciably different from that that produced the copper-zinc-lead ores of Lakeland. The reported depositional temperatures for the Cumberland iron ores (Shepherd, and others, 1976) are based on studies of the fluorite present in them. On this fluorite basis, it has been suggested that the filling of inclusions in fluorite at 115°C indicates that the hematite and other minerals that are associated with the hematite were formed in the vicinity of 115°C. Normally these hematite associated minerals are considered to indicate much higher temperatures of formation than 115°C, and this still may be the case. It appears probable to me that the fluorite is late in the paragenetic cycle and that actually does not give a true picture of the temperature at which the hematite ores were formed. More evidence is needed if the hematite ores are to be considered the result of low-temperature formative processes.

The major structure of Lakeland (Eastwood, 1959) is an anticline with the early Ordovician Skiddaw slates (the oldest rocks) being bordered on either side by the less old Ordovician Borrowdale volcanics; these rocks are overlain by those of Silurian age (although these do not outcrop along the northern rim of the district, being covered there entirely by Carboniferous beds). The anticlinal axis runs through the center of the Skiddaw outcrop area in an east-northeast direction. This axis curves to the south at both its ends. On the flanks of this major fold are several lesser, though large, folds and a huge number of small ones that are cut by numerous faults, in some of which lie the various ores of the Lake district. All of the rocks show some cleavage, but the Skiddaw slates are better described as shales, flags, and grits, and they resemble the killas of Cornwall to a considerable extent.

Although there are five large intrusions in Lakeland and many small ones, ore veins cut only the Skiddaw granite and the Carrock complex, both near neighbors at the north end of the district. Despite the agreement on the Devonian age of the intrustions, the veins penetrate, though barely, Carboniferous rocks. This suggests rather than a Carboniferous age for all the ores that further igneous activity at depth produced small quantities of Carboniferous ore-forming fluids that followed the paths of their Devonian forebearers. Thus, the age of the ores is here given as Middle Paleozoic with the understanding that this probably includes earliest Carboniferous to a minor extent. The late Tertiary uplift of the Lake district probably did not produce any ore mineralization even though it did affect the topography and the degree to which the ore veins were exposed at the surface.

The north-south veins of the district show an upward zoning with chalcopyrite and sphalerite at depth, then with galena and sphalerite, followed above by barite and dolomite, and finally by pyrite and manganese minerals. Normally, chalcopyrite and galena do overlap, but not to a large extent. There does not appear, however (Eastwood, 1959), to be any lateral zoning as is true in the Pennines. The tungsten ores are definitely older than the lead-zinc ores, but they still are Middle Paleozoic in age. On the contrary, the copper veins normally are older than the lead-zinc but can be later in places; the same is true of quartz, a common gangue mineral of the copper ores. Most of the copper ores are in east-west veins, and the north-south faults that contain the lead-zinc ores have displaced the east-west veins but this rule has many exceptions.

The wall rocks, except where soft, compact ones barred the upward passage of the ore fluids, apparently did not localize the ore mineralization except through the structures that earth movements developed in them. The important structures are the relatively minor fractures and not the major folds or even the minor ones. The ore bodies are in many places about equal in vertical and horizontal dimensions, and, in a number of instances the vertical dimension is greater than the horizontal, as at Greenside and Coniston. Eastwood (1959) suggests that several ore bodies, not completely mined out, probably have a greater vertical extent if the lowest ore mined was galena and sphalerite. Thus chalcopyrite ores might be found at depth. If this concept is correct, then, although there may have been a copper-rich and a lead-rich mineralization, the one following the other, lead-zinc mineralization at depth also may be expected to pass into a copper-zinc one. From the common occurrence in the Coniston area of ore bodies that do not have an upper extension into lead-zinc ores and are in faults of different strike than those of the lead-zinc veins that are known to grade downward into copper-zinc phases, it would seem to follow that the first ore fluids that were given off from the magma chambers at depth were capable of producing chalcopyrite-arsenopyrite-pyrite ores (at least at the levels at which such ores are now known), whereas the later north-south faults tapped magma chambers that produced ore fluids that ran the gamut in the ores they produced from copper-zinc to pyrite-manganese ones. This would explain why lead-zinc ores that have been followed to sufficient depths grade into copper-zinc ores whereas those that outcrop in the copper-arsenopyrite range may have had an upward extension into lead-zinc ores but have long since lost these to processes of erosion.

Although most of the copper veins (with minor or no lead-zinc) are irregularly distributed in the Borrowdale volcanics and show no preferred formations within the volcanics, a few are known to be in the Skiddaw slates. One vein was explored for copper in a southern, xenolithic portion of the Eskdale granite, but no economic mineralization appears to have been found.

In the Coniston area, the tuffs and rhyolites and andesites to a minor extent are the host rocks for the ore mineralization. The rocks strike nearly northeast-southwest and dip uniformly to the southeast at high angles. The faults are most abundant in the rhyolite, which probably accounts for such rocks being the most productive in the area. The faults appear to have resulted from a pair of conjugate shear planes at 345° and 80°, plus a set of tension fractures at 300°; these orientations are consistent with compression at right angles to the strike of the beds. The strike faults are high-angle thrusts that are localized in the softer horizons and probably were concomitant with the folding. Some of the faults displace the north-south mineral veins. Open spaces were created along the veins by the refraction of the shear planes as they passed from one rock type to another. Of all fault types, the strike faults are least favorable to mineralization because they rarely intersect bedding planes and, for this reason, refraction-produced open space normally is not developed on them.

Wall-rock alteration is confined to the veins and to narrow zones next to them. Such alteration produces quartz and chlorite, both of which also are common in the unmineralized volcanics. Further, these two minerals are the main gangue minerals in the veins and are accompanied by only minor quantities of carbonate and feldspar; barite and fluorite are completely absent from these deposits. The order of deposition of the primary ore minerals is: magnetite, arsenopyrite, pyrrhotite, pyrite, marcasite, chalcopyrite, sphalerite, galena, bismuth, and bismuthinite. All of these minerals, except pyrite, occur only in the veins; pyrite may impregnate the wall rock, particularly the tuffs. Dagger (1977) considers that the textures of the ore minerals are consistent with a single phase of mineralization; replacement textures are only minor in extent. Some early arsenopyrite and pyrite have been broken and then in-filled by quartz, chalcopyrite, and sphalerite, indicating some small-scale movement along the veins during mineralization. Locally carbonate minerals were deposited and these seem to post-date all minerals.

The veins show vertical zoning in that pyrite and chalcopyrite dominate near the surface, arsenopyrite is added to these with further depth, and magnetite is present instead of arsenopyrite at still greater depths. This magnetite phase is found only in the Bonser mine at Coniston, which is the deepest mine in the area. The magnetite may have been a replacement of even earlier hematite as is suggested by the lamellar habit of the magnetite crystals.

The Carrock Fell mine is on the northern edge of the Skiddaw granite and is about seven kilometers south of the village of Caldbeck (Dunham, 1976). During the late stages of the crystallization of the granite, a small greisen zone was developed in which several quartz veins and stringers were developed. These extend from Caldew Valley on the south to Drygill on the north. In about the middle of this set of quartz veins, the veins travel upward not only through the greisenized Skiddaw granite but on into the mica schist of the metamorphic aureole of the granite into the Carrock gabbro of the Carrock complex. In the entire vertical extent of the veins, tungsten minerals are present. The veins are vertical or nearly so; these veins are cut by minor east-west fissures that contain uneconomic amounts of galena and sphalerite. The three main veins are the Emmerson (east), the Harding (180 meters to the west), and the Smith (another 45 meters west of the Harding). The veins are from 10 to 15 centimeters to 1.5 meters wide and contain mainly quartz with randomly located wolframite and scheelite, followed by arsenopyrite, pyrite, pyrrhotite, iron-rich sphalerite, and ankerite as the principal minerals. Eastwood (1959) says that the ankerite has reacted with wolframite to provide additional scheelite. Other minerals locally present are galena, molybdenite, chalcopyrite, bismuth, tourmaline, apatite, dolomite, calcite, and barite. Some assays have been reported to contain tin, but no tin mineral ever has been identified. Wolframite normally occurs near the vein walls and quartz mostly, but not solely, in the central portions of the veins. In places, the greisen is said to have been heavily kaolinized. At least on the Emerson vein, the throw on the fault has been considerable. Dunham states that the mineralization was emplaced about 390 m.y. ago.

Although Dagger (1977) thinks that the as-yet unverified granite at depth provided only heat that heated water in the overlying rocks enough that it was able to extract ore elements from the Borrowdale volcanics and transport them upward to be deposited higher in those rocks, it seems much more probable to me that the ore fluids came from the hidden granite and moved upward from far below the basal Borrowdale rocks to deposit their load in them, with some spilling up into the Skiddaw beds.

From this discussion of the copper deposits of the Coniston area, it would appear reasonable to assume that they were deposited at high enough temperatures to be classified as Hypothermal-1 to Mesothermal, although microscopic study of the Coniston chalcopyrite and sphalerite might determine definitely if mesothermal should be included.

The considerable amounts of silver in the lead of the lead-zinc ores in the Lake district suggests that they were formed at higher temperatures than telethermal and that they should be categorized as Leptothermal or, perhaps, better as Mesothermal to Leptothermal. The lack of detailed publications on the mineralogy of the lead-zinc ores of Lakeland prevents a more positive statement as to classification to be made here.

Despite the short distance that lies between the Lake district and the Pennine ore bodies, it appears probable that there was an appreciable difference in time and character between the Lakeland ores and those of the Pennines. The Lakeland ores are almost certainly Devonian (with perhaps a minor addition to the lead-zinc ores in the Carbono-Permian) whereas the Derbyshire ores are most likely early Mesozoic and those of the Askrigg and Alston blocks are more probably early Mesozoic. The Lakeland ores range in wide variety from Hypothermal-1 and -2 to Leptothermal, whereas those of the Pennines are true Mississippi-Valley-type telethermal ores. These classifications remain correct no matter where the ore fluids came from or how they managed to transport and deposit their mineral load. It seems reasonable to suppose that the close spatial relationship of the Lakeland ores to granitic igneous rocks and the essential lack of such in the Pennines must be considered in any discussion of their respective manners of genesis.

SOUTHWESTERN ENGLAND LEAD AND ZINC

Late Paleozoic	Lead, Zinc, Silver Barite, Fluorite	Leptothermal to Telethermal (minor)

Dewey, H., 1921, Lead, silver-lead and zinc ore in Cornwall, Devon and Somerset: Geol. Surv. Great Britain Spec. Repts. Mineral Res., v. 21, 70 p., particularly p. 10-59

Dines, H. G., 1956, The metalliferous mining region of south-west England: Dept. of Sci. and Ind. Res., Geol. Surv. Great Britain Memoirs, 2 vols., 795 p. (mainly descriptions of individual mines but includes most of those that were mined exclusively for lead, zinc, silver, and barite)

Schnellmann, G. A. and Scott, B., 1970, Lead-zinc mining areas of Great Britain, in Jones, M. J., Editor, Mining and petroleum geology: Ninth Min. and Met. Cong. 1969, Pr., v. 2, p. 325-356; (disc., p. 357-365)

Vipan, P. G. L., 1959, Lead and zinc mining in south-west England, in Future of non-ferrous mining in Great Britain and Ireland: Inst. Min. and Met. Symp., London, p. 337-353

Notes

The positions of the lead-zinc lodes in southwestern England already have been located in the discussion under the heading: "British Lead-Zinc". Schnellmann and Scott report that the area produced nearly 320,000 tons of lead (probably lead concentrates) and almost 90,000 tons of zinc concentrates. The largest amount of production came in 1845 through 1886. Although the area contained, at one time or another, more than 200 mines, 71 per cent of all reported lead production came from nine mines and 63 per cent of all zinc from three. Zinc was produced from 63 mines. More than seven million ounces of silver are reported to have been obtained from the lead.

Normally, the veins have a north-south trend with the faults being normal and the dips steep to vertical. The ore shoots in these structures mainly are fissure fillings, but some replacement of the immediate wall rocks has been reported. Some of the ore shoots, at least, were emplaced in soft shale and ended when the fracture passed into harder, more siliceous rock. The ore shoots were up to 800 meters along strike and extend to at least 400 meters below the surface. The veins are located in Devonian and Culm Measures (lower Carboniferous) beds. They usually are spatially related to granites, and the mineable mineralization is found both within and beyond the envelopes of thermal metamorphism that surround the granites. In addition to galena and sphalerite, the ores include barite (200,000 tons produced from the district), fluorite (5000 tons recovered), quartz, and calcite. For what the results of age determinations made from lead-isotope studies are worth, these deposits appear to have been formed in the late Carboniferous or early Permian and are associated in space with granites that appear to be of about the same age. For an outlier of the tin-copper deposits of Cornwall, the assignment of this age appears to be reasonable.

Studies of the wallrock alteration (killas) are far from thorough, but what has been done indicates that the sediments containing the ores have been converted to silica and chlorite for a few meters into the walls.

The rich silver content of most of these southwestern England ores suggests strongly that they were precipitated, and in considerable part, under leptothermal conditions, but the lack of uniformity in the silver content of the galena may indicate that part of the ore was deposited in the telethermal range. This is, of course, exactly what would be expected of ores that lie on the margins of a huge district in which major deposits of hypothermal and mesothermal ore minerals are centrally located in relation to the lead-zinc-silver ores.

The first structural event in the area appears to have been the development of normally east-northeast trend folds, cleavage, and jointing. The Paleozoic rocks so affected appear to have been shales and mudstones with intercalated grits and conglomerates that have been converted by the intruded granite in the early Permian into "killas". The granites occur as bosses, usually of considerable size, that are surrounded by their metamorphic halos of killas. The areal distribution of the granite increases with depth so that no point on the surface probably is very far removed from granite at depth. Mafic dikes, sills, and lavas are common in the region but appear to have no genetic connection with the ores. In contrast to the tin and copper ores of Cornwall, which are mainly in east-west lodes, the lead-zinc ore shoots are in north-south "crosscourses", both due to the effects of the granite intrusions. All of the lead-zinc ores are either in killas or the unaltered mudstones and shales that lie outside it.

What follows here is largely summarized from Vipan (1959) and indicates that the Southwestern England lead and zinc district, although it is known only to have produced about 1/16th of the total obtained in Britain, would have been considered of far greater importance had it been by itself and not a minor offshoot of the Cornish tin-copper district.

In the Perranzabuloe-Newlyn East area, the country rock is made up solely of lower Devonian sandstones, grits, conglomerates, and shales in which the ore veins run northeast-southwest (with the exception of the East Wheal Rose mine where the two main veins run north-south) and dip steeply. In these veins, the ore shoots were normally as much as two meters wide but locally much greater widths were known. In some mines, the killas country rock was quite weak and mine openings had to be strongly supported. In places, the ore shoots were well developed in the soft shales but pinched out where the country rock changed in both directions to hard siliceous material. More than one space of soft shale on a given fissure might have been converted to ore shoots; the East Wheal Rose and the Chiverton mines occupied such a relationship. In this district, 115,000 tons of lead concentrates contained more than 1,750,000 ounces of silver; 1.2 million ounces came from the West Chiverton mine alone.

The Perran Iron Lode was poor in lead, although much iron ore was recovered from it, and this area produced over 37 per cent of all zinc recovered in the Southwest England district. The Perran Lode proper had a length of nearly six kilometers; it dipped to the southeast at, for the district, a very low angle: 35°-65°. From the usual width of a few meters, the width locally rose to over 20 meters, although such wide volumes contained large horses of barren country rock. Each wall of the vein was made up of a wide band of brecciated killas, cemented by quartz, iron oxide, and/or sphalerite. The bodies of sphalerite and iron ore were not continuous but would stop against barren killas or against fractures that cut northeast-southwest across the vein.

The Menheniot area produced about 28 per cent of the lead from the whole district and over 40 per cent of the silver; only a minor amount of sphalerite was recovered from the entire area. Of the three major mines, the Mary Ann was the most productive, having an output of nearly 30,000 tons of lead concentrates and 1.125 million ounces of silver. The Trelawny and Herodsfoot mines produced, respectively, 19,000 and over 23,000 tons of lead concentrates; and, respectively, nearly 700,000 and just over 600,000 ounces of silver. The country-rock killas contained minor lavas and mafic intrusions; two major veins extended north-south through the area, one for five kilometers and the other for over seven kilometers. The Mary Ann and Trelawny mines adjoined each other on the same vein, being worked for a total strike length of more than 1.5 kilometers and perhaps to a depth of as much as 550 meters. Below about 370 meters, the ore was far less continuous than the 80 per cent of total recovery stoped out above that depth. The combined vein was 0.3 to 1.2 meters wide; it dipped steeply east in the Trelawny mine but much less so near the southern boundary of the Mary Ann. The vein at the Herodsfoot mine was up to 1.2 meters wide and extended along strike for about 1500 meters and went down during its final period of mining to a depth of 400 meters. When the mine was reopened in 1844, the shafts were found to be open only down to 100 meters, and what was produced from these upper workings never was recorded; the total production from this mine, therefore, was appreciably larger than the 23,000 tons mentioned above. All mining in the area ended in 1884.

The Tamar Valley area produced only eight per cent of the lead from the district and 8.6 per cent of its silver; no sphalerite was reported but 3240 tons of fluorite was sold. Some mines in the area were worked as far back as the 13th century, but, of course, no production records obtained from those times, so the percentages just given certainly are low, but by how much will never be known. The country rock is Devonian killas with some outliers of the Culm Measures being known at the north end of the district. Two veins ran north-south and dipped east in the area of the Tamar River. The greatest depth of workings was about 450 meters. In the richest mine (South Tamar), silver averaged 37 ounces per ton, but this mine was so close to the Tamar that it was flooded out after only a short life. A small tonnage of fluorite (about 800) was produced from the district and sold in 1879 through 1882.

Although the Teign Valley area was the smallest producer in the district it did contribute 370,000 tons of barite, plus 9.1 per cent of the lead concentrates but only 5.1 per cent of the silver and 1.8 per cent of the sphalerite. At the surface of this area, the country rock is Culm Measures shales, mudstones, cherts, and tuffs with appreciable intrusive bodies of igneous rocks. Near the margins of the Dartmoor granite, the killas is metamorphosed. Parallel to the western border of this granite, outside the metamorphic aureole, is a belt of north-south trending fractures some 8000 meters long. Three lead mines were worked on two main veins and several minor branches; these structures dipped steeply but Vipan (1959) does not say in which direction. Production, after recording began, amounted to about 26,500 tons of lead and 1564 tons of zinc concentrates, the lead containing nearly 370,000 ounces of silver. Since one of these mines had been opened a considerable time before production statistics were kept, the totals for lead and silver almost certainly were at least appreciably greater than the figures quoted here. Barite and fluorite undoubtedly were present in these mines in considerable quantity but probably were not recovered or, if recovered, not concentrated and sold. At the north end of the fracture belt, the most successful barite mine (the Bridford) in southwestern England was found; apparently, the mine was opened about 1875 and continued in production until 1958. During that period, it is indicated that upwards of 500,000 tons of barite were recovered. In the Bridford mine, two main veins were exploited as well as a number of minor branches; the strike length was nearly 250 meters and the depth reached was about 180 meters. Between the 120 meter level and the 250 meter level, the concentration of barite per foot of depth dropped from 870 tons of barite concentrates to about 180 tons. This major decrease was caused not only by the falling-off of the dimensions of the ore shoots but also to an appreciable increase in the content of sulfides of lead, zinc, and iron. East of the main barite lodes, a stockwork area was developed but not stoped. This ore was not mined, it seems, not because of a decrease in barite grade but due to the increased sulfide content that could not be satisfactorily separated in the mine's gravity ore dressing plant. It would seem almost certain that considerable sulfide ore exists in the Bridford mine, but obviously it did not appear, to the operators, of sufficiently large size to warrant the investment in a processing plant capable of separating barite from sulfides and sulfides from each other.

At the time that Vipan wrote (1959), the value of silver was low, and he thought that this, as much as anything else, was what kept the owners of various lead-zinc-barite-fluorite properties from investing in modern processing plants. So far, even the much increased price of silver, about $10.00 per ounce at present (1982) prices, does not appear to have tempted any group into reopening or reprospecting known veins in the Southwest England district.

If the Bridford property, on further exploration, were found to pass from essentially massive barite into a polymetallic mine at depth, it would not be the first to do so, but it appears that the current chances of making a profit by such exploration do not seem good to those who have detailed knowledge of the area.

It is almost certain that the deposition of the Southwest England lead and zinc ores was carried out, at least in part, at temperatures above the telethermal range. Although the ores do not carry to sulfosalts that are typical of most leptothermal deposits in the world, the high content of silver (over 22.5 ounces to a ton of lead concentrates) places the district firmly in the leptothermal category, although minor amounts of the galena may have been deposited in the telethermal range.

Further, the age of the mineralization definitely is that of the Cornish tin-copper deposits of which these lead and zinc ores are the outer and lower-temperature fringe, and, therefore, they should be dated as Late Paleozoic.

SCOTLAND

Leadhills-Wanlockhead, Lanarkshire, Dumfrieshire

Late Paleozoic	Lead, Zinc, Silver	Leptothermal to Telethermal

Gallagher, M. J., 1978, Scotland, in Bowie, S. H. U., and others, Editors, Mineral Deposits of Europe, Volume I, Northwest Europe: Inst. Min. and Mineral. Soc., London, p. 270-273

Mackay, R. E., 1959, The Leadhills-Wanlockhead mining district, in The Future of Non-Ferrous Mining in Great Britain and Ireland: Inst. Min. and Met., London, p. 49-64

Mitchell, J., 1919, The Wanlockhead lead mines: Min. Mag., v. 21, p. 11-20 (mainly historical and technical)

Temple, A. K., 1954, Rammelsbergite from the southern uplands of Scotland: Mineral. Mag., v. 30, p. 541-543

____ 1959, The Leadhills-Wanlockhead lead and zinc deposits: Roy. Soc. Edinburgh Tr., v. 63, p. 85-113

Watson, J. A., 1937, Crawford Muir; its mining history and minerals: Edinburgh Geol. Soc., v. 13, p. 98-105

Wilson, G. V., 1921, The lead, zinc, copper, and nickel ores of Scotland: Geol. Surv. Great Britain Spec. Repts. Mineral Res., v. 17, 159 p., particularly p. 9-43

Notes

The mining district of Leadhills-Wanlockhead is located in the Southern Uplands of Scotland on the lower slopes of the northern flank of the Lowther Hills. This area is between the Nith Valley and the head of the Clyde Valley.

The two villages for which the mines are named lie on either side of the boundary between the two counties, Leadhills in Lanarkshire (northeast) and Wanlockhead in Dumfrieshire (southwest). Although the first recorded production was in 1239, some of the veins probably were worked by the Romans for lead and almost certainly for silver; zinc of course was not recovered by the Romans. In the 16th century, gold was mined from massive quartz veins that contained pyrite, muscovite, and albite as well; production of gold in this century was worth over £100,000 but has long since ceased. Since 1637, mining has been carried out in the district for lead and silver, but it was not until 1871-1880 that the first commercial production of zinc was recorded. Zinc production appears to have been confined to the Wanlockhead portion of the field, and silver was appreciably more important in that area as well. No silver was reported as mined from Wanlockhead after 1935 and from Leadhills after 1920. Except for one year (1934), Wanlockhead was shut down from 1929 until 1956-1957. The mines were put on care and maintenance in 1958 and have remained so to the present time so far as I know. The total production (estimated) of the district from 1785 through 1957 was about 315,500 tons of lead (in concentrate form?), about 14,000 tons of zinc (as concentrates?), and nearly 850,000 ounces (31 metric tons) of silver as a by-product of lead smelting. The policital division between the mines also was true in the physical and ownership senses. The lowest workings reached to about 425 meters below the surface at Wanlockhead and about 445 meters below the surface at Leadhills.

The oldest rocks in the district are of the Ordovician Arenigian stage and consist of volcanic and intrusive rocks (older) and radiolarian cherts (younger). Unconformably above the Arenigian rocks are those of the Caradocian stage that include graywackes, shales, and grits (older) and black shales passing into graywackes, shales, and conglomerates (younger). Apparently conformably above these rocks are those of the Ordovician Ashgillian stage that are designated as the lower Shales with conglomerates and graywackes. This sequence, given by Pringle (the South of Scotland: British Regional Geology, 2nd ed., 1948), differs considerably from that given by Wilson (the Lead, Zinc, Copper and Nickel Ores of Scotland: Mineral Resources of Gt. Britain, Spec. Rep., 1921) that was based on the studies of Peach and Horne in the 1890s. From Wilson's work, it appears that the thickness of the post-Arenigian rocks was slightly over 315 meters; the radiolarian cherts had a thickness of more or less 20 meters, and the thickness of the volcanic and intrusive rocks is unknown as their base never has been seen.

The rocks that contain the ore veins are graywackes, and these appear to be stratigraphically above the series of black shales and cherts. The only cherts noted in Pringle's stratigraphic sequence are the Arenigian Radiolarian cherts. From this it follows that the Black shales are Caradocian, and it would seem that the ore-vein-containing graywackes are Ashgillian. The Arenigian and Caradocian rocks are limited to two northeast-southwest-trending anticlinorial belts, one to the northwest of Leadhills and the other to the southeast of Wanlockhead. The synclinorial areas between these two belts contain the younger ore-vein-bearing Caradocian graywackes; the Caradocian beds are composed of graywackes and shales and probably are slightly over 300 meters thick.

The ore-bearing graywackes range from coarse breccias to fine shaly graywackes. The major component of these rocks is angular to subangular grains of quartz; also present is some feldspar that in many places is partly decomposed. The graywackes also include fragments of different rock types, particularly graywacke, slate, quartzite, and fine-grained igneous rocks. Hornblende also is present, as are white mica, chlorite, garnet, and zircon. The cement of these various constituents is a dark chlorite with appreciable amounts of cryptocrystalline silica; the presence of carbonate is shown by the noticeable degree of effervescence exhibited by the graywackes.

Although much of the rock of the Scottish Uplands is isoclinally folded, the rocks of this ore district do not appear to have been so affected. Actually, in the ore area, all the contacts that were mapped underground in the examination of the property from 1952 on were considered to be tectonic and the structures generally were demonstrably reverse faults. The surface mapping above the adit in which this mapping was done showed the structures to be isoclinal folds. MacKay (1959) believes that the underground mapping should be considered as the better interpretation. MacKay believes that the faults of the area can be assigned to one of two distinct sets; one trends northwest-southeast and the other generally north-south.

A geological feature in the district of considerable importance is a reverse fault with a low angle of dip (30°-40°) to the northwest. This fault is certainly known only underground, never having been seen on the surface; its relationship to the ore veins is not stated. This fault marks the northwest margin of the ore-bearing zone, a barrier broken only by a few of the ore veins.

Temple (1959) expresses the problem somewhat differently, saying that, "although the universal strike of the country rock is northeast-southwest, the dominant structural feature within the graywacke belt is the marked north-northwest-south-southeast jointing". In places, however, Temple sees two sets of joints that may depart from each other by as much as 40°. He says that 75 per cent of the joints dip to the east. The main trend of the mineral veins in the field is N15°W (or north-northwest-south-southeast), and they dip to the east at various angles, most of which are less than 30°. In the Leadhills area, there is a minor set of veins that strike west-northwest and dip southwest. The north-northwest-trending veins in the Wanlockhead area (and probably over the entire ore district) have directions that change abruptly. Where these veins make such changes in direction, minor veins take off from the main ones. Where they leave the east side of a main vein, these minor fractures head toward the north; where they leave the west side of a vein, they head generally south. Although these minor veins are mineralized only in a few places, they may form branch veins or loop back to the main vein. Slickensides are common on the walls of the main veins; such structures normally plunge north. Locally, veins have been seen to displace microdiorite dikes to the left.

The veins in the graywackes are filled with brecciated fragments of that rock and fragments that are cemented gangue minerals, mainly ankerite and calcite. Sulfide minerals also make part of this filling, but the ore minerals normally are on the foot or hanging walls of the veins with the footwalls being the more common loci of their deposition. Clay gouge in most places outlines the margins of the veins and, in narrow veins, may partly provide the cement for the fragments. A sharp bend in a vein in many places changes its character from a thin clay band to a much wider structure that may contain appreciable galena. Some veins are displaced by as much as three meters by strike faults; the example given was in a right-hand direction.

All the veins in the Wanlockhead area (to the west of the Belton Grain Vein that lies on the northeast edge of this portion of the district) end against a low-angle thrust fault that dips to the northwest. This fault is the southern margin of the northern anticlinorium and separates cherts and black shales to the northwest from graywackes to the southeast. The plane of movement of this fault, at least locally, is a wide shatter zone of crushed rock that has been laminated by pressure and contains inclusions of graywacke. Although this fault has not been traced into the Leadhills area, the way in which the Leadhills veins end against the southern margin of the northern anticlinorium indicates that the structure along this margin probably is the same, or a similar, fault although farther east the fault probably has been displaced northwest by a left-handed movement along one or more north-northwest-trending faults. This suggestion appears to be confirmed by the northwestward displacement of the Leadhills chert outcrops in relation to those in the Wanlockhead area.

The left-handed displacement of the low-angle thrust fault just mentioned has been essentially confirmed by the left-handed displacement for over 65 meters of a microdiorite dike by the Middle Grain vein at the extreme northeast edge of the Leadhills area.

The shatter zone of the low-angle thrust fault, which carries Llandeilian and Arenigian rocks southeastward over those of Caradocian age, probably furnished an impermeable or semipermeable barrier to the movement of the ore fluids farther to the northwest. This appears to be the only area in the district where Llandeilian beds are known.

MacKay points out that it has been the custom to divide the veins of the district into those of Leadhills and those of Wanlockhead. Although this division is based on political boundaries and not on geology, some differences exist between the veins in the two areas. Most of the Wanlockhead veins are sub-parallel, and all of them, with one exception, dip to the slightly north of east. In Leadhills country, however, the pattern is less regular with two major directions of strike being discernible, north-south and northwest-southeast. The dips of the veins are usually southwest for the northwest-southeast veins and east for the north-south veins. If the two strikes are combined in one vein, the zig-zag pattern that results, plus minor horizontal displacements (both left-handed and right-handed), produces appreciable volumes of open space in which the principal ore shoots were emplaced. The amount of horizontal movement ranged between 1.5 meters and 30 meters. The size of the ore shoots depended on the amount of horizontal displacement and the degree to which the vein in question zig-zagged. For example, the major ore shoot on the New Glencrieff vein reached from the surface to a depth of about 370 meters and had a length down the plunge of nearly 500 meters. The length of the ore shoot in a horizontal direction was, as a maximum, 370 meters but was not consistently of this length. Ore shoots of much smaller horizontal dimensions (45 meters) were common, and these usually were separated by twice this distance of poorly mineralized vein. Smaller patches of ore have been found, developed around even smaller volumes of open space, and such also have been mined.

The total lengths of the various veins ranged between a few hundred and over 2000 meters with all of them being contained in faults.

MacKay reports that the faulting and fracturing in the mine was studied in great detail by geologists of Siamese Tin and Rio Tinto, and he gives their findings as follows in their order of occurrence:

(1) intense shearing, which locally converted bands of graywacke to a soft shale-like rock;

(2) low-angle faults (probably slightly later and minor relatives of the major thrust fault) may have displaced item (1) shearing;

(3) barren quartz veins, in numerous places showing no movement but in others displacing the bands of intense shearing (1);

(4) steep faults with both vertical and horizontal components of movement that usually were of small magnitude;

(5) mineralized veins, most of which, or perhaps all, had considerable movement in the planes of the veins; and

(6) post-mineral faulting.

Although these six items of tectonic movement listed above are given in a definite order, it appears probable (MacKay, 1959) that they generally are not neatly separated in time but two, or sometimes three, of them were contemporaneous in some portions of the various mines.

A considerable quantity of igneous material (microdiorite or microgranodiorite) as dikes intrudes the district; these dikes are Caledonian in age. Near the mineralized veins, the dike material is altered, and these dike rocks are locally found as components of the vein breccia.

In addition to the micro-rocks, dikes of monchiquite trend northwest-southeast and outcrop to the east of Leadhills. Further, a member of the Tertiary Mull dike system cuts across the northeast corner of the mineralized district. What relationship (probably none) that these dikes have to the process in mineralization is unknown.

Subsequent to the Caledonian folding and faulting, the intersecting joint system was imposed on the area; the acute bisectrix of the joint system is north-northwest-south-southeast; this is the principal discordant feature in the mineralized area and probably was caused by a north-northwest-south-southeast compression. (Temple, 1959). The presence of Caledonian dike material in north-northwest fractures indicates that these structures that received the dike material were produced by the Caledonian orogeny. Although Temple holds that the northwest-southeast Caledonian compression would have developed the north-northwest-striking tear faults, the joints that intersect them can only have been formed by a change in the Caledonian stress from northwest-southeast to north-northwest-south-southeast. He thinks that further change toward a north-south stress would have produced a left-handed movement on those joints that trended most nearly north-south. The effect of such movement on the veins formed by the intersecting joints would have been the production of open space on the more northwest portions of the veins, such formation of open space being more pronounced in the graywackes (competent) than in the black shales (incompetent) and the cherts (splintery). This development of open space by left-handed movement provided the loci for mineral deposition; galena is concentrated in the more northwest-trending portions of the veins for this reason.

Such later movement occurred along these fractures prior to the introduction of the minerals of the deposits and before the Hercynian dikes were emplaced. Temple believes that these various movements occurred over so long a period of time that the mineralization must be dated as Hercynian or Late Paleozoic.

Although the early mineralization in the Leadhills-Wanlockhead veins does not contain galena and sphalerite, it is widespread and abundant and consists of pyrite, muscovite, albite, and gold. The quartz is massive and white and is present as a joint filling material in the graywackes. In the lead-zinc veins, these early minerals occur as breccia fragments or are slickensided on the walls of the veins. The broken quartz in many places contains lead-zinc mineral veins. This gold mineralization certainly is appreciably earlier than the lead-zinc deposition.

In most of the larger quartz veins of the district, quartz is abundant and pyrite and muscovite are common; the albite is much more localized, apparently being present only in the Lowther Hills area where it is associated with white mica. The only known mining for gold was in a vein outcropping in the Long Cleuch and that was carried out in the 16th century. Alluvial gold, however, is found throughout the district with the most favored localities being in the Long Cleuch and the Windgate Burn. In the area of Windgate Burn, the pockets containing the gold bear no geomorphic relationship to the stream as it exists at present, and these probably were produced in meanders of a pre-glacial stream.

No gold has been seen in spatial or temporal association with the lead-zinc mineralization. One quartz vein is known to contain small veinlets of galena but no other minerals, ore or gangue, of the lead-zinc ores. Whether this galena is genetically related to the quartz or to the main ore mineralization is uncertain.

The presence of white mica, which is a diagnostic mineral of the minor Caledonian intrusive rocks, argues that the gold-quartz mineralization may have been produced during some phase of the Caledonian orogeny.

The country rock that borders the lead-zinc veins originally was a dark green graywacke that has been converted to a light gray rock in which pyrite is a prominent constituent. The pyrite, which generally is euhedral, contains residual fragments of the graywacke and almost certainly was emplaced by replacement of that rock. In addition to the pyrite, the replacement process also appears to have introduced the abundant ankerite, also largely in euhedral form (rhombs). Chlorite is common in the altered graywacke and occurs in brown crystals or massive green form. Sericite is present only in small amounts. Temple (1959) interprets the paragenetic sequence as being pyrite, chlorite, ankerite, and quartz, the last mineral being, of course, largely deposited in the veins. Just what was added to the graywacke from the ore fluid to produce these replacement minerals is uncertain; definitely, much of the material now present was in the original wall rock, but iron at least has been shifted from the dark chlorite to pyrite and the calcereous matrix of the graywacke has largely been removed.

Of the ore minerals, galena is the most widespread and occurs as massive bands mainly deposited in the hanging- and footwalls of the ore shoots formed in the open parts of the veins. The less abundant, but still important, sphalerite was deposited at much the same position in the paragenetic sequence as the galena, that is, after the ankerite and the first generation of iron sulfide. These two minerals precede the calcite and barite and, in places, followed these two gangue minerals to produce a second generation of lead and zinc sulfides. Chalcopyrite, though less abundant than the lead and zinc sulfides, is widespread: it is earlier than the lead-zinc minerals but later than the ankerite and the first generation of iron sulfide. A minor second generation of chalcopyrite is associated with the second generation of lead and zinc sulfides. In addition to the early massive pyrite, that mineral is present in well-formed crystals that are later than the first generation of barite and calcite and accompany the second-generation lead and zinc sulfides. Marcasite is not abundant; it occurs with ankerite, pyrite, chalcopyrite, and quartz and not with the second generation or pyrite. Rare niccolite and rammelsbergite are later than the first generation and earlier than the second galena. Cobaltite is even more rare but probably is paragenetically associated with the two nickel arsenides. The ankerite contains various proportions of Mg, Fe, Mn, and Ca ions. Calcite is abundant in connection with all primary vein minerals, being both earlier and later than barite. Rare, but primary, minerals in the ore include plumbocalcite (maximum of $PbCO_3$ ranges from 9.5% to 26.0%), aragonite, and witherite. Quartz is the last mineral to have formed.

Temple is puzzled by the repetition of the sulfide sequence for the second time. He prefers to explain this by assuming that the later replacement reactions, the replacement of the sulfides by non-metallic gangue minerals, freed enough of the constituents of the sulfides that the solutions that accomplished these replacements eventually were so concentrated in lead and zinc and sulfide ions that these sulfides began again to precipitate. When the sulfides of this second generation were deposited, again largely by replacement, the metasomatic reactions transferred enough calcium and carbonate ions to the ore fluid to permit the somewhat later development of a second generation of calcite. The first generation of barite, however, appears to have been insoluble enough that it was unaffected by the replacing reactions that emplaced the second generation of sulfides.

I do not understand why Temple says that "the greatest replacement was effected by the last mineral in the paragenetic sequence, quartz". He goes on to say that "the late introduction of quartz is considered to have been the main factor in the formation of the second generation of sulphides". How the deposition of quartz, which may have been emplaced by the replacement of the ore sulfides, can have aided in the formation of the second generation sulfides when these sulfides had been emplaced before the deposition of quartz began is not clear. Whatever sulfide material was removed when quartz was metasomatically emplaced, it does not appear to have left any trace behend in the higher levels reached by the ore fluids unless, of course, the sulfide material so removed from the solid state was deposited at elevations above the present level of erosion. Yet, if this is the reason for the lack of a third generation of sulfides, Temple can only guess that such had been formed during the mineralization cycle and cannot say that the late introduction of quartz was a main factor in the formation of the second generation of ore sulfides. He surmounts this problem by suggesting that the ions of the replaced minerals moved upward faster than the silica of the quartz and, therefore, were, on the upper levels, deposited before the slower moving quartz - rather an unlikely explanation. Nor does he show this in his paragenetic diagram.

Secondary processes have affected the upper parts of the veins. Most of the common oxidized minerals of lead and zinc and several of those of copper are known here, but a considerable number of rare and unusual ones are known. These include leadhillite, lanarkite, susannite, and caledonite, all of which are sulfates of lead or of lead and copper. Vanadinite and jamesonite also have been reported from the district as have pyromorphite, plumbogummite, cotunnite, lead hydroxyapatite, beudantite, and phoenicochroite. Copper minerals due to secondary enrichment are covellite and chalcocite. These two sulfides are paragenetically earlier than the first of the oxidized minerals of lead, anglesite; this was followed by cerussite or hydrocerussite. Most of the additional oxidized minerals of lead were derived from the replacement of cerussite and were pyromorphite and polyspharite (Ca-rich pyromorphite); the end stages of the paragenesis are in places vanadinite and the desclosite group. Another sequence of oxidized minerals stemming from cerussite, but less common than that just given, was leadhillite [$Pb_4(SO_4)(CO_3)_2(OH)_2$] and lanarkite [$Pb_2(SO_4)O$]. Where mixed galena-chalcopyrite ore was oxidized, the first mineral, anglesite, may have been directly replaced by caledonite [$Pb_5Cu_2(SO_4)_3(CO_3)(OH)_6$] and linarite [$PbCu(SO_4)(OH)_2$]. Other minerals also included in this paragenesis derived from lead-copper ore were, after cerussite, malachite-linarite, leadhillite-caledonite, hydrocerussite-malachite and caledonite-anglesite. The most common arrangement of alteration minerals, however, is anglesite-cerussite, followed by pyromorphite [$Pb_5(Pb_4)_3Cl$] or by leadhillite-lanarkite.

Where chalcopyrite is present in considerable masses, it normally is replaced by chalcocite and then by goethite. Malachite may follow goethite, then come brochantite, linarite, and cerussite. Only three oxidized zinc minerals are known in the district - hemimorphite, aurichalcite, and hydrozincite. Greenockite is locally present on sphalerite. In places, the oxidized zone extends nearly 400 meters below the surfacce. This depth is not related to the present water table; it probably related to one that existed when the relief was greater and the water table lower.

Temple considers that, if the ores in the Leadhills-Wanlockhead district were deposited from solutions derived from magmatic activity, they cannot have been related to any of the Caledonian orogeny but must be discussed in relation to activity of Hercynian time. The only Hercynian igneous activity associated with an orogenic belt in the British Isles is that in the southwest of England, particularly Devon and Cornwall. In the remainder of the country, the non-orogenic part during Hercynian time, alkaline and calc-alkaline volcanic types predominate. Temple thinks that his evidence indicates that the mineralization in the Leadhills-Wanlockhead district was later than both the alkaline and calc-alkaline suites of the Hercynian volcanic association. Certainly, it is unlikely that the mineralization can be thought to have accompanied Tertiary volcanic activity as the igneous rocks (in the broad sense) of that time are notably poor in associated ore deposits. Temple argues that, if these deposits were produced concomitantly with igneous activity associated with the Hercynian orogeny, the ore fluids must have been derived from high in the tholeiitic layer or from the granite crust. He believes that the elements contained in the Leadhills-Wanlockhead ore indicate a magmatic source for the ore fluids must have been developed in a locus including the top of the tholeiitic layer and the bottom of the granite crust.

The lateral and vertical zoning of the ores and their essential confinement to the vein structures indicate that the ore fluids derived from magmatic centers and not from meteoric fluids that were heated at depth and rose through, and scavenged from, the sedimentary rocks of the district. The more or less metamorphosed sedimentary sequence in the district does not appear to have contained sufficient zinc and lead, particularly lead, to have provided enough of the ore metals to the ore fluids to have permitted the development of the ore shoots in the veins of the district.

The Leadhills-Wanlockhead deposits are not typical of Mississippi-Valley-type deposits as are those of the Pennines and the Mendip Hills and are closer in character to the deposits of the Lake Country. The presence of rather considerable amounts of silver in the ores appears to put them above the telethermal range as the solution of silver in galena depends on the temperature being above that of the inversion of argentite to acanthite. The structure of argentite permits its ready acceptance in galena; the much different structure of acanthite does not. As this inversion temperature is approximately 173°C, the deposition of the galena above, or appreciably above, this temperature puts these deposits out of the Mississippi-Valley class. On the contrary, the lack of any exsolution textures in chalcopyrite and sphalerite indicates that these minerals were not formed in the hypothermal range. The only sulfosalt reported is rare jamesonite, so the leptothermal range is not certainly to be included in the classification of the deposits. Nevertheless, it seems better to include it in the categorization of the Leadhills-Wanlockhead ores, and they are here characterized as Mesothermal to Leptothermal.

WALES

Mid-Wales Ore Field, Cardiganshire, Montgomeryshire*

Late Paleozoic	Lead, Zinc, Silver minor Copper, Gold Barite, Pyrites	Mesothermal to Leptothermal

Ball, T. K. and Nutt, M. J. C., 1976, Preliminary mineral reconnaissance of central Wales: Inst. Geol. Scis. Rept. no. 75/14, 12 p.

Hughes, W. J., 1959, The non-ferrous mining possibilities of central Wales, in The Future of Non-Ferrous Mining in Great Britain and Ireland: Inst. Min. and Met. Symposium, London. p. 277-294

Jones, O. T., 1922, Lead and zinc. The mining district of north Cardiganshire and west Montgomeryshire: Geol. Surv. Great Britain Spec. Repts., Mineral Resources, v. 20, 205 p.

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* The old names of the shires of Wales are used here as most of the descriptions of the geology and the details of ore characteristics are based on these older geographical designations.

Nutt, M. J. C., 1978, Central Wales, in Bowie, S. H. U., and others, Editors, Mineral Deposits of Europe, Volume I, Northwest Europe: Inst. Min. and Met. and Mineral. Soc., London, p. 277-279

Phillips, W. J., 1972, Hydraulic fracturing and mineralization: Geol. Soc. London Jour., v. 128, p. 337-359

Raybould, J. G., 1973, Framboidal pyrite associated with lead-zinc mineralisation in mid-Wales: Lithos, v. 6, p. 175-181

____ 1974, Ore textures, paragenesis and zoning in the lead-zinc veins of mid-Wales: Inst. Min. and Met. Tr., v. 83, Sec. B (Bull. no. 816), p. B112-B119; disc. 1975, v. 84 Sec. B (Bull. no. 822) p. B67-B69; (Bull. 828)., p. B154-B155

____ 1976, The influence of pre-existing planes of weakness on the location of vein-type ore deposits: Econ. Geol., v. 71, p. 636-641

Notes

Mining in the lead-zinc region of Mid-Wales may have begun in pre-Roman times, although the evidence for this is not completely convincing. Certainly, gold was sought and found by the Romans during their occupation of Britain. After, at best, scattered instances of mining over about 800 years, work of some importance was carried out sporadically from the 12th century until about the middle of the 19th when operations in Mid-Wales reached a peak that extended between 1840 and 1880. Between 1840 and 1920, some 10 to 12 per cent of the lead and zinc produced in Britain came from the ore deposits of Mid-Wales. The greatest lead production was in the 1840 to 1880 period when the annual production of lead concentrates ran between 8100 and 16,250 tons; for zinc concentrates annual production between 1870 and 1908 was between 2000 and 5100 tons. Total production after 1845 (when official records began to be kept) was 510,000 tons of lead concentrates, which contained at least 110,000 kilograms of silver, and over 150,000 tons of zinc concentrates. In addition, some 7000 tons of copper concentrates, about 2350 troy ounces of gold (73.1 kilograms), 4100 tons of pyrite, and 4400 tons of barium minerals were recovered. Cadmium was present in the sphalerite, antimony in the galena, and silver in the sphalerite, but none of these appears to have been recovered in refining or, if they were, to have been credited to the mining companies. In this area, about 275 mines are known to have been in operation at one time or another, but only 50 of these provided 95 per cent of the total output. Most mines were closed by 1920, and the last did so in 1942 (Ellis, in Dunham and others, 1978).

All of the ore (except gold ore) was mined from lodes in the various vein systems with the only real exception being in the Van mine, east of the Cylwedog Dome, where flats, associated with the veins, also were worked. The gold that was mined, however, came from only one mine, the Oogofau (or Roman Deep), where it was recovered in part from quartz veins and in part from gold-bearing pyrite impregnated into wall-rock shales. The system of lead-zinc veins was best developed on and around the Plynlimon and Cylwedog domes with the veins on the western limbs of these anticlines being appreciably better mineralized than those on the east. Where the strike of a lode-vein was deflected as it crossed from one limb to the other of an anticline, ore shoots on the one side were of higher grade than on the other. Ore shoots containing galena and sphalerite are most common in the portions of the Van, Frongoch, and Cwmystwyth formations where they consist of interbedded sandstones and siltstones (turbidites).

The ore are localized on east-northeast-trending normal faults and breccia zones that cut across the Caledonian folds that have a north-northeast regional strike.

Most of the outcrops in the district are of Silurian beds, but the Ordovician Van formation occurs as inliers in the centers of the principal domes or anticlines. The stratigraphy (modified by Nutt, 1978, after Jones, 1922) is as follows:

Silurian (top to bottom)	
Upper Llandovery*	Cwmystwyth formation, 1000 meters of alternating sandstones and mudstones Frongoch formation, 910 meters of alternating sandstones and mudstones between pale mudstones, and all topped by dark pyritic mudstones
Lower Llandovery	Gwestyn formation, 76 meters of dark, pyritic mudstones
Ordovician	
Ashgill	Van formation, 1050 meters of pale mudstones above alternations of sandstones and mudstones

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* the lowest Frongoch and uppermost Gwestyn are middle Llandovery

The north-northeast-trending folds in these rocks produced many periclinal structures that (paradoxically) are stated to be elongated along the north-northeast trend. These structures do not persist along strike but are arranged in an en echelon fashion to each other. Although reverse faults follow the trends of the prominent fold structures, the east-northeast-striking normal faults are the economically most important fractures in the district, and these cut obliquely across the folds. Folds and faults both almost certainly resulted from Caledonian earth movements (Raybould, 1974).

The east-northeast-trending normal faults range in length from one to ten kilometers (or even more) and are downthrown by as much as 200 meters. Most of them are on the flanks of the Plynlimon Dome with the major ones on its west flank. Important deposits, however, were developed on fractures crossing domes in Montgomeryshire to the east and on the Llynfnant Dome to the northwest. On the slopes of the Plynlimon Dome, the larger faults are varied in structure, relationships that appear to have resulted from the interaction of the fracture-making stresses with the varied folded rocks of the dome. Three major lode-vein systems on this dome (Camdwr North, Dylife, and Castell, south) change from being normal faults with large displacements and flattish dips to vertical breccia zones with essentially no displacement. Phillips (1972) believes that these differences in the veins were effected by a gradual lessening of differential stress as the fractures were extended, the greatest initial stress having been centrally located and gradually diminished as the fractures were lengthened to the west and east. Raybould thinks that these breccia zones are the upper edges of fault planes that are more extensive and somewhat flatter at depth; he says these are typical of terminal breccia zones.

In ore bodies that were emplaced in the fracture zones, the ore lodes became larger and more frequent the farther they were followed from the centers of the fractures outward. Rich pockets of ore were found where branching and deflection of the ore-containing veins occurred and where the lodes encountered the anticlinal axes. The upper surface of the Gwestyn mudstone was, in many veins, also the base of the ore lodes, suggesting that both stratigraphy and structure controlled the localization of the ore lodes.

The following discussion of the minerals of the Mid-Wales deposits is paraphrased and summarized from Raybould (1974).

The primary minerals in the Mid-Wales veins are: chalcopyrite, sphalerite, galena, pyrite, marcasite, arsenopyrite, cobaltite [CoAsS], quartz, and ankerite. The most common secondary minerals are hydrozincite [$Zn_5(CO_3)_2(OH)_6$], gypsum, and limonite [mainly goethite]. Uncommon to rare secondary minerals are: smithsonite, hemimorphite [$Zn_4Si_2O_7(OH)_2 \cdot 2H_2O$], cerussite, and covellite.

The chalcopyrite is present in one millimeter to five millimeters disseminated blebs in both quartz and ankerite. Only seldom is it found in larger amounts or sizes in veins or in breccias. These small bodies of chalcopyrite either have their shapes governed by the earlier-crystallized minerals with which they are in contact or they grew out into cavities in quartz or ankerite, and there show simple tetrahedral crystal faces. Most chalcopyrite is developed with, or as exsolution blebs in, sphalerite; lamellae of chalcopyrite are present along cleavage planes and sphalerite-grain boundaries. Such relationships of chalcopyrite to sphalerite occur only where chalcopyrite is abundant. If galena is an abundant mineral, chalcopyrite can be dispersed as spots in the galena. The detection of some slightly deformed lamellae of chalcopyrite indicates that minor tectonic deformation took place after chalcopyrite deposition.

Sphalerite is to be found in larger or smaller amounts in all the ore lodes; in open space, it developes into large aggregates of crystals and may form layers of several centimeters in thickness, thereby producing cockade or other radial growths. Many of the ore breccias are cemented by sphalerite, and, if the distance between breccia fragments is large enough, radial growths also are normally dominant. In veins, sphalerite and quartz usually form alternating layers or, less commonly, a crystallized mix of the two minerals. In breccias, sphalerite and quartz are closely intergrown but locally may have a crude layered relationship where they surround rock fragments. In some veins and breccias, the mixture of quartz and sphalerite is so fine-grained that the individual minerals apparently cannot be recognized and the mix has a muddy brown color. The sulfide most commonly accompanying sphalerite is galena; the two minerals are precipitated in alternate layers that range from a few millimeters to several centimeters in thickness. This arrangement normally occurs in veins, but around breccia fragments a poorly defined layering may occur. The general sphalerite-galena form in breccias is an intimately intergrown mixture. In at least one place (Dylife mine), sphalerite is disseminated in the wall rock adjacent to the vein; whether the introduction of the sphalerite was by solid-state replacement or by the actual penetration of the wall rock by the ore-forming fluid is uncertain. Normally, the sphalerite is brown, but rarely it can be yellowish brown. Slight color differences in the sphalerite have been interpreted by Raybould as indicating distinct sphalerite generations; it is certain that color zoning is common. Some deformation of sphalerite, as is true of chalcopyrite, indicates that some minor movement took place after the mineralization.

Galena shows much the same relationships to quartz and ankerite as does sphalerite but does occur as much larger (cleavable) bodies and as intimate mixtures with chalcopyrite and sphalerite. Where the galena grows outward into open space, the crystal form developed is the octahedron, forms that may be as much as 1.5 centimeters in cross-crystal direction. Galena may be disseminated in small bodies in quartz and ankerite; galena also fills cavities left by the crystal growth of the earlier minerals. As chalcopyrite is not, the galena may be present in the gangue minerals as idiomorphic crystals, and large masses of galena are present in vugs in the veins. Breccia fillings contain smaller crystal aggregates and smaller disseminations of galena than those in the vugs. Continuously layered galena is present in veins and around breccia fragments, but it is less well developed in this form than is sphalerite; galena rarely forms radial growths that are so common in sphalerite. Raybould considers that the three-point contacts of galena grain boundaries might indicate that the lead sulfide have been recrystallized as a result of post-depositional deformation. He points out, however, that even well crystallized and excellently cleaved galena shows this three-point texture; since this texture in the undeformed galena could not be due to tectonic effects, it probably was not caused by deformation in any case. There is, however, a foliated type of galena that Raybould says is found in or near places in which deformational movement took place, and such movement is emphasized by elongated fragments of pyrite and small inclusions of arsenopyrite oriented in the same direction as the foliation, so some post-deformational movement must have occurred, the galena moving by flowage and the fragments of the more brittle minerals being carried along with it.

Pyrite is present in the wall rocks and the breccia fragments as framboids; these shapes are small. Raybould considers that the framboids were formed at the beginning of the main vein mineralization. In the veins, pyrite developed as small crystal aggregates (not more than two centimeters across) but much more importantly as separate cubic crystals, confined largely to the early quartz layer that lies on most rock surfaces; these crystals are seldom larger than 0.02 centimeters across. Almost certainly, pyrite was the first sulfide to deposit.

Marcasite, which Raybould does not put in his paragenetic diagram, is present in appreciable amounts only in a few places, but locally it is abundant enough to make it impossibly difficult to concentrate the ore. The tiny, tabular crystals of marcasite were late deposits on the surfaces of breccia fragments and line the cavities in the breccias in what Raybould considers "the manner of a supergene deposit". Its presence, in a few places, as small veinlets that cut across the earlier minerals, indicates, however, that it was a primary mineral and probably the last of the main sequence of sulfides to be precipated.

Arsenopyrite, found only in microscopic amounts in the veins, is in tiny grains in galena in many places, including the foliated variety of galena where inclusions of FeAsS are elongated parallel to the foliation. In places, crystals, up to 0.6 millimeters across, are in some cavities in quartz, in association with chalcopyrite and galena.

Cobaltite, in a few small, idiomorphic crystals, is present in arsenopyrite. These two minerals were unknown in Mid-Wales prior to Raybould's work.

Quartz is ever-present in the ore, constituting the major gangue mineral; ankerite is found in abundance only in a few places. The initial layer, present on vein walls and rock fragments almost everywhere, is a thin layer of quartz in a crude comb structure. This relationship obtains even where ankerite is abundant. Quartz deposition continued throughout the mineralization process; that of ankerite did not. The quartz deposited after the initial layer either forms alternating layers with the sulfides or is mixed thoroughly with them. Where quartz is in veins, it is clear or iron-stained, and the crystals in vugs may be as much as several centimenters long. Breccia quartz may be as vug crystals, but it mainly is white and less well crystallized.

Ankerite always has rhombic shapes and is pink or pale brown, though it becomes dark with weathering. This carbonate forms a large percentage of the gangue only in a few places, but in others it is localized in specific portions of the ore lode and is not widely and sparsely distributed in it. Locally the carbonate may be calcite or dolomite, but such instances are far from common.

The crustification of the minerals in the veins of the Mid-Wales deposits provides an excellent set of clues as to the order in which they were deposited. In interpreting these relationships, Raybould has found that quartz is ubiquitous and involved in every possible arrangement with the other gangue and ore minerals. It is, therefore, shown in his paragenetic diagram as having begun its deposition at the same time as the pyrite and having continued to precipitate until it, and the galena, stopped depositing at the same time. Ankerite, on the contrary, is an early mineral, beginning to deposit at the same time as pyrite finished its precipitation; this follows because, in the majority of breccia fragments, ankerite veins were in those fragments before they were broken off. Ankerite deposition continued for a short time after the (early?) brecciation had stopped but continued only through part of the chalcopyrite cycle and ceased just as sphalerite began to precipitate. The very minor amounts of arsenopyrite (and cobaltite associated with it) began to come down on a small scale almost as soon as the sphalerite started and endured through the sphalerite cycle and on into the term of galena deposition. Arsenopyrite still was precipitating after sphalerite had ended doing so but, in turn, ceased before galena stopped. Galena began deposition about half-way through the sphalerite cycle. Marcasite appears, from what he says, not to have begun until all other sulfides had ceased depositing. It must be emphasized, from the relations of the ore and gangue minerals to each other in the vugs particularly, that the cycle of mineralization was repeated once or more often than that is certain from the numerous layers of alternate quartz and the various sulfides, particularly sphalerite. Any explanation of the chemistry of Mid-Wales sulfides ore must account for such reversals.

The scheme outlined in Raybould's paragenetic diagram is that which prevails in many of the ore shoots. In many other places, however, more than one stage of brecciation occurred and appreciable cross-cutting of earlier minerals by later took place. In some places in the mines, galena began to precipitate before sphalerite. This and similar events fit into Raybould's concept of lateral zoning even though Williams questions some of his conclusions in his written discussion (p. B68-B69) but Williams depends on old production tonnages, many of which were compiled in the days when sphalerite was essentially a valueless gangue mineral. Thus, despite Williams' comments, Raybould considers that the Camdwr lode shows a significant lateral zoning. At a mine on the eastern end of this lode, the relative abundance of the ore sulfides was: sphalerite, chalcopyrite, and galena in decreasing order. In ore bodies farther west on this lode, sphalerite became less common, and, at a mine on the western end of the vein system, the ores were composed almost entirely of galena. Raybould also reports that the same relationships obtain on the Castell lode, but he admits that, in the Dylife lode, on the eastern side of the Plynlimon dome, the zonation pattern is much harder to define, but he believes that galena proportionally decreases toward the west, that is toward the central part of the dome. This lode begins in the north central part of the Plynlimon Dome and continues northeast for some nine kilometers to end at the Dylife mine. At the northeastern end, both sphalerite and galena are abundant, and as the lode is followed to the west, galena is proportionally reduced, as Raybould's general rule would suggest. He cites a few other less clear examples, but all seem to indicate that his rule is essentially correct.

In short, it appears probable that deposition of sulfides usually began at the eastern ends of the lodes with chalcopyrite being the first mineral to be deposited, followed by sphalerite, and finally be galena. Only when the proportion of galena (no matter how it was carried in the ore fluids) was high enough did galena change places with the other two sulfides in the order of depositon. If, as Raybould speculates, the metals of the sulfides were carried in bisulfide complexes, the following three equations will show how deposition could have been achieved.

(1) $Fe^{+2} + Cu(HS_2)^{-1} + H^{+1} = FeCuS_2 + H_2S$

(2) $Zn(HS_2)^{-1} + H^{+1} = ZnS + H_2S$

(3) $Pb(HS_2)^{-1} + H^{+1} = PbS + H_2S$

The normal order of precipitation ($FeCuS_2$, ZnS, PbS) in the Mid-Wales district (as in so many other lead-zinc districts in the world) probably depended on the relative stability of these bisulfide complexes. This normal order, then, is explainable by assuming that the lead-bisulfide complex is the most stable, the copper-bisulfide the least stable, and the zinc-bisulfide complex intermediate between the other two. If galena, because of a locally increased abundance, relative to the other two complexes, was the first sulfide to deposit, what was the reason? Almost certainly each additional increment of lead-bisulfide complex, over that present in the usual composition of the district's ore fluid, would add additional PbS to that required to maintain equilibrium in equation (3) above. If sufficient of these increments were added, relative to those of ZnS (the amount of which had already been reduced by appreciable precipitation of sphalerite), so little ZnS would be in solution (as such) as to allow the precipitation of sphalerite to cease while that of galena would be made possible. Continue this precipitation of galena long enough (at the same time adding, thereby, to the abundance of the zinc-bisulfide complex relative to that of lead), and galena would stop precipitating and sphalerite would begin again. Given a low enough rate of upward and/or outward progress of the ore fluid past a given point, this alternation of these two minerals could be repeated as long as the solutions contained appreciable amounts of both complexes. The rate of complex dissociation would not have to be changed, only the relative abundances of the two, if alternations in precipitation were to be effected through mass-action. The lack of the participation of chalcopyrite in these alternations is an expression, not of changes in stability of the copper-bisulfide complex, but rather that it was so much the least stable of the three complexes, that the bulk of it was deposited when precipitation began and never was built up sufficiently to allow its deposition to begin again.

In addition to the time-dependence of mineral zoning, late galena being farthest from the apparent entry points of the ore fluids into the various vein systems and the early chalcopyrite the nearest, Raybould also believes that mineralization is structure-dependent. He thinks that the greater the dip of the lodes and the less the displacement on them (as is true of the western parts of the veins), the greater the proportion of galena. The possibility exists, however, that late galena normally deposited farthest of the ore minerals from the eastern ends of the vein structures (and presumed entry points of the ore fluids). Granted this entry relationship is correctly stated and that normally the lead-bisulfide complex is the most stable of the three in question, it would be expected that lead sulfide would be most abundantly deposited in the western portions of the vein system systems no matter what the structure there, as long as the ore solutions had a reasonably clear channel to reach them.

Because of the long time since the Mid-Wales mines have been closed, Raybould did not have much opportunity to study the ores underground. He is, nevertheless, of the opinion that the deposits do show a vertical as well as a horizontal zoning. He is able to give a few examples. One of these is from a mine on the Castell lode where, at the level of the lowest adit, the ore consists mainly of sphalerite, but, at the surface, sphalerite is associated with galena. In another example, taken from Jones (1922), a clear description was given of the change from sphalerite at depth to galena toward the surface with chalcopyrite also being on the lowest levels. Although it is probable that these two examples are typical of what would have been found in all, or nearly all, the mines, it is too late in the history of the district for this concept of vertical zoning to be more than that.

Raybould's paper was published not long after his faculty supervisor (W.J. Phillips) had published his suggestion (1972) that hydrothermal fluids might produce breccias of three types, namely: (1) hydraulic breccias in which rock fragments are highly angular but are divided each from its neighbor by only narrow distances; (2) (at the other extreme) intrusion breccias in which the fragments are much separated, completely disoriented, and, in many places, rounded by abrasion; and (3) (and intermediate between the other two) breccias with considerably separated but angular fragments. Phillips has argued that, in many ore veins, the same breccias characterize the sequence from vein center to vein walls with the matrix-fragment ratio decreasing from the vein center to the walls as it would do along the vein length.

The lack of breccias of type (2) (intrusion breccias in which the fragments are widely separated and completely disoriented, and, in many places, rounded by abrasion) indicates that the formation of breccias in the Mid-Wales veins went only through type (3), the fragments never having been sufficiently agitated to be well rounded. In all Mid-Wales breccias, the fragments are separated by considerable distances, and the minerals deposited around them tend to show radial growth. These minerals are particularly quartz and sphalerite, and, in many places, the fragments are completely ringed by such radially arranged minerals. Further, the widely separated positions of the fragments could have been maintained while the matrix material (radial or otherwise) was deposited around them only if they were in some manner held in place, out of contact with each other, so rounding could not take place. There is no evidence that these fragments (before the cementing matrix was sufficiently deposited) were supported by anything but water. Granted that this is true, the water must have been moved upward by tremendous force if the fragments were to be kept from settling to the bottom of the chamber in which they had been developed. Still further, this upward moving water could not have been highly turbulent or rounding would surely have taken place, but it must have moved upward with such force that a large proportion of the fragments produced by the initial explosive action must have been moved out of the breccia zones, probably having been thrown out on the earth's surface. If this were not done, the widely separated character of the fragments could not have been achieved.

In zones of lower solution velocity (away from the center of upward moving water), the force applied would have been great enough only slightly to separate the fragments, one from the others; in fact, they would have remained close enough to each other to have provided mutual support after only slight relative movement. A major result of these differences in relative fragment movement would have been an outward decrease (along the length of the vein) in the matrix-fragment ratio. Raybould believes that his concept of rapid crystallization of the breccia matrix is upheld by the fine-grained texture of the quartz layers around the fragments and by the disordered and fine-grained intergrowths of galena and sphalerite that normally form the matrix of this type of breccia.

Many of the breccia fragments show evidence of rebrecciation; second-stage fragments may be angular and completely surrounded by matrix material, yet, within these fragments are textures that clearly show that they were once part of a mass produced in the first cycle of brecciation. As many as four stages of brecciation have been recognized by Raybould; this suggests, since the fragments of all stages appear to have been supported by no material more dense than water, that the force behind that water was immeasurably strong and long, if intermittently, continued. The problem of the development of these breccias needs further study, both in the field and in the laboratory, particularly because modern laboratory techniques permit the use of pressures of the order of magnitude that appear necessary to produce the type of brecciation characteristic of the Mid-Wales ore field.

As regards the precipitation of the sulfides in the Mid-Wales ore veins, Raybould holds that, if the metal in these minerals had been carried as chloride complexes, they might have picked up enough sulfur when they passed through the pyritic Gwestyn mudstones to have caused the rapid and abundant precipitation of ore-metal sulfides. Instead, the sulfides are found well above the mudstones indicates that they were not transported as chloride complexes. He believes that positive evidence of the transport of the ore metals in bisulfide complexes is provided by the local reversals of the order of the beginning of PbS and ZnS. Normally, the zinc-bisulfide complex is less stable than that of lead so the first of the two sulfides to have precipitated would have been sphalerite with galena following after the conditions for exceeding its stability had been reached. As has been pointed out on page above this one, the presence of unusually high concentrations of lead-bisulfide complex in the ore fluid at a given increment of time might, through a mass-action effect, have permitted a major amount of galena to deposit before sphalerite. It may be that this relatively greater abundance of the lead- over the zinc-bisulfide complex may have resulted from a large-scale early precipitation of sphalerite in one part of the vein having depleted the zinc-bisulfide complex relative to that of lead by so much that galena, in the locality in question, precipitated in considerable amount before sphalerite. Raybould suggests that the difference in stability between the two complexes may have been sufficiently small for this reversal to be possible.

Raybould ends his paper by suggesting that the ore fluids originated as interstitial brines that, either before or during migration, took the ore metals into solution as bisulfide complexes. These brines, he supposes, migrated because of regional compaction and folding and that the low-angle fractures formed during compression became the sites, first, of, quartz deposition and of that of very minor lead-zinc mineralization. During subsequent relaxation of compression a new set of normal stresses developed and produced normal faults and provided enough momentum for huge quantities of ore fluids to build up sufficient pressure to fracture the rocks into the pattern of veins that the present ores occupy. This means that the introduction of the ore minerals took place toward the end of the Caledonian revolution and that, therefore, the Mid-Wales ores should be dated as mid-Paleozoic. The two major objections I have to this scenario is that: (1) I doubt if a "new set of normal stresses" could have produced the huge pressures that were necessary to develop the unusual breccias of the Mid-Wales ore district and (2) connate brines could scavenge the huge quantities of metal necessary from the rocks beneath the ore district or could provide the energy needed to convert these metals into bisulfide complexes. I wish that Raybould had elaborated on his statement that "there is now abundant evidence that formation waters are capable of leaching those metals to form mineralization fluids". If he had said "abundant speculation" instead of "abundant evidence", I would agree with him. He is, of course, correct in saying that igneous rocks are known to be associated with the Mid-Wales ores, and this argument has been used to justify a wide variety of theories as to other unknown sources that are supposed readily to supply the ore fluids that the unknown igneous rocks are thought to be unable to do.

Putting aside the controversy as to where the ores fluids came from or how they originated there, the deposits can be classified on the bases of what the fluids produced. The simple mineralogy of the deposits, however, makes it difficult to position them with certainty in the modified Lindgren classification. The small quantities of chalcopyrite, relative to sphalerite and galena, diminish its importance economically but still may provide clues as to the temperature of ore deposition. Chalcopyrite contains exsolution blebs of sphalerite and vice versa. This indicates that the deposition of these two minerals probably began in the upper portion of the mesothermal range or even in the lower part of the hypothermal. The silver content of the galena is sufficiently large to make it almost certain that the galena was not formed in the telethermal range. This means that the galena is certainly leptothermal, in part at least, and that the remainder may be mesothermal. The minor inclusions of arsenopyrite and cobaltite in galena, suggest the mesothermal rather than the leptothermal portion of the classification. Despite a suggestion by El Shazley and others (British Lead-Zinc, 1956-1957) that the trace elements indicate that the mineralization was of a relatively low-temperature variety, the evidence of the silver in the galena and the chalcopyrite blebs in the sphalerite seems to indicate that their concept was not correct. The best compromise that can be made, lacking data from fluid inclusions, is to classify the deposits as mesothermal to leptothermal, and that is done here.

NORWAY

SULITJELMA, NORDLAND

Middle Paleozoic	Copper, Pyrites, Zinc	Hypothermal-1 to Mesothermal Metamorphic-C

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Notes

The town of Sulitjelma (67°10'N, 16°05'E) is located 75 km slightly south of east from the town of Bodø on the North Sea and only 15 km from the Swedish border, less than 70 km north of the Arctic Circle and on the southeast end of the north side of the major lake, Lomivann. Sulitjelma is at least 10 km east of the most easterly of the currently operating mines.

The first discovery of ore in the district was made in 1860 by a Lapp, Mons Petter; exploitation did not begin until 1891. The early geological exploration was carried out by, or under the direction of, Hjalmar Sjögren at a time when Sweden and Norway were part of a single kingdom. At about 1975, the total production from the district had been 20 million tons containing 370,000 tons of copper. Of this 20 million tons, 14 million came from the northern part of the district and 6 million from the southern (Baldoaivve) section. Production in 1974 was 420,000 tons of ore from which were obtained 19,129 tons of copper concentrates (29 per cent Cu), 1000 tons of zinc concentrates (45.66 per cent Zn), and 59,000 tons of pyrite concentrates (49.9 per cent S). The copper concentrates are smelted to blister copper in a local electric smelter.

The first comprehensive account of the geology of the Sulitjelma district was published by Th. Vogt (1927), but this work did not discuss the ore geology; for many years it was understood that Th. Vogt would provide a second part to his earlier work, but his untimely death came without that ever having been done. Between Vogt's 1927 publication and the succession of major works that began with papers by Nicholson (1966, 1967), Nicholson and Rutland (1969), Henley (1970), and Mason (1967, 1971), the only major publication was that of Kautsky (1952) and this was based mainly on work done on the Swedish side of the border.

At present, mining is essentially confined to the northern part of the district, principally from (from west to east) the Bursi, Mons Petter, Furulund, Charlotta, Giken, Hankabakken, and Ny Sulitjelma mines. The most prominent mines in the southern area are: Jakobsbakken (exhausted) then clockwise to Baldoaivve, Ingeborg Stålhaug, Furuhagen, and Sagmo. Most of the sulfide-ore mines are located in the upper part of the Furulund group (see below) - Giken zone and Ingeborg-Stålhaug zone; or at the contact of the Sulitjelma amphibolite-Charlotta zone and Jakobsbakken zone. The Mons Petter mine is in the banded amphibolite and the minor Fjeld Gruve is at the top of the amphibolite a little more than one km NNW of Ny Sulitjelma. The uneconomic mineralization in the Kong Oscar Grube is within the Sjønsta group.

The basement of the Sulitjelma district is granite gneiss that is (Wilson, 1973) a continuation of the Precambrian Baltic Shield. The Caledonian rocks that lie above the basement are quite thin; these make up several separate structural units that are divided from each other by early thrusts. Both the rocks and the planes of the thrust faults have undergone later folding and metamorphism. The nappes in the district are (from oldest to youngest); Beiarn, Gasak (or Rødingsfjall), and Seve-Köli. Except for short distances above which the Gasak nappe has been eroded, most of the surface is composed of rocks brought in above the Gasak nappe.

In the Sulitjelma district, two nappe units have been recognized; the upper is a series of various metasediments known as the Sulitjelma Schist sequence and the lower made up of amphibolites that were emplaced above calcareous schists and phyllites of the Furulund group. Although Kautsky (1953) thought that these Sulitjelma amphibolites lay above another nappe surface, recent work (Rutland, 1969, Wilson, 1973) has shown that no such thrust exists. No ore bodies are known in the Sulitjelma schists and the amphibolite itself contains only the Mons Petter mine within it and the Fjeld Gruve on top of it. Most of the ores are in the upper part of the Furulund group or directly beneath its contact with the marginal brecciated equivalents of the Sulitjelma amphibolite that separate the Sulitjelma schists from the Furulund rocks. Below the Furulund is the Sjøstra group, unmineralized except for the minor Kong Oscar Grube, and below the Sjønstra is the Pieske marble and certain psammites and the ancient basement.

The Furulund group is composed of calcareous phyllites and schists; these are well-banded and, in parts, are rich in garnet and hornblende, that is, are above the garnet-hornblende isograd (Winkler's reaction isograd, *mutatis mutandis*). This group also contains mafic intrusions, now metadolerites or metadiabases; these have rather coarse-grained interiors, but their margins are highly deformed (Wilson, 1973). These once-continuous intrusions have been broken, thinned, and stretched in boudin-like bodies that are elongated parallel to fold axes. In some instances, these intrusive bodies appear to cut across the schistosity and may, therefore, not all be pre-tectonicism. West of the Ny Sulitjelma mine (at the eastern end of the north ore belt) the contact between the Furulund group and the amphibolite has been confused by late deformation; in the Ny Sulitjelma area and to the east, however, a gradation between these two rocks has been demonstrated. In this eastern portion of the area, the number of thin amphibolite bands that were inserted into the Furulund group gradually increased. These were accompanied by the development of quartzite horizons, quartzite conglomertes, and calcareous bands. The zone in which these various lithologic elements were produced ranges from 1 to 30 m. Above this banded rock section, the amphibolite is massive.

In the eastern part of the north ore belt, the amphibolites are fine-grained and well-banded (minerals are hornblende, plagioclase, and clinozoisite). Included in the amphibolite body are bands of schist that are as much as 5 m thick, plus bands of conglomerates that contain amphibolite pebbles with a matrix of mica schist. The rocks, and their varied thicknesses, suggest to Wilson (1973) that the intrusive rocks originally were mafic volcanics.

In the western part of the north ore belt, the amphibolite thins and some of the bands are coarse-grained, although some remain thin. The coarse-grained bands are in places amphibolites and in others quartz keratophyres. The amphibolite unit has been strongly strained to produce a tectonic breccia in which boudinage has been a major process. The ore bodies characterize the junction of the Sulitjelma sequence and the amphibolite, and the ores are accompanied by late chloritization and zoisite formation. Major disharmonic folding was developed in the vicinity of the Bursi mine. In the far western part of the area and where amphibolite is known south of Langvann (lake) the entire thickness of the amphibolite unit is no more than a few meters. In this section, the zone between the top of the normal Furulund group and the base of the normal Sulitjelma Schist group (that is, the thin section of amphibolite) has been designated by Henley (1970) as a "junction unit".

Wilson (1973) recognizes three deformation phases within the Furulund group schists and designates them as D_1, D_2, and D_3. The first two events were associated with the main metamorphism and thrusting, whereas the third was confined to the post-schistosity folding.

The principal metamorphic fabric was developed in the course of D_1 and D_2. It is possible to distinguish these two phases because the garnet and hornblende porphyroblasts were produced during D_2 and because the Sulitjelma gabbro was intruded into the Sulitjelma schist during a pause between the two metamorphic phases (Mason, 1967, 1971). This break, however, was not particularly well marked especially in the Furulund group. Mason considers the finite strain ellipsoid for D_1 and D_2 to be indicated by the schistosity and the mainly E-W- or NW-SE-trending penetrative mineral lineation. The maximum elongation of deformed pebbles is east-west oriented. The greatest part of the deformation of tectonic breccias (in the Sulitjelma amphibolites) is thought to have taken place during the D_1 and D_2 folding.

During the D_3 deformation, large-scale folding was related to basement doming and took place to the west of the Sulitjelma region (Nicholson and Rutland, 1969). Under the same tectonic forces, large-scale gentle warping occurred in the Sulitjelma district proper. In association with localized sliding, important harmonic folds were developed in the Bursi area as well as at Furuhaugen on the south side of Langvann.

The ore bodies in the Sulitjelma district have been affected by the early metamorphism and by the regional metamorphism. Wilson (1973) advances the following evidence for this statement: (1) the ore bodies are highly elongated into ruler-shapes that are oriented NW-SE and plunge NW (Kautsky, 1953). The lineation of the ore bodies is about parallel to the stretching lineation of the rocks and with the "Gleitstriemen" lineation of the rocks which is normal to the strike of the Caledonides. [This is taken essentially from Wilson's translation of a statement by Kautsky in his 1953 paper.]

Such an elongate shape is best illustrated by the Jakobsbakken body that is at least three times as long as it is wide. The shapes of other bodies do not show this elongation nearly as well because of erosion, the present land surface being parallel to the elongation direction. It has been suggested (Vokes, 1968, disc.) that the elongated shapes of the ore bodies [at Løkken in the example quoted] may, in part, be primary because the internal strains observed in the country rocks do not seem enough to account for the extreme elongation of the ore bodies [those at Løkken are 10 times as long as they are wide]. Such elongation is not matched at Sulitjelma, but it still is possible that some of it is due to primary factors [including possible replacement of primary sedimentary-sulfide structures by sulfides].

(2) At Bursi, a small ore body has a NE-SW strike where it has been concentrated in the neck of a large boudin; the development of which was caused by the D_1-D_2 penetrative strain.

(3) The ores are closely associated with the tectonic breccia and this probably formed through the D_1-D_2 penetrative strain. Fine-scale brecciation is quite usual in the western portion of the Sulitjelma area; there are seen numerous examples of "the highly penetrative type of rotational strain known as "Durchbewegung" [literally "through-movement"] which is often found in this type of sulfide deposit" (Vokes, 1968). It is thought possible, however, that these Durchbewegung structures were produced by highly localized movements of D_3 age.

(4) The Sulitjelma ores have been considered as examples of metamorphosed sulfide ores in which large pyrite porphyroblasts overgrew other sulfides and were locally considerably deformed themselves.

(5) The ores were affected to an appreciable degree by D_3 events; unusual post-schistosity disharmonic folding [characteristic of rock layers that have significant contrasts in viscosity] and sliding that produced concentration of the ores in the cores of the folds. This folding usually was preceded by very strong brecciation; later the district underwent retrograde metamorphism in which chlorite was developed. Previous authors have thought that this chlorite be produced by hydrothermal solutions during the emplacement of the ores. Wilson, however, believes that the evidence is good that the chlorite replaced biotite in the matrix of the tectonic breccia. If this is true, then the chlorite probably was pre-ore.

Wilson, on the basis of these five points, is convinced that the ores were introduced before or during the early penetrative phases of deformation (D_1-D_2).

The basal granitic gneisses of Nordland probably belong to a single unit of basement rocks that are, at least in the main, authochthonous and are joined to the Baltic Shield. Wilson and Nicholson (1973) say that a line that cuts across the Nasafjäll abd Rombak windows in a NNE-SSW direction is the eastern boundary of Caledonian recrystallization and of basement doming. East of this line, the gneisses exhibit structures that are not related to the overlying Caledonian thrust blocks; also, in places, these rocks are covered by thin authochthonous sediments.

Rb-Sr whole rock isochron ages have been produced for Nasafjäll rocks, and these give an age of 1780±43 m.y., suggesting that the basement gneiss in this window attained their pre-Caledonian form during the Svecofennian orogeny.

Rai (1977) says that the mineralized zone on the northern side of Langvann (lake) can be traced for 100 km, but economically workable deposits are limited to about 8 km of this length. Mining operations south of this lake were abandoned many years ago (Rai, 1977).

The ore bodies consist of at least three morphologic-mineralogic types: (1) massive pyritic ore, (2) disseminated ore, and (3) pyrrhotite-chalcopyrite ore. The three ore types can, in many ore bodies, be found intermixed in proportions that differ considerably over short distances. The mineralogy of the ores is most simple. The only economically important minerals are pyrite, pyrrhotite, chalcopyrite, and sphalerite; the minor and rare minerals make up between 2 to 3 per cent of the total ore mass. Such minerals, widely dispersed through the ore, include: galena, arsenopyrite, tetrahedrite, molybdenite, mackinawite, bournonite, plus vanishingly small amounts of many other sulfo-salts. When Ramdohr studied the Jakobsbakken ores (1938) he reported numerous Cu-, Pb-, Ag-, As-, and Sb-sulfides and sulfosalts, plus Au, Ag, and Sb as native metals; these Jakobsbakken ores are remarkably rich in antimony for this type of ore. Both Ramdohr and Rai are convinced that the ores have been affected to various degrees by high-grade, regional metamorphism.

Rai (1977) approached the problem of the genesis of the Sulitjelma, and by inference of all other massive-sulfide, deposits in Norway at least, by determining the contents of trace elements in the typomorphic minerals (i.e. pyrite, pyrrhotite, chalcopyrite, and galena) from the various ore bodies of the area. He quantitatively determined such elements as Co, Ni, Mn, Mo, Cr, Ti, V, Ga, Cd, Ag, and Pb. As, Sb, Bi, Sn, and Se were determined semi-quantitatively. His results, given in ppm, show that two types of ore deposits (the first and third types listed above) exhibit the same suite of minor and trace elements through their occurrences in the deposits. Nickel, for example, is quite constant, far more at least than cobalt. Similarly constant are Mn, Ga, Ag, As, and Sb within moderate limits. The behavior of cobalt is quite erratic, which Rai explains as being due to its extraordinary sensitivity to post-depositional events such as metamorphism, recrystallization, and remobilization of the ores. Rai suggests that such geochemical regularities of the type he found in the Sulitjelma ores indicates that the ores were formed sedimentarily or by volcanic-exhalative processes. These patterns are, he contends, in direct contrast to the abundance patterns of minor and trace elements in typomorphic minerals of magmatic and hydrothermal deposits. He quotes two Czech authors (Cambel and Jarkovsky in a publication not available to me) as saying that such deposits are characterized by varied and irregular contents of minor elements due, they believe, to influences of wall rocks, temperature of ore solutions, periodicity of mineralization, and several other (unnamed) factors. He remarks, however, that a particular ore type generally shows slightly, yet characteristically, different levels of concentration of an element in a particular mineral in different ore bodies of the deposit but exhibits remarkable consistency in the abundance of that element in that mineral on the scale of the ore body. Rai goes on to add that such definite and consistent differences in the minor- and trace-element composition of ore in different ore bodies of the deposit, joined with the differences observed in their base-metal composition, appear to reflect primary differences in their overall composition. These differences, he thinks may be explained by rather small changes in genetic, thernodynamic, and other conditions of ore deposition in different ore bodies. This all may be quite true, but it tells very little as to the extent to which each of these factors affected the problem.

Rai also observed, so he reports, that a particular ore type shows closer and quite interdependent geochemical relations with other ore types associated with it in the same ore body rather than with the same ore type in other ore bodies of the deposit. He thinks that this means that this interdependence of abundance patterns of different elements and their distribution ratios in typomorphic minerals in these ore types offers strong support to the concept of a palingenetic mode of origin for the pyrrhotitic ores.

It appears that the meaning applied to "palingenetic" or the "palingenesis" in Scandinavia is not what is given as a definition in the "Glossary of Geology". There palingenesis is defined as: "formation of new magma by the melting of pre-existing magmatic rock in situ". It would appear that the word that would suit better what is being said by Rai and other Scandinavian ore geologist is "anatexis", which is defined by the same source as "melting of pre-existing rock; the term is commonly modified by such terms as intergranular, partial, differential, selective, or complete". The words partial, differential, or selective should be added to anatexis, and the term would better fit the process they have in mind than does palingenesis. [Incidentally, these remarks about palingenesis apply equally well to the use of that term in the discussion of the deposits of the Røros-Kvikne-Killingdal area as it does to that on Sulitjelma.]

Rai also has considered the isotopic ratios obtained from the Sulitjelma sulfides and their relations to possible processes of ore genesis. As a whole, the deposits show a spread in delta-^{34}S values between -0.1$^{\circ}/_{oo}$ and 8.44$^{\circ}/_{oo}$. The massive ores (of both pyritic and pyrrhotitic types) show a much narrower spread (+3.55$^{\circ}/_{oo}$). The delta-^{34}S values outside this narrower range appear to be confined completely to the disseminated pyritic ores.

Despite this difference in values between ore types, Rai considers the overall spread to be quite narrow and close enough to the meteoritic standard (allowing for reactions between the source of the ore fluids and their arrivals in the rock volumes where they deposited their sulfide loads) to make it possible to think that the fluids came from an independent and fairly homogeneous deep-seated magmatic source. Allowing that this is true, it remains to decide whether these solutions deposited their sulfide contents entirely with unconsolidated sediments or entirely in already lithified rock or a combination of both, that is, with the primary pyrite being volcanic-exhalative and the other base-metal sulfides being hydrothermal in the sense of depositing in solid rock.

Rai believes that his study of the distribution of major metals, particularly copper and zinc, in the various ore types and ore bodies of the Sulitjelma deposit bring out certain important aspects of the metallogeny of the region. These facts suggest to Rai a primarily mafic igneous source of the ores and indicate a close genetic connection between massive pyrite and disseminated types of ore with both of them deriving from the same pulse of ore-forming fluids. It would be surprising if, on their upward journey, the various entities of ore-fluid had not reacted differently enough within themselves and with their surroundings to produce somewhat different ore-mineral and trace-elements contents in the ore bodies they formed to account for the (relatively minor) differences that hold among the various ore bodies of the district. Rai believes that a "palingenic origin" or pyrrhotitic ores, presumably during the regional metamorphism of the deposit, is supported by his study. The major problem needing resolution, in his opinion, is that of the effects of metamorphism on the primary geochemistry of the ores.

The minor, but definite and consistent, differences in composition of ores in the various ore deposits of the district, however, seems to him to have been a primary feature of the deposits, and this appears difficult to explain by either a simple hydrothermal or simple sedimentary scheme of ore genesis. The complications that he sees, however, he believes can best be explained by the volcanic-exhalative type of deposition having occurred over a moderate time span, geologically speaking.

Kautsky (1958) thought that the sulfide mineralization took place at the same time or after the thrusting of the nappes; that is after or during Wilson's D_3 stage of deformation. From this it follows, if true, that the Sulitjelma ore bodies are definitely younger than the effusive rocks and schists in the nappes; therefore, the ore bodies could not, in his opinion, be syngenetic but epigenetic. Included in the rock types that are either adjacent, or in close proximity, to the ores are no effusive silicic rocks; instead the host rocks are mafic effusive greenstones, gabbros, and coarse-grained granites. Kautsky makes this point because Oftedahl's 1958 paper requires silicic effusive rocks to produce the volcanic-exhalations needed, as he thought then, to develop volcanic-exhalative ores. Since Kautsky wrote these remarks some 24 years ago, it has become apparent to most ore geologists studying the region that the ores were in place far enough before the end of metamorphism for considerable changes to have been wrought in the textures of the ores so they definitely bear the imprint of appreciable metamorphic action.

Kautsky, however, points out that the different shapes of the Sulitjelma ore bodies nearly always were the result of tectonics; he holds that this applies especially to the elongate ore bodies that follow the regional lineation produced during the metamorphic episodes. Oftedahl assumed [and it is now forgotten how wrong some of his assumptions were] the shape of the Caledonian sulfide ores might have been due to the action of ocean streams on the exhalative-sedimentary materials when they were deposited on the ocean floor.

Further, Oftedahl (1958) said that the Caledonian sulfide deposits in Sulitjelma and elsewhere show every variation from the primary exhalative-sedimentary ores [and these are now almost universally admitted to be such] of the Vignes-type through all stages of metamorphism, shearing, and metasomatic recrystallization to the most metamorphosed ore bodies in the country. Kautsky, however, points out that for the two major types of pyritic ore there are no intermediate transitional types. The ores of the Leksdals type are thin layers of pyrite sandwiched between major amounts of sediments; these usually are near mafic extrusive rocks of Ordovician age (Vignes-vasskis-Leksdals type). The second type, that shows no transition types with the Vignes type, are large pyritic deposits that normally contain amounts of copper economically worth recovering, that show considerable to major degrees of metamorphism (Kautsky considers this metamorphism to have been produced by lower Devonian deformation). The compact ore of this type normally is rich in massive pyrite in their central parts, whereas the peripheries and an aurole of impregnation ore outside the massive material are normally rich in chalcopyrite. The pyrrhotite-chalcopyrite-rich ores probably developed from the massive pyrite type by replacement reactions promoted by copper- (and usually zinc-)rich ore fluids that followed those that deposited the pyrite or were deposited, in some ore bodies, after the pyrite had been introduced but by the same generation of ore fluids. The major question as yet unresolved is: were they deposited syngenetically on the ocean floor or epigenetically by the replacement of already lithified rocks?

Vokes, for example, in his study of the Birtavarre district suggested that these deposits were formed by palingenetically generated fluids that deposited the ores of this deposit epigenetically. He has since, of course, changed his mind as any true scientist has the right to do. This change was due in large part to his having seen many more of such massive sulfide deposits in the field and having reached the conclusion from this study that the sulfides were emplaced before the orogeny. Further so little time was available after the deposition of the sediments and the beginning of the orogenic epoch that they, the primary sulfides, must have been deposited while these sediments still were unconsolidated. This certainly, as Vokes admits, is not complete proof that the Norwegian massive sulfide deposits in general or the Sulitjelma deposits in particular are syngenetic, but this concept has been accepted by most ore geologists who have studied the Sulitjelma (and similar deposits). It is easy to say, for example, that "chalcopyrite [is] injected plastically into the enclosing schists". But the absolute truth of this statement is much more difficult to educe.

There is no doubt but that, under sufficient metamorphic pressure, massive sulfides can flow over short distances and that during this flowage or before pyrite porphyroblasts can be produced, swimming in a matrix of more plastic chalcopyrite (and sphalerite). But, again, this is not proof as to the exact time, relative to the beginning of the metamorphism, that that melange of sulfides was introduced. I think that it is only fair to render the old Scottish verdict of "not proven" for the concept that all massive sulfides in stratified rocks must, initially, have been introduced syngenetically.

Thus, the Sulitjelma ores (or the ores of any other Norwegian massive sulfide deposit or of any other such deposit anywhere in the world) can be classified as II-A-3 plus Metamorphic-C. But it also is reasonable to suggest that they also may be categorized as Hypothermal-1 to Mesothermal plus Metamorphic-C, as is done here.

If the deposits were formed under category A-3 plus Metamorphic-C they would be early Paleozoic in age, but if they are correctly classed as Hypothermal-1 to Mesothermal, plus Metamorphic-C they must be middle Paleozoic in age.

BLEIKVASSLI, NORDLAND

Middle Paleozoic	Lead, Zinc	Hypothermal-1 to Mesothermal

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Notes

The Bleikvassli mine (65°50'N, 13°50'E) lies nearly 50 kilometers south-southwest of Mo i Rana (if Vokes map is to be believed over Bugge's statement that the distance is 70 kilometers to the south). The distance by road is about 75 km, the area ranges from hilly to mountainous, and the altitude of the mine, on the north slope of Kongsfjell Mountain, is 450 m; the mountain itself reaches an elevation of 1060 m. The discovery of the ore was made in 1914, but only in 1948 was serious exploration and development undertaken. Total production through 1974 was about 2,100,000 tons, and 2,600,000 tons remained in reserve of which the grade was estimated to be 2.4 per cent Pb; 4.2 per cent Zn; 0.22 per cent Cu; and 17 per cent S. Production in that year was 140,000 tons of ore. In 1963, the concentrates were brought by truck to a shipping dock on Ranafjord, slightly west of Mo (Bugge, 1978).

According to Vokes (1963), one of the principal features of this area of Paleozoic folded mountains is a thrust plane that trends generally north and provides a boundary between Nordland facies rocks to the west and Trondheim facies rocks to the east. The rocks differ greatly in both lithology and metamorphic grade. The Nordland rocks are thick pelites with minor quartzites, and, in the upper levels, much limestone and dolomite; rocks that were apparently volcanic in origin are minor in quantity. The metamorphic grade of these Nordland rocks is the epidote-amphibolite or amphibolite regional-metamorphic facies. On the contrary, the Trondheim rocks (east of the thrust plane) are of low metamorphic grade and include high proportions of silicic and, more importantly, mafic volcanic lavas and pyroclastics. These rocks have been intruded by large numbers of igneous bodies of various sizes that range from partly serpentinized peridotites through gabbro to trondhjemites (quartz diorites).

The Bleikvassli ores and the highly metamorphosed rocks that contain them, therefore, have been thrust eastward as the Rødingfjell nappe, together with several sulfide prospects of probably little or no economic importance, to lie over rocks of appreciably lower metamorphic grade. These underlying rocks, especially across the Swedish border, also contain massive sulfide deposits that are rather small and include various amounts of copper, plus some lead and zinc. In the neighborhood of Bleikvassli (7 km southeast on the southern slope of Kongsfjell) is an area of weak chalcopyrite-pyrrhotite and galena-sphalerite-pyrite mineralization that had not, in 1962, been much investigated. Another lead-zinc occurrence is on Rostafjell, 30 km due north of the mine, and northeast of Rostafjell, across a small fjord, are both lead-zinc and copper-pyrrhotite showings near Svalengen. At Mofjell, in the immediate area of Mo i Rana is the working mine of Mofjell that was discovered over 300 years ago, although it was not mined seriously until the late 19th century. Up to 1974, this mine produced 2,400,000 tons of ore from which were extracted 188,000 tons of zinc concentrates, 22,000 tons of lead concentrates, 21,000 tons of copper concentrates, and 177,000 tons of pyrite; silver runs 700 grams to the ton of concentrated galena. The literature about Mofjell is sparse and mostly in Norwegian; this is a pity as it seems to be of equal geological interest and importance to Bleikvassli.

The Bleikvassli deposit is strata-bound (used here in non-genetic sense) within a series of rocks assigned by Ramberg (1967) to the Kongsfjell group that is composed of a succession of garnet-mica schists that contain horizons of quartzite, graphite schist, calcareous mica schist, and amphibolite. The rocks associated with the ores are staurolite-kyanite gneisses and microcline gneisses. Ramberg considers the microcline gneisses to be metamorphosed sediments because of the regular stratigraphic relation to the beds with which they are intercalated. In addition to the sulfides at Bleikvassli, sulfides also are present in similar gneisses in other places in the area. Because of the stratified nature of these volcanics as well as of those at Bleikvassli, Bugge, Vokes, and other authors favor a volcanic-exhalative manner of formation for the sulfides throughout the Kongsfjell district. This matter will be considered in detail later.

The general strike of the rocks in the vicinity of the mine is northeast with dips to the northwest, but, in detail, the structure is far more complicated than that statement would suggest. This mine is located around the edges of a major synform (rather than syncline), and the ore bodies apparently are on both sides of that structure. In the northernmost stopes, the microcline gneiss is on both walls of the ore body. Highly schistose gneisses and micaceous schists lie along the southern part of the mine and, in Vokes' report (1963), no definite stratigraphic units had been used to designate them. On the lower levels of the mine, highly micaceous schists and gneisses, similar to those known in the southern part of the mine on higher levels, are present on both walls of the ore. Bands of quartzite are interbedded with the schists and gneisses; the quartzites are from less than one meter to two or three meters in thickness, though locally the thickening may be due to the repetition of isoclinally folded beds. Two types of quartzites are known in the mine area: (1) gray to white, fine- to medium-grained banded rocks essentially without muscovite and (2) dark-gray to black with its color being caused by various amounts of finely divided graphite disseminated through the rock; iron sulfide (pyrrhotite) is abundant in this type of quartzite. The graphite-sulfide quartzites have not been found on the surface. Also, near the footwall of the ore, concordant bands of coarse garnet-mica schists are present; these range in thickness from a few centimeters to perhaps one or two meters. The garnets are from 0.5 to 2.0 cm in diameter. The bands contain considerable amounts of disseminated sulfides. The hanging-wall rocks are quite like those in the footwall in the southern part of the mine, being schists and schistose gneisses with major amounts of mica and local bands of coarse-grained garnet schist. Surface mapping in the district was most difficult because of the problem in distinguishing one rock unit from another over any but the shortest distances.

At the north end of the mine, both the hanging- and footwalls contain microcline gneiss, which, in places, has a streaky and contorted look with the micas being segregated into lenses, bands, or schlieren of irregular shape.

In the central portion of the mine, on the hanging-wall, there is a local occurrence, as far as was known in 1962, of the microcline-quartz-biotite gneiss that is typical of the footwall. In some places in the mine, staurolite and kyanite are present in the gneisses.

The working of the mine has shown that the ore body is made up of a series of branching and reconnecting lenses or plates of massive sulfides that appear to have been concordantly introduced into the schistose and gneissic country rocks (at some stage in their development). The schists are more common in the southern part of the ore body and here the concordant stratigraphic relationship between ore and schists is most marked. Nevertheless, Vokes notes some slight cross-cutting of the wall rocks by the ore in the northern section, but the meaning of which Vokes (1962) had not then fully determined.

The individual plates or lenses of the massive sulfides differ greatly in thickness both along strike and down dip. The greatest width of the ore (1978) is reported by Bugge to average between 20 to 25 meters. Locally, the ore lenses contain barren schist intercalations of lesser thicknesses. At other points, however, the plates have contracted to no more than strips of sulfides one or two decimeters thick. Further, individual plates and lenses thin out along both strike and dip and, in many places, completely disappear only to reappear in an en echelon position to those that had died out. It appears to be difficult, amid these branching, and reuniting, and disappearing ore bodies, to determine which is the main one and which are dead ends. If a lens is followed to the northeast along strike (northeast-southwest) the new ore mass shows up in the hanging-wall of the lens it replaces. When followed down dip, the new ore appears in the hanging-wall of the one above it. It would seem logical to believe that, from a single ore lens, branches on the hanging-wall side would turn to the north, whereas those on the footwall would diverge to the east or southeast, but (1962), this problem did not seem to have been settled. The dips of the ore lenses range between 40° and nearly 90°NW. In 1962, Vokes suggested that the ore pitched at 90° to the horizontal.

Some of Vokes' mapping indicated to him that the detailed form of the ore body in many places is controlled by tight, often isoclinal, overturned folds that involved both wall rocks and ore. The plunge of the fold axes and of the lineation in the ore area is southwest or west-southwest at between 18° to 36°, which indicates that there is no direct correlation between this feature of the structure and the form of the ore zone (Vokes, 1963). Although the walls of the ore bodies are normally conformable to the schistosity in the wall rocks, cross-cutting relationships occur between the ore lenses and the fold structures. Contacts between ore and rock generally are sharp but locally the ore grades from massive through steadily decreasing impregnations of sulfides (mainly pyrite) into barren wall rock.

The ore within the lenses and plates is of massive-sulfide character with the gangue minerals never making up more than 40 per cent by weight of any sample of ore. Beyond (outward from) the massive ore, the wall-rock schists are mineralized to a decreasing degree, especially on the hanging-wall. The disseminated ore can, in places, be quite rich, particularly in lead. Both ore types are stoped together in many places.

The ore at Bleikvassli is of two types: (1) massive ore and (2) wall-rock mineralization. In turn, the massive ore is divided into two varieties: (a) pyrite-sphalerite-galena ore and (b) pyrrhotite with about the same galena and sphalerite content as the pyrite variety but with appreciably more chalcopyrite. The pyrite variety is much more abundant than the pyrrhotite ore, in almost all places, is adjacent to the footwall. The percentage of sulfur in the massive ore grades upward from foot- to hanging-wall.

The massive ore is quite fine-grained and is of a largely homogeneous texture; this is particularly true of the pyrite variety; the composition also is generally homogeneous as well with the sulfides being quite so but with the gangue minerals differing to a greater extent. The mineral compositions of the two massive ore varieties average:

Mineral	Pyrite Ore in %	Pyrrhotite Ore in %
Pyrite	55.6	2.5
Pyrrhotite	5.4	49.0
Sphalerite	17.4	19.3
Galena	6.1	6.5
Chalcopyrite	0.8	1.8
Gangue	14.7	20.8

The principal gangue minerals are: quartz 70% to 90% and mica 10% to 20%. Minor and accessory minerals are: plagioclase, chlorite, gahnite [$ZnAl_2O_4$], apatite, and tourmaline.

In more detail, Vokes measured the pyrite variations across the ore from foot- to hanging-walls, finding that one section showed a steady decrease, one was higher in pyrite in the center, one was lower in the center, but nine showed a steady increase from foot- to hanging-walls. For non-ferrous sulfides, the following relations obtained:

	Steady Decrease	Steady Increase	Higher in Center	Lower in Center	Irregular
Galena	7	1	3	0	1
Chalcopyrite	4	3	2	1	2
Sphalerite	3	1	3	2	2

So far as gangue minerals are concerned, nearly 60% of the sections showed a steady decrease, 25% showed a lower percentage in the center, and two of the remaining three showed eight per cent each.

In addition to these sulfides and gangue minerals mentioned above, arsenopyrite and magnetite are present in the massive ore in sufficient quantity to be noted; arsenopyrite is a prominent constituent in the disseminated ore in the hanging-wall, particularly in the southern half of the ore body. All of the sparse magnetite in the massive ore is in the pyrite variety and only locally is it abundant enough to be included in a statement of mineral percentages.

The range of the sulfide percentages in the pyritic ore is: pyrite - 30% to 85%; pyrrhotite - 0% to 15%; sphalerite - 10% to 30%; galena - 0% to 12%; chalcopyrite - 0% to 5%; gangue - 0.2% to 40%. Other minerals, in amounts less than 1% are: molybdenite, cassiterite, magnetite, tenantite-tetrahedrite, stannite, boulangerite, valleriite, arsenopyrite, gudmundite, breithauptite, bournonite, and ruby silver (proustite-pyrargyrite). Pyrite (and the minor arsenopyrite and pyrrhotite) is the only mineral showing crystal outlines; all other minerals are interstitial to the pyrite and are intimately intergrown with each other. Normally no banding is observed in the sulfide portion of the ore except at the thinning ends of the ore body; there banding is observed between the sulfides and the gangue minerals, the ratio between the two mineral categories changing quite appreciably across narrow distances.

In many places in the ore area, the schist bands have been folded, the folds having amplitudes of only a few centimeters or one or two decimeters at most. The folding does differ considerably in intensity from one place to another, locally being essentially isoclinal. In others, folds have been disrupted and the fragments detached one from the others and stretched out in the massive ore. The ore-forming process does not seem to have caused any appreciable disturbance of the fold axes in the schists, granted that this process took place after the folding. On the other hand, Vokes points out that these same relations would obtain if the folding took place before the introduction of the sulfides.

In places, local and irregular patches of vein quartz are found in bands generally parallel to the general strike and dip of the ore zone and of the enclosing rocks. These could be detached pieces of original quartz lenses, disrupted during folding or they might be locally formed metamorphic segregations of quartz or they could have formed by replacement during a hydrothermal stage of mineralization.

Vokes believes that the typical textures of the sulfide minerals - pyrite showing mainly crystal faces against the other sulfides - resulted from major recrystallization (under metamorphic stress) of the pyrite and the adaptation of the other sulfides (under the same stress) to positions around the pyrite crystals. In other areas, however, pyrite grains have been drawn out into lenticular masses that show essentially no traces of original crystal outlines. The development of these elongate grains, Vokes thinks, requires rotation of the crystals and additional deposition of galena or sphalerite on the corners aligned to the stress (i.e. in the stress shadows), the mineral so depositing depending probably on what was available. Such stress appears to have been greater along the margins of the massive ore or the massive ore was better able to resist this stress than the disseminated sulfides.

Sphalerite and galena usually are present in any given area in a ratio of about 3 to 1, so the sphalerite largely determines the textural relations between the two. The sphalerite normally is in an even-grained, allotriomorphic mosaic with no banding or other orderly arrangement. In such situations, the sphalerite generally has rounded grains, convex against the galena, the galena, therefore, being in a network of irregular and disconnected interstitial fillings and being concave toward the sphalerite. Pyrrhotite, such as is present in the pyrite variety of ore, usually is intergrown with non-ferrous sulfides, particularly chalcopyrite, but retains convex outlines against the other sulfides. Pyrrhotite appears to have replaced both pyrite and magnetite to a minor degree. This phenomenon almost certainly was the result of solid diffusion, but the extent to which actual fluids may have been involved is not known.

In general, the Bleikvassli pyrite ore variety does not show late cataclastic breaking after sulfide crystallization with such tectonic disturbances as occurred taking place before and during the final crystallization of the ore. A few widened cleavages and cracks in pyrite do exist and have been filled by the sulfides of the matrix. No evidence has been seen by Vokes of such veins cutting the matrix sulfides, only the pyrite is so treated because it was by far the most brittle mineral present.

As for the pyrrhotite-rich ore at Bleikvassli, Vokes reports two types: (1) the massive type of the same general character of the pyrite variety and (2) irregular bodies (mainly toward the ends of the ore body) that are high in quartz. In this second type, the sulfides occur in a loose, sugar-grained quartz gangue that is contained in heavily crushed zones of schist. The non-ferrous sulfides are erratically distributed in this type, and the chalcopyrite content is much higher than normal. Vokes suspects that these masses of the chalcopyrite-rich type are the result of selective remobilization of massive pyrrhotite-rich ore on a local scale.

The massive, pyrrhotite-rich ore normally is present on the footwall of the ore lenses; nevertheless, this ore is an integral part of the massive ore.

No structural break is known between the two types, and the change occurs by the pyrite grains of the pyrite type becoming more widely separated and pyrrhotite forming the dominant sulfide in the matrix between them. The texture of the pyrrhotite-rich ore is not uniform, and modal analyses of this type show a wide range as follows:

Mineral	Average %	Range %
Pyrite	2.5	0-12
Pyrrhotite	49.0	33-63
Sphalerite	19.3	9-24
Galena	6.5	3-9
Chalcopyrite	1.8	0.5-3
Gangue	20.8	11-39

This type also contains minor molybdenite, tennantite-tetrahedrite, and stannite. Supergene marcasite is present. Trace minerals are: arsenopyrite, boulangerite, gudmundite, and breithauptite. The proportions of non-ferrous sulfides and gangue minerals are somewhat different in the two ore types, but the ferrous sulfides, of course, are widely different. Chalcopyrite is about twice as abundant as in the pyrite type.

The footwall band of pyrrhotite-rich ore is not completely continuous along the ore-body footwall; in the thicker parts of the ore body, the pyrrhotite ore may be one to two m thick (of a total width of 10 to 15 m), but there is no general and firm relationship between total band thickness and that of the pyrrhotite layer. In the central part of the mine, there appears to be a thin lens of pyrrhotite ore that may be as much as 2 m thick but is definitely quite different in thickness from one area to another. The mineralogical composition of the pyrrhotite ore is much more varied than that of the pyrite type and has the appearance of a breccia (as Vokes says, without any genetic significance being applied to that term). The broken fragments are both schist and quartz; these fragments are essentially always rounded, their shapes being elongated or ovoid (for schist) and spheroidal (for quartz). Vokes believes that these fragments were subjected to a rather violent tectonic action that rolled them sufficiently to account for their highly polished surfaces and the distortion of their schistosity. Whether all this action, plus some penetration of the quartz fragments by sulfides, took place after the sulfides had been inserted in the Bleikvassli rock environment or whether the sulfides were introduced by replacement late in the cycle of sedimentation-metamorphism Vokes is uncertain. Some movement of the sulfides occurred after they had been introduced, but he does not find definite evidence that they were present through the whole of the sedimentary-metamorphic process or not.

The principal texture of the pyrrhotite ore is a fine- to medium-grained intergrowth of major pyrrhotite with minor chalcopyrite and even lesser galena and sphalerite. Certainly pyrrhotite is the only mineral that exhibits any evidence of having undergone stress. Either this means that the entire assemblage was subjected to at least some stress, the effect of which only pyrrhotite shows, or only pyrrhotite was present when the stress was applied and the non-ferrous sulfides were introduced after the stress had been relaxed. If the latter scenario is the correct one, the non-ferrous sulfides can have been emplaced only by the replacement of a gangue of quartz and schist fragments of various degrees of fineness and compaction. The final verdict is not yet in.

The contrast between the massive ores at Bleikvassli and their wall rocks is normally sharp. In many places, however, the hanging-wall schists contain various proportions of scattered sulfides with pyrite being the most abundant of these and being in elongated crystals up to 1 cm or more in their longest dimension. The other sulfides are distributed with varied regularity between the pyrite crystals contained in the schist.

In addition to the variety of scattered sulfides just described in the hanging-wall ore, veins, patches, and impregnations of sulfides are, mainly, in the rocks on the hanging-wall side of the ore body. With one exception, all the minerals in this ore type have been identified above; this exception is geocronite [Pb_5SbAsS_8].

Vokes (1966) outlines his ideas as to the two most likely concepts to explain the genesis (manner and time of emplacement) of the Bleikvassli sulfide ores.

Assuming an epigenetic emplacement, the development of schistosity must be assigned to shear movements that preceded the deposition of the sulfides and produced the structure that now hosts the ore. The essentially concordant schist bands now in the ore could be considered as remnants of a schistose body that must have extended from one wall to the other of the present ore body. The schistose structure would have served as an excellent channelway to upward and outward movement of the ore-forming fluids, whatever they were and wherever they came from. Some deformation, though certainly not the major portion that at least the host rocks of the ore body have undergone, would have had to take place after the sulfides had been introduced into the schistose rock. The portions of the schistose rock remaining between the hanging- and footwalls of the ore body would be those portions of that zone that were least susceptible to replacement. If this scenario is correct, Boyle (quoted by Vokes, 1966) would be right when he said that: "the massive sulfides are contained in [a] conformable structure and this is the only reason for referring to the [ore body] as conformable. If these assumptions are correct, the ore body at Bleikvassli could be considered to be a metamorphosed, epigenetic sulfide ore body.

Vokes says that an epigenetic derivation for the Bleikvassli galena (at least possibly) would explain the apparent ages derived from Pb-isotope determinations are such (80±70, 160±60 and 160±80) by an addition of radiogenic lead to normal lead. These various ages, as Vokes says, are not compatible with the conformable class of geosynclinal deposits as defined by Stanton and Russell (1959). Such an addition of radiogenic lead could have been made during the movement of ore-fluids toward, and into, the shear structure assumed to have been developed late in the Caledonian metamorphic cycle.

On the contrary, Vokes points out that it can be argued from the fact that a remnant of layered structure can be detected in the microcline gneiss body that the ores were emplaced epigenetically. In particular, bands of white and gray quartzite can be followed from the schists in the south far into the area of microcline gneiss, and these bands maintain a rather constant position relative to the footwall of the ore zone. The quartzite becomes progressively better-defined as it is traced northeastwards into the center of the gneiss body; in this place, the gneiss loses its identity. It is Vokes' opinion that these quartzose bands have the appearance of remnants of a pre-existing layered sequence that was almost entirely changed by the production of the microcline gneiss. Resoning from this concept, he believes that the sulfides themselves are the remnants of an earlier-existing sulfide-rich unit in this layered sequence that has been preserved to an appreciably greater extent than the quartzitic layers. Vokes does not say, although he points out the alternatives, which of the two concept he prefers, but he does comment that it does not matter, for his then current purposes whether the sulfides were emplaced syngenetically with the original sedimentary series or by later processes (diagenesis or other). Further, the embryo ore bodies may have consisted of iron sulfides only if they were so produced.

He does, however, believe that subsequent deformation and metamorphism during the Caledonian revolution would have been capable of producing the present ore bodies as they are now known.

The epigenetic hypothesis requires: (1) folding and metamorphism of the original sediments and the production of the "remarkable" microcline gneiss by intrusion or metasomatism; (2) shearing movements to produce the host-rock structure across the junction between the microcline gneiss and the schists adjacent to the gneiss on the south; (3) the entry of a hydrothermal ore fluid to deposit the sulfides, and (4) a later metamorphic event to deform the sulfides and their neighboring and surrounding schists.

For the syngenetic scenario, the sulfide lens would have been part of the original sedimentary sequence before any metamorphism or folding. The later geological events would have included the folding and metamorphism of the sulfides and their enclosing rocks, the formation of the microcline gneiss (by a process that left it partially enveloping the ore zone), and later shearing along certain zones that included the ore walls.

Ramberg (1964) has shown that the rocks were subjected to two metamorphic events; the first was characterized by tight, isoclinal folding along a generally east-west axis with metamorphism that reached the lower amphibolite stage. The second event was a less intense metamorphism that reached only to the lower epidote-amphibolite facies, the accompanying folding being more open. Ramberg also thinks that the crystallization of the microcline gneiss occurred during the second metamorphic event at a temperature that he thinks was probably not more than 250°C. The area around the ore body was a zone in which differential movements took place in which the footwall rocks (to the east) were displaced to the north, relative to the overlying rocks to the west. The shearing that affected the Bleikvassli rocks and the ore body itself, he refers to these differential movements.

He further believes that metamorphic recrystallization continued after the main deformation period as is shown by the lack of cataclastic textures in the ore, particularly, and by the presence of undeformed and metablastic pyrite in the highly stirred-up pyrrhotite-rich zones. In short, the second phase of the deformation, because it was affecting rocks that had been subjected to a higher phase of metamorphism than it (the second phase) could produce, left little detectable evidence of its occurrence on the higher-grade phase and particularly on the sulfide minerals in the ore.

Vokes thinks that the epigenetic sequence is much less easy to fit into the geologic history of the Bleikvassli district, but he cannot, he says, exclude it at his 1966 state of knowledge. Only through the epigenetic scheme can the anomalous leads be fitted reasonably into the geologic picture as this could have been done while the hydrothermal fluids were making their way toward the area of ore deposition and were their depositing the Bleikvassli ores. This seems more reasonable to me than that the anomalous lead could have been added to the Bleikvassli ores long after they had been laid down as sediments and in such complete conformity with the ore body as it then existed.

Obviously, Vokes thinks the syngenetic production of the original sedimentary sulfides is the more likely of the two possible explanations; my opinion is exactly the opposite. Vokes has spent years in working on the Bleikvassli deposits; I have never seen it. Nevertheless, I believe it more probably is a metamorphosed epigenetic deposit than that it is a metamorphosed syngenetic deposit even more highly metamorphosed, but the final verdict certainly is not in.

In a somewhat more recent paper (Sen and Mukherjee, 1973), the authors conclude that the rocks at Bleikvassli have suffered two phases of folding: (1) F_1 in which tight isoclinal folding took place and (2) F_2 that produced a set of disharmonic kink folds that finally became nearly the same in geometry as those of F_1 (but with an axial plane cleavage in places nearly co-axial with F_1). The F_2 stage contained a shear component as its extreme development that rotated the noses of the earlier mesoscopic folds. The "stirring up" of the ores described by Vokes (1963, 1966) is ascribed to this F_2 shearing component.

These authors assume that the original layering (S) was emphasized as S_1, the regional bedding schistosity during the F_1 folding phase. The axial plane of the F_1 folds is parallel to S_1. F_2 folds produced an axial-plane cleavage S_2, that only shows restricted recrystallization.

These two authors explain the local truncation of the ores by local disharmony in disharmonic kink sets of F_2 folds (with the ore generally participating in this folding).

The schistosity, S_1, is emphasized by quartz, plagioclase, epidote, biotite, muscovite, and amphibole along its orientation. Thus, these minerals are pre- and syn-kinematic in relation to the F_1 folds. The superposition of the F_2 folds resulted in recrystallization and elongation of these minerals, plus the development of some new minerals. The amphiboles have two generations of crystallization, along S_1 and growing across the re-emphasized S_1 and post-dating it.

The degree of recrystallization that took place during the F_2 folds differs importantly from place to place. Garnet probably had two periods of crystallization. These authors recognize at least five varieties of biotite, most of which were due to F_2 folding. Staurolite was developed after the mica; garnet is earlier than staurolite. Staurolite crystallization began in the late F_1 folding and continued through the F_2 stage, even later it was rotated syntectonically. Some staurolite has been affected by retrograde metamorphism, being converted to sericite.

Kyanite in spots appears to be parallel to F_1, whereas bladed kyanite probably is synkinematic with respect to F_1 (later than biotite). Sen and Mukerjee believe that the euhedral pyrite may be an example of mimetic recrystallization (i.e. reproducing any pre-existent anisotropism, bedding, schistosity, or other structures). The various degrees of recrystallization have been charged to differences in the intensity of local deformation and to non-cylindrical structural domains.

Their opinion is that the main metamorphic minerals and the ore minerals crystallized during the F_1 folding and recrystallization continued through the F_2 stage. They consider, however, that pyrite recrystallization overlapped that of all the other minerals and continued into the late F_2 period. They give the following table to show the relationships of mineral growth to periods of folding:

Mineral	Periods of Growth
Garnet	2 periods of growth (1) F_1 (2) F_2
Kyanite	2 periods of growth (a) F_1 (2) F_2
Biotite	2 periods of growth (a) F_1 (b) late to post F_2 (patchy biotite)
Pyrite	2 periods of growth (a) F_1-F_2 (b) late F_2, overlapping other ore minerals

Note: Kyanite, sillimanite, and staurolite align parallel to S_1 and overgrow it; they are suggested to have crystallized in F_1 and through F_2.

They think that the ore minerals began to form (do they mean recrystallize?) during the F_1 phase of folding and were recrystallized and elongated during the F_2 phase.

Assuming that this order of events is correct, the ore minerals (or the material from which they were produced during metamorphism) had to be introduced into the rocks before any folding had taken place. If this is correct, then it tilts the scales in favor of a syngenetic manner of formation of the ore materials in the deposit. If, however, the first introduction of the ore minerals was after the first stage of folding, and they replaced rocks that had inherited the structures (produced by F_1 folding) in the rock minerals, it would be difficult to say whether or not the ore minerals were pre- or post-F_1 folding. Obviously this matter needs further consideration.

If the parents of the ore sulfides in the Bleikvassli district were deposited syngenetically or diagenetically, they would be classified in the modified Lindgren classification as "Deposits chemically concentrated in quiet waters by introduction of gaseous igneous emanations and water-rich fluids", or IIA3. The possibility of their having been introduced as part of the sedimentary cycle by precipitation of metal ions through the agency of hydrogen sulfide formed by bacterial action on organic matter seems remote.

If it were possible to consider the pyrite (and pyrrhotite?) as having been introduced syngenetically and the non-ferrous metals added later (after the rocks of the area had been lithified, but before the F_1 stage of folding), this would provide a more acceptable explanation.

I am, however, of the opinion that the concept of hydrothermal fluids invading the rocks of the Kongsfjell group after pyrite had been deposited syngenetically (or essentially so) and the non-ferrous metals after the pyrite had been metamorphosed during the F_1 folding is possible but not probable I still favor, nevertheless, the idea that the sulfides were brought in in their entirety after the rocks had undergone the F_1 folding and that the structures of the pyrite was largely due to the F_2 folding. The bedding schistosity (S_1) is parallel to F_1 folding, so pyrite replacing the rocks after the F_1 folding had taken place would have the same alignments as the minerals they replaced. The recrystallization of the pyrite in the F_2 folding phase would have been due to the same folding that produced the textures present in the sulfides that provide the matrix around the pyrite crystals. Both the pyrite and pyrrhotite ores contain fragments of schist (Bugge, 1978) which he thinks were enclosed in the sulfides during metamorphism. These schist fragments could equally well be thought of as replacement remnants of schist that had escaped metasomatism during the period between the F_1 and F_2 folding.

On the basis of this concept and considering the minerals present in the Bleikvassli ores, it seems best to classify them as Hypothermal-1 (in non-calcareous rocks) to Mesothermal, and this is done here.

RØROS-KVIKNE-KILLINGDAL AREA, SØR TRØNDELAG

Middle Paleozoic	Copper, Zinc, Pyrites	Hypothermal-1 to Mesothermal Metamorphic-C

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Notes

The Trondheim region of which the RØros-Kvikne-Killingdal area is a part, has been the largest and most important copper producing area of Norway. It contains many more or less strata-bound sulfide deposits; those outside the RØros area here under discussion include: LØkken, Folldal, Undal, and Tverrfjellet. The RØros-Kvikne-Killingdal area lies in the southeastern part of the Trondheim region with the town of RØros being located at 62°35'N, 11°23'E; Kvikne at 62°36'N, 10°19'E; and Killingdal at 62°47'N, 11°26'E.

The five mines in the RØros area for which appreciable geologic information has been published are: RØros proper, Kvikne, Killingdal, RØdhammeren, and KjØli. Three of these mines (RØros, Kvikne, and Killingdal) are at the apices of an isosceles triangle. The line connecting Kvikne with Killingdal is the hypotenuse and strikes at about N60°E. The distance from RØros to Kvikne and to Killingdal is about 35 km in each instance; that from Kvikne to Killingdal is about 55 km. KjØli is about 15 km N55°E of Killingdal and RØdhammeren 17 km N13°W of Killingdal.

In the RØros area, the individual mines fall into two areas; (1) in the eastern field, 10 km northeast of RØros, the principal mines are: Hestkletten, Christianus Quintus, Storwartz (the most important mine), and Kronprins Olav and (2) in the northern field, 10 to 15 km northwest of RØros, are: Leirgruvebakken, Kongen, RØdalen (probably a strike continuation of Kongen but offset by a fault), Sextus, and Mugg. Leirgruvebakken was the last mine to cease operation.

In the Kvikne area, the oldest mine was Prestens Gruve (now the upper part of Segen Gottes Gruve); the Gabe Gottes mine is some hundred meters south of Segen Gottes. Further prospecting found such mines as: Vansgruvene, Banken Gruve, Gruve (no modifier). Odden, Odden II, and Dalsgruven were the last to be found and worked in the Kvikne area. The Røstvangen mine (62°23'N, 10°22'E) is 25 km SSE of Kvikne parish and is considered with the Kvikne mines.

In the Killingdal area, only the mine of that name has been studied in any detail. In the Kjøli area, only the mine of that name was worked on any appreciable scale; others of minimal importance were located south-southwest and north-northeast of Kjøli. The Rødhammeren mine is 17 km N13°W from Killingdal.

Copper mining began in the Røros area proper with the discovery of the Storwartz mine in 1644 under the direction of the Røros Kobberwerk A/S and continued until 1977, when the last mine was closed. During the more than 300 years of operation, the company produced about 35,000 tons of copper from the Storwartz mine; the Kronprins Olav produced about 13,000 tons of copper from 1,000,000 tons of ore, in the period following its discovery in 1935. The Leirgruvebakken mine was found by an airborne electromagnetic survey in 1934, aided by later ground geophysical measurements and diamond drilling, and its reserves at that time were estimated as 1 million tons of 0.74 per cent copper and 7.8 per cent zinc ore. This mine was the last in the area to close. In 1975 production was 6680 tons of zinc and 1495 tons of copper concentrates. The Mugg mine was operated from 1717-1919, the Rodalen mine from 1924-1945, and the Kongen mine from 1657-1933.

The Killingdal mine (Rui, 1973) produced more than 2.5 million tons of ore through 1973 that averaged 1.7 per cent copper and 4.5 per cent zinc; the mine was discovered in 1674. Although it originally belonged to the Røros Kobberwerk, it sold in 1891 to the Killingdal Gruveselskab A/S that have operated it, except for the period 1895 to 1946 when it was rented to an English company. The last reported production figures are for 1975 was 50,000 tons of ore from which 2600 tons of copper concentrate, 2650 tons of zinc concentrates, and 9000 tons of iron-sulfide concentrates were recovered, plus some galena and arsenopyrite. So far as I know, production still continues, but all I know for certain is that it operated through 1977.

Ore deposits in the Kvikne area were first discovered in 1629 with mining beginning in 1631, 13 years before mining began at Røros. Production continued, with numerous interruptions, until 1812 when all of the various, but small, mines were closed. Prospecting and mining did continue sporadically on a small scale after this date and until World War I. In the 180 years before 1812, 7343 tons of copper were recovered; how much was produced since 1812 is unknown, but the amount must have been small. Nine mines in the Kvikne area were important enough to be mentioned by name by Bugge (1978).

The Røstvangen mine (see above for location) was discovered in 1905 and work started there in 1908. Mining ceased there in 1921. Røstvangen is reported (Rui, 1973) to have been "one of a number of rich, but small massive chalcopyrite-bearing pyritic ore deposits known from the Kvikne district". Rui gives no production figures for Røstvangen or the small mines associated with it. In addition to Røstavangen, mining also was conducted at the adjacent Finnhaug, Hamdal, and Børsjøhø mines about 5 km WNW of Røstvangen, and apparently they are of the same type.

In the Kjøli area, Rui and Bakke (1975) report that a great number of small, now-abandoned mines and prospects mark out a major area of abundant sulfide mineralization that extends southwest from Tydal (63°04'N, 11°33'E). Only the Kjøli mine appears to have been of any economic importance; Rui and Bakke (1975) report that its production was about 250,000 tons; this mine was found in 1766 and was worked intermittantly, being abandoned for the last time in 1941. All the remaining mines probably did not recover more than 15,000 to 20,000 tons of ore with about half of it coming from Svensk Menna.

The grade of the Kjøli ore probably averaged a little over 2.0 per cent copper, traces of zinc, about 46 per cent sulfur, and about 45 per cent iron. The remaining mines (Mitgruben Godthåb, Røros Menna, and Svensk Menna had about the same grade, except that Røros Menna appears to have run over 7 per cent zinc.

The Rødhammeren mine (62°56'N, 11°25'E) is part of a belt of sulfide deposits in the eastern Trondheim region and is located 1.5 km north of the valley of the Holta River. The Røros deposits, by far the most productive along the belt, had infinitely more production than came from Rødhammeren; what ore was mined at Rødhammeren ran from 0.1 to 0.5 per cent copper and 35 per cent sulfur.

Bugge (1978) gave a most detailed stratigraphic table for the Røros district, and he points out that strata-bound deposits occur at six different stratigraphic horizons in the Cambrian through Silurian sequence known in the area. These rocks are mainly metasediments and metavolcanics. It is possible that these rocks have been overturned and belong to a major and recumbent nappe that has its roots at a considerable distance to the west. If this is the case, Bugge's stratigraphic sequence that follows should be overturned. This overturning has been suggested by several authors - Wolf, 1967; Nilson and Mukherjee, 1972; Rui, 1972; Rui, 1973. If this is true, then the rocks are successively younger to the east. Bugge (1978) considers that this "overturning" is a reasonable subject for controversy and suggests that radiometric dating and diagnostic fossils, if and when found, would help solve this problem. Bugge, however, points out that graded bedding in several places supports the idea that the beds are *not* overturned as does correlation with other, more northerly parts of the Norwegian Caledonides.

Thus, the stratigraphic sequence given here is in the order favored by Bugge, but it must be examined with the caveat that it may be necessary to turn it upside down. The formations, then, from bottom (in Bugge's sense) to top are: (1) Hummelfjell, feldspathic quartzite with meta-dolerite dikes - perhaps late Precambrian; (2) Røros formation - four subdivisions, bottom to top, meta-dolerite sills and dikes; thin beds of graphite schist and marble, serpentinite, calcareous mica schist; (3) Røsjo, metagraywacke and mica schist with thin intercalations of greenschist; (4) Fjellsjø, graphite schist; (5) Saetersjø, thin beds of greenstone and keratophyre under meta-arenites and polymict conglomerate; (6) Kjurudal, mica and graphite schists with beds of greenstone; (7) Hersjø, basal polymict conglomerate, then interlayers of pyroclastics and green sediments with siliceous banded sediments; meta-dolerite sills and dikes; layered gabbro (and trondheimite), finally greenstone-keratophyre; (8) Gula group, quartz conglomerate (lowest), *dictyonema* schist, trondheimite (in part semi-stratiform) and metagabbro; mica schist, kyanite-staurolite schist, graphite schist, greenstone and peridotite, calcareous mica schist and meta-arenite (on top). Obviously, if the sequence is overturned, these beds should be listed in the reverse order from that given here.

The various, normally strata-bound deposits in the general area are distributed rather widely through the stratigraphic sequence. In the Røros formation are the Storwartz, Lossius, and other ore bodies; in the Røsjø formation, the Kongen, Sextus, Mugg, Leirgruvebakken mines (Røros area); Kjøli, Menna mines (Kjøli area); in the Kjurudal formation, Hesjedalen mine (to the west of the Kongen mine); in the Hersjø formation, Løkken mine (southwest of Trondheim), Tverrfjellet mine (southwest of Kvikne), Killingdal mine (see above), Hersjø mine (some 10 km west of Sextus mine), Rodhammeren mine (some 17 km slightly west of north of the Killingdal mine), Gula group, Kvikne mines, Røstvangen. This scattered distribution throughout the stratigraphic sequence of the Røros, and continguous, districts means that, if the ores are of exhalative-sedimentary origin, the process of sedimentary ore formation took place at least six different times from the Cambrian to the Silurian. If, on the contrary, the ores were deposited in lithified rocks by hydrothermal fluids, they may have been emplaced all during one major period of mineralization or they may have been (as seems less likely to me) have been the result of a successive series of invasions by hydrothermal fluids entering the rock volumes in question between Cambrian and Silurian time. This problem is well addressed later in this discussion.

The geological environment in which the Røros and adjacent ores was formed was one of spatially related volcanic and sedimentary rocks that were laid down in the geosynclinal stages of the Caledonian orogeny [Rui, 1972, 1973 (both papers), and Rui and Bakke, 1975]. The volcanic rocks (now mainly greenstones and quartz keratophyres) were introduced into the stratigraphic sequence by a succession of eruptive episodes that began in the (probably late) Precambrian and continued until well into the Silurian. The sediments now are principally meta-argillite and meta-arenites of major thickness.

Rui and Bakke (1975) have added two stratigraphic divisions to the six proposed by Rui (1972) so that the eight divisions correspond to those given by Bugge (1978). The major difference between the two stratigraphies is that the one is in the reverse order from the other. The sedimentary rocks of the Røros area have been intensely folded, metamorphosed, and overthrust during the Caledonian (Silurian) orogeny. Before and during this orogeny, the sedimentary rocks were in places intruded to a considerable degree by pre-metamorphic gabbros and younger syn- and post-metamorphic trondheimites.

The major structure of the district is the Seve-Köli-nappe. To the northeast of Røros, at a distance of about 35 km, is a zone of intense mylonitization that marks the base of the Trondheim/Seve-Köli nappe complexes where they ride up over the Eocambrian and Precambrian rocks of the Sylane window. The Cambrian rocks in the Røros district are the lower, recumbent limb of that thrust, and (granted that Rui and Bakke and Nilsen and Mukherjee are correct) the west-dipping rocks in the Røros district were overturned in the process of nappe-formation.

A second thrust fault has been identified to the north of the Røros area proper and is designated by Rui and Bakke as the Tydal thrust. This thrust has been followed south to near the Sextus mine in the northern Røros field. North of the Sextus mine, the thrust divides the Kjøli area into two structural units in which the expressions of the tectonics that affected them are quite different. Above the Tydal thrust, the Gula and Hersjø units have been deformed into tight isoclinal folds and exhibit quite steep, west-dipping schistosity and bedding with the axial trends of the regional folds being parallel to the axes of minor folds; in turn, these minor folds are parallel with the elongation direction of the deformed conglomerates of the Gula group.

East of the Tydal thrust, the beds lie more flatly, are gently undulating, and have dips normally between 15° and 20°. Minor folds are usually more open than to the west of the fault-front with axial trends ranging from NW-SE to NNW-SSE and turning to north-south farther north. The thrust zone is well exposed in several places and is marked topographically be a steep wall. This zone is composed of dense mylonites and ultramylonites. Rui and Bakke can recognize the discordant character of the thrust zone by its effects on the phyllites of the Kjurudal and Fjellsjø formations in its southern portion, nearest the Røros area proper.

Below the main thrust (the Sevi-Köli nappe), mylonitic zones are common, either in the Røros formation or at the border between those beds and the Røsjø formation. The zones form topographically low ridges that may be several kiometers long. The contact between the Røros and Hummelfjell formations normally is characterized by strong deformation and disharmonic folding (Rui and Bakke, 1975).

The Caledonian intrusive rocks in the Røros district include: (1) equigranular and porphyritic meta-diabases that occur mainly in the western portion of the area, above the Tydal thrust; (2) saussuritized gabbros that are most common in the Røros formation; (3) dark amphiboles that are confined to the Hummelfjell formation, and (4) scattered occurrences of trondheimite that are either syn- or post-metamorphism.

The meta-diabases are mainly in swarms in the Gula group but also are present to a lesser extent in the Hersjø formation; they are known in the Køli mine area below the Tydal thrust. These fine- to medium-grained rocks normally are found in sills that are from a few decimeters up to 10 to 15 m in thickness; discordant sheets are rare. The only important difference between the two rock types is the presence of major quantities of phenocrysts in the porphyrytic ones; these phenocrysts may be as much as 10 mm in diameter but usually are 5 mm or less. Chilled margins are common. The average mineral composition is hornblende 49 per cent, plagioclase 48 per cent, zoisite-clinozoisite 5 per cent, 2 to 3 per cent sphene, and minor chlorite and biotite.

The saussuritized meta-gabbros are in conformable sheets and lenses of fine- to coarse-grained rocks; normally they are massive but exhibit fine-grained schistose contacts; these rocks are most commonly in the Røros formation. The mineral content of the meta-gabbros is essentially that of the meta-diabases, but the plagioclases everywhere have been altered to aggregates of albite/oligoclase and clinozoisite or are largely recrystallized to a granoblastic mosaic. It still is possible to discern remnants of ophitic texture. Locally, a few ultramafic bodies (now much altered to mixtures of antigorite and talc) occur in the meta-gabbros; pseudomorphs of antigorite after talc are readily identified.

The amphibolites that are found in the Hummelfjell formation differ from the meta-gabbros in being darker in color although they locally show a lighter banding. The amphibolites are much more schistose and exhibit slender parallel prismatic crystals in thin section. These dark rocks are almost entirely made up of sub- to euhedral hornblende and plagioclase (An_{20-40}), normally in nearly equal quantities. As accessories, some epidote/clinozoisite is present with minor carbonate, chlorite, and biotite. In places, the rocks may contain as much as 80 to 90 per cent of the silicates as hornblende, but the lighter bands can have as much as 80 per cent of plagioclase.

These amphibolites produce high aeromagnetic anomalies, this separating them from the less iron-rich mafics.

The trondheimites appear as scattered bodies in the northern part of the Røros formation or above the Tydal thrust. They are thin, concordant sheets up to a few meters in thickness that cross the formations in which they are enclosed. The rocks are light in color, and the major minerals are sodic plagioclase and quartz with lesser muscovite, biotite, chlorite, and epidote/clinozoisite. Farther north, Rui and Bakke (1975) report that the structures of the trondheimites favor a syn- to postmetamorphic age for the trondheimites.

The ore bodies of the Røros area definitely are strata-bound, but various hydrothermal phenomena are shown and small veins that appear to be epigenetic are common. The ore bodies are most prevalent in volcanic greenstones and keratophyres but also are present in graywackes and mica schists. Pyroclastic material has been identified in the metasediments of the Røros formation, but some of it is present where no ore is observed. Some of the ore bodies in the east part of the district are contained in the Hersjø greenstones, such as the major Killingdal deposit. This greenstone is thought, by Rui and Bakke (1975), to be lower Ordovician, but this depends on the validity of the concept that the beds in the district are overturned.

In the eastern mining field at Røros, about 900 m above sea level, several mines, now worked out but once of much importance, were present in the Røros formation and underlay the Hestkletten meta-gabbro. One ore horizon is immediately below the gabbro (Hestkletten and Christianus Quintus mines); another is about 60 to 80 m below the gabbro [Storwartz (the most important mine) and Kronprins Olav]. Also included in the eastern mining field are the old and new Sunshine mines that are in a higher horizon than the Olav mine with the former overlying the latter. The ore in these mines is in more or less irregularly horizontal layers that are 0.5 to 12 m thick; these parallel the planes of schistosity and the schist-gabbro contact. The ore lenses are up to 300 m wide and quite irregular in outline. Zinc ore lies outside the zones of copper ore with copper ore to the west being in pyrite and to the east in pyrrhotite. Much chlorite accompanies the ore, but it has been largely affected by tectonic forces as witnessed by its schistose character. The Sunshine mines and the Olav below them are cut off to the west by a west-dipping fault; the old and new Storwartz mines (to the west of the Olav) are apparently on the same horizon. The Sunshine mines outcrop at the surface and are separated (?) by an eroded interval. If the fault cutting off the Sunshine and the Olav at their western extremities is a reverse fault, then the Olav possibly is the equivalent of the Storwartz and the Sunshine of the Quintus and Hestkletten (the latter two having been joined by mining), although each outcrops on the east and west side of the hill topped by the gabbro sill. In the southeast part of the Olav mine, a fault raised the ore about 10 m and a second dropped it back down again.

The northern mining field at Røros contains at least four mines arranged roughly in a north-south line; from south to north, these mines are: Leirgruvbakken (also spelled Lergruvbakken), Kongen, Sextus, and Mugg. The Fjellsjø and Hesjø lie more or less west of the Sextus mine with the Fjellsjø being the nearer to Sextus. All of these mines are in the graywacke and mica schist of the Røsjø formation. The Leirgruvebakken mine is at an altitude of 860 m and consists of seven sheet-like bodies in graywacke and chlorite schist, but only two of these were of mineable grade and size. Chalcopyrite, sphalerite, and pyrite are contained in a matrix of pyrrhotite. The ore has been brecciated and includes fragments of the folded mica schist. Most of the other mines (worked before the Leirgruvebakken) were rod-like with their long axes being parallel to a prominent tectonic axis. The Mugg mine had an ore length of 1500 m with a width of 110-160 m and a height of 0.2 to 2.0 m. The Kongen mine had a mineralized length of 2500 m, a width of 50 to 120 m, and a height of 1 to 4 m; the ore body plunged west at 10°. The Rødalen mine was 600 m long and is separated from the Kongen by a fault but the one appears to be a continuation of the other. Still farther west, a geophysical anomaly probably indicates a continuation of the Kongen-Rødalen pair to the west.

Although most of the cited publications on the Røros-Kvikne-Killingdal area are the work of the "Røros Project" of the Institute of Geology of the University of Oslo, Blindern, Norway, I have no recent work on the geology of the Røros area per se. Much of what I said derives from my own notes taken in 1954, so the discussion of the eastern mining field comes from them; the material on the northern field comes from Bugge (1978).

The Killingdal mine (ca. 62°48'N, 11°30'E) is about 25 km north of Røros. The mine is one of the largest of the numerous massive-sulfide deposits known in this district; production from it puts it in the same class as the larger mines of the Røros district proper. The rocks of the Killingdal area are eugeosynclinal rocks formed during the Cambrian through Silurian periods; these rocks now are made up mainly of metamorphosed greenstones of submarine volcanic genesis and of a variety of schist-types that originally were argillites and arenites. These beds are intruded in many places by what are now amphibolites but initially were gabbros. Some younger silicic rocks, mainly quartz diorites, are spatially related to both the metavolcanics and to the intrusive amphibolites as well. The Cambro-Silurian formations involved include members of the Gula group, the Hersjø formation, the Kjurudal formation, and the Røros schists. A thrust plane, the Tydal fault, lies along the contact of the Kjurudal formation (below) and the Røros schists (above). At Killingdal, the Gula group is composed of non-calcareous gray phyllites and mica schists with some conglomerate. The Hersjø formation is mainly metamorphosed basalts, now termed green hornblende schists; the Kjurudal formation is composed of calcareous light to dark phyllite; and the Røros schists are calcareous and gray and initially were argillites and subgraywackes. All of these rocks are quite well-bedded, layers usually ranging between a few centimeters and several decimeters in thickness.

The Killingdal region is one in which the concept that the formations in the area have been overturned generally is accepted. The principal structures of the area are caused by two generations of folds. The interactions of these two fold systems are well shown by the complex pattern exhibited by the Hersjø formation in which the greatest complexity is that in the Killingdal area proper. The early F_1 folds are isoclinal and initially had axial trends striking at about NNE-SSW. The primary bedding surfaces of these rocks (S_0) normally are parallel, or nearly so, to the secondary schistosity (S_2) that is much the more prominent of the two and is axial planar to the F_1 folds. The F_2 cross-folds are more open than the F_1 that they deform. The thrust faults stand out in relief because of the dense, dark mylonites developed in zones along them; the mylonites probably were produced in late stages of the F_1 folding, movements on the faults being east or ESE. During the F_2 folding, flexures were formed on the thrust planes.

The Killingdal ores occur in two ore bodies, Main and North; both are markedly elongated and have lens-shaped cross sections. These ore bodies are parallel for their known length of 2500 m and have a mean axial dip of 30°W. The North ore body lies above and somewhat to the north of the Main one. In 1973, the mine was developed down to 1400 m below the surface (500 m b.s.l.). In the higher levels of the Main mine, the ore body normally is between 40 and 80 m in width, and its thickness averages about 3.5 m with a maximum of 10 to 12 m. As this ore body is followed downward, it narrows considerably and finally degenerates into impregnated schists. As for the North ore body, its cross section is much less and mining on it has been much less impressive.

The massive ore is concordant with the schistosity of the adjacent wall rocks; in many places, however, the ore splits into separate, parallel bands that grade into the impregnated schist of the enclosing wall rocks. In detail, the structures of the ores are tight folds that affect both the ore and the wall rock; these folds are overturned to the north and the reclining fold axes coincide with the major axes of the deposits, that is, they plunge west, following the dip of the surrounding schists. The ore is concentrated in fold crests and troughs; the problem of the time relations between ore formation and folding will be discussed later. In the north end of the Main ore body, isoclinal folds are repeated to form strongly compressed fold hinges. In such areas, the wall rock is brecciated and contains highly irregular veins that are enriched in copper.

The minerals in the Killingdal ore are, in order of abundance: pyrite, sphalerite, chalcopyrite, pyrrhotite, galena, arsenopyrite plus accessory tetrahedrite, bournonite, mackinawite, molybdenite. The first three of these minerals account for over 90 per cent (by weight) of the ore, and pyrite is by far the most common ore mineral. The first minerals to deposit were fine-grained sub- to euhedral pyrite and scattered grains of arsenopyrite. These grains determine the texture of the ore in that the other sulfides are molded around them. The pyrite and arsenopyrite are quite brittle and may be broken into subregular to most irregular networks of fractures; these are occupied by the minerals later than the FeS_2 and FeAsS. Rui (1973) is of the opinion that post-ore deformation by folding is responsible for these fracture textures and these were filled by remobilization of the more plastic minerals of which chalcopyrite appears to have been the most so. Rui is convinced that these deformation-induced mineral movements make it impossible to determine with real accuracy the paragenesis of the ore minerals.

The ore minerals commonly are banded with darker bands being richer in sphalerite.

The Main ore body is quite regular in composition and has a much higher sulfide to gangue ratio than does the North ore. The wall rock alteration around the Main ore, from the ore body outward is: (1) quartz-muscovite schist; (2) chlorite schist, and (3) ordinary hornblende schist. The North ore body differs in that the zone farthest from the ore is ordinary phyllite and not ordinary hornblende schist. In the Main ore zone, alteration zones (1) and (2) may be, each, as much as 10 to 15 m wide and are found beyond the ore bodies in both length directions. Around the North ore body, the alteration zones are thinner, and the chlorite schist is much less well developed.

Normally, perpendicular to the schistosity, the borders between the ore and the quartz-muscovite schist are quite sharp but those between the two schists are much more gradual. In the northern portion of the Main ore body, however, the quartz-muscovite schist dies out and the chlorite schist becomes the wall rock of the ore.

Parallel to the strike of the schistosity, the massive ores normally will finger out into the schist; further on, the schists finger into each other and change gradually into ordinary, unmineralized country rocks. The Main ore body also wedges out downward. Massive ores have never been seen in direct contact with either ordinary schist or phyllite.

The RØdhammeren mine located in the HersjØ formation of probably early Ordovician age. RØdhammeren mine is on the eastern slope of the mountain of the same name toward the valley of the Skjelåfjell river. The deposit was discovered in 1774 and was mined, with a good many interruptions until the beginning of the 20th century. Although much exploration was conducted in the district, beginning in 1917, no ore worth further mining was discovered. The mines are now flooded and are essentially inaccessible.

Here another author accepts the concept that the rocks of the general district are overturned, and, although Nilsen designates the metavolcanics as the StØren group, this name covers the same beds as the term HersjØ. The rocks below the HersjØ are black shales (uppermost), polygenetic quartz conglomerate, crystalline limestone, and pelitic schists. Both silicic and mafic rocks have intruded the sediments and volcanics at a later time.

The nearest rocks of the Gula group to the RØdhammeren mine are a band of hornfelses that parallel the northward-narrowing wedge of HersjØ beds that contain the mine. They have no connection with the location of the ores or with their genesis.

The Kjoli mine area is divided geologically into two parts by the Tydal thrust fault. Most of the ore bodies of the KjØli area are contained in the RØsjØ formation where it is exposed just beyond (east of) the eastern margin of the Tydal thrust plate. Of the six formations identified by Rui and Bakke (1975), all are present in the RØros area proper (south of the southern end of the Tydal thrust). Two of the formations in the RØros area are joined together as the Kjurudal and FjellsjØ formation. The SaetersjØ formation that is identified at RØros as being between these two has been eliminated in the KjØli area.

The displacement on the Tydal fault appears (Rui and Bakke, 1975) to be no more than two kilometers.

The KjØli area is divided into two structural units by the Tydal thrust; these show marked differences in their tectonic features. Above (west of) the Tydal thrust, the Gula and HersjØ formations are in tight to isoclinal folds, and the schistosity and bedding have quite steep westward dips. The axes of the regional folds are parallel to the axes of the minor folds. Where the orientation of the minor fold axes depart from those of the major ones, this probably is due to a super-imposed gentle flexing of the beds.

The thrust zone is well exposed in many places and forms a topographically steep wall; it is composed of dense, well-jointed mylonites and ultramylonites. East of the thrust wall, the beds are flatter lying, undulate gently, and dip in many places between 10° and 20° to the west. Minor folds are more open than west of the thrust wall, and their axial planes trend, in the southern portion, NW-SE to NNW-SSE but turn to north-south farther north than Kjøli. The thrust zone is easily recognized by its discordant relations to the phyllites of the Kjurudal and Fjellsjø formations in the south and to the Røros formation farther north. The contact between the Røros and Hummelfjell formations shows strong deformation and disharmonic folding.

The mines in the Røsjø formation are located from the Allergot subarea in the north, through the Grønskar subarea (the center of which is about 4 km southeast of the center of the Allergot), through the Kjøli subarea proper (about 6 km SSW from the center of the Grønskar subarea) as far as the Svensk Menna mine (about 7 km southwest of the Kjøli mine). Thus, the total length of mineralization is nearly 20 km on a line curving SSE (in the north) to southwest from the Grønskar subarea to the Svenska Menna.

The Kjøli area contains two types of ore bodies, both of which are of the massive strata-bound type. These deposits are: (1) those dominated by pyrite and (2) those in which pyrrhotite is more important. The silicate fraction of the gangue increases as pyrrhotite increases in proportion to pyrite; in the pyrite variety, the silicates are not more than 10 per cent of the total mineral content of the ore. On the contrary, in the pyrrhotite-dominant ores, green aggregates of silicates are between 20 and 50 per cent of the minerals present; these silicates, Rui and Bakke (1975) report, give the ore the appearance of a breccia.

In the entire Kjøli district, the ore bodies in the Kjøli subarea contain that mineralization that is most rich in pyrite. The ores are composed of a dense aggregate of fine-grained, sub- to euhedral pyrite in which chalcopyrite, pyrrhotite, and minor sphalerite are interstitial. The pyrite normally shows evidence of clastic deformation. The silicate gangue minerals in the typical pyrite-type ore from the Kjøli ore bodies are quartz and chlorite that average about 5 per cent by volume. Muscovite is rare. Rather unusually for the Røros district, local concentrations of magnetite, containing minor amounts of sulfides, occur. Bands of this type of magnetite have been found underground, usually next to the hanging-wall but in places, these bands are in massive sulfide ore.

In the massive pyrrhotite-rich pyrite ores from such mines as Svensk Menna, Røros Menna, Godthåb, and Mitgruben, pyrite grains are embedded in, and locally isolated from each other by, abundant pyrrhotite; this latter mineral, plus chalcopyrite and sphalerite provide the matrix for the pyrite. Very rarely, grains of arsenopyrite and galena have been seen. Such pyrrhotite-rich pyrite ores in places contain almost pure spherical aggregates of quartz; these have sharp boundaries with the sulfides. Fragments of wall rock are, in places, included in the sulfide masses, and they are more intensely altered than the wall rock in place. These silicates are cut by irregular branching veins of sulfides.

The pyrrhotitic ores are found in the Guldal mine and some prospects near to it and in all deposits in the Grønskar and Allergot subareas. These ores contain major inclusions of green silicate fragments and quartz eyes such as are present in the pyrrhotite-rich pyrite ores. These inclusions are cemented in a groundmass of pyrrhotite and chalcopyrite aggregates and minor sphalerite. Some euhedral grains of magnetite and pyrite also may be present. Chalcopyrite normally is clustered around silicate fragments. The copper content of these ores is so varied that averages of them are meaningless.

The green silicate fragments in this ore type are composed of mixtures of chlorite, biotite, or blue-green hornblende; these fragments are bent twisted, and rotated, probably by plastic flow of the sulfides. The silicates also are cut by veinlets of sulfides. Hornblende needles in many places are bent or broken. Magnetite is quite common in this ore type. Locally tremolite may be present instead of hornblende and is similarly altered to chlorite.

These pyrrhotite ores normally contain some magnetite, rather evenly scattered through the sulfides in places but more commonly concentrated in dark and massive bands in which only minor amounts of sulfides occur. The silicate bands contain appreciable green hornblende and subordinate quartz with the hornblende subparallel to the banding. Some of the Grønskar ore bodies contain so much massive magnetite that they are reported to have been mined for iron ore.

The wall rocks in not only the Kjøli area but also in the Røros district as a whole form haloes of chlorite-rich schists around the ore masses; these zones are in places as much as several meters thick and grade out into ordinary, less chloritic schists. Zones of muscovite may lie between the chlorite schists and the ore or the ores may contain internal muscovite with or without chlorite. Subordinate minerals in the chlorite-rich schists include sodic plagioclase, epidote, carbonate, sulfides, and garnet and biotite relics. Epidote and feldspar are common minerals in the Røsjø formation, but muscovite is not.

The Kvikne mine area is about 30 km^2 and contains a large number of workings. The first mine to be worked in the area was the Prestens Gruve that is now united with the Segen Gottes mine, the former being the upper levels of the latter. A richer mine, some 100 m south, was found in 1631 and was worked as the Gabe Gottes; it was mined for 120 m along strike and 200 m down dip. After 1730 extensive prospecting was conducted and several new mines were discovered. The last new mines to be exploited were opened in 1795 and 1796, and only a little mining was done after 1812.

The Kvikne mines are located in the strongly folded Gula schists; the Gula group rocks in this area are chiefly metasandstones and calc-silicate-bearing mica schists and phyllites that have been variously metamorphosed. Within the Gula metasediments are some thin layers of metavolcanics (Gula greenstone), limestones, conglomerates, and black schists. The numerous sulfide deposits are contained in the narrow horizons of Gula greenstone and extend in a NNE-SSW direction from Røstvangen through the Kvikne area proper to Guldalen. The Cambro-Silurian rocks of the area are intruded by small bodies and swarms of trondheimite. Moderate-sized bodies of amphibolites (originally gabbros) are conformably emplaced in the Gula greenstone; associated with these are smaller bodies of now altered ultramafics, now serpentinites and metaperidotites, that are conformably emplaced in the enclosing schistose rocks.

The Gula group rocks in the Kvikne area are folded into tight isoclinal folds the fold axes of which are nearly horizontal and have a NNE-SSW trend (F_1). This folding provides a uniform schistosity to the Gula rocks in the Kvikne area; the strike is NNE-SSW, and the dip is 20° to 40°E. The rocks of the mine area strike between N35°E and N45°E, and the dip is 25° to 35°E.

A younger (F_2) deformation developed a strong parallelism of aggregates of individual minerals (biotite, amphibole, and calcite) and, in the mine area, has a NW-SE trend and a southeast plunge. Small kink bands were formed locally on the planes of schistosity; these overprint the F_2 lineation, but are not well-developed in the mine area. A prominent joint system is present in the Kvikne area and is the youngest of the tectonic elements; these joints strike in a WNW-ESE direction and have nearly vertical dips. The courses of some of the river valleys were determined by major regional joints. In some of the mines, small-scale faults parallel this joint system.

Nilsen and Mukherjee (1972) divide the Kvikne ores into two types: (1) pyritic-chalcopyrite ores and (2) pyrrhotite ores. Type (1) ores dominantly are massive, although in the Dalsgruven ore, pyrrhotite is the major sulfide in the massive variety. Pyrrhotite apparently always is the dominant iron sulfide in the disseminated variety. Only minor quantities of oxides and silicates are included in the massive ores. Again, based on the grain size of the pyrite contained in these massive ores, these authors class them as fine-grained, medium-grained, and coarse-grained ores. The medium-grained ores (1 to 2 mm pyrite grain size) contain the maximum concentrations of chalcopyrite and sphalerite; the coarse- and fine-grained ores apparently never are of economic value. The pyrrhotitic ore contains almost no other iron sulfide; they may include some ore sulfides but apparently only locally and no economic amounts. Even so, both chalcopyrite and pyrrhotite are appreciably more important than pyrite in the disseminated ores.

Two ore bodies in the Kvikne area, Kaltberget and Olkar, are chemically different from the other ore bodies in that they are almost totally lacking in zinc and include appreciable nickel not present in the other Kvikne ores. The close association of these two deposits with ultramafic rocks strongly suggests that these two deposits were magmatic, and not hydrothermal, in genesis.

Texturally, the Kvikne ores show prominent exsolution, replacement, and fracture filling. Except for pyrite, all the major and several of the accessory minerals show exsolution textures. The most impressive replacement textures are produced by the replacement of pyrite by chalcopyrite. Chalcopyrite and sphalerite fill fractures in pyrite. The cataclastic textures in pyrite indicate that that mineral, at least, has been affected appreciably by the deformation that the rocks underwent.

The various rock types that Rui (1973) recognized at Røstvangen and the mode of the occurrence of sulfide ores in them is much the same as that described for the Kvikne mines (Nilsen and Mukherjee, 1972). The location of the Røstvangen ore bodies is spatially connected with the border zone of the ordinary greenstone of the Gula group; some 100 claims of low-grade pyrrhotite and pyrite, plus magnetite, mineralizations are, on the contrary, spatially related to a rock composed of garnet, quartz, and colorless cummingtonite. The most conspicuous structures in the Røstvangen area are tight to isoclinal folds that are reflected in the outcrop pattern of the greenstone. These regional folds are overturned to the west and show a moderate-angle dip to the east; the dips usually are parallel to subparallel to the dominant secondary schistosity of the rocks. Numerous joints and minor faults strike about NW-SE.

The massive Røstvangen ore bodies contain fine- to medium-grained aggregates of sub- to euhedral pyrite in a matrix of chalcopyrite, pyrrhotite, and minor sphalerite. Locally magnetite concentrations (low in sulfides) and of high-grade chalcopyrite-pyrrhotite ores are present. During the working of the mine three types of ore were obtained by hand-cobbing; these were high-grade (pyrrhotitic) about 8.5 per cent copper and 35 per cent sulfur; export ore (pyritic) about 2.4 to 3.4 per cent copper and 42 to 46 per cent sulfur; and processing ore (pyritic) about 1.7 per cent copper and 30 per cent sulfur. The average grade of bulk production was 2.65 per cent copper, 43 per cent sulfur, and less than 1 per cent zinc. Chemical analyses showed arsenic to be about 0.01 per cent, only traces of lead, and silver between 10 and 80 ppm and gold 0.3 to 2.0 ppm.

The ore bodies in the Røstvangen mine are elongated, lenticular, and small and are arranged in an overlapping en chelon pattern. The major axes of the ore bodies plunge with the preferred orientation of the minor fold axes and the linear structures of the country rocks Rui (1973). Such ore shoots as were mined out had lengths between 150 and 200 m and widths between

15 and 60 m. The thickness ranged between 0.3 to 0.5 m up to a 15 m maximum; the mining cutoff was about 2 m.

The lower western and the lower eastern ore bodies are located at the junction of the greenstone with the lower sedimentary beds and were connected by a zone of heavily impregnated pyrite (processing) ore. The upper eastern ore body was located in the same stratigraphic position as the lower eastern and is separated from that lower ore body by locally thick masses of massive magnetite ore. The lower eastern and the lower western ore bodies lay on the same fold and the upper eastern and western probably were located on a similar and stratigraphically higher fold. Greenstone largely occupies the volume between the upper and lower western ore bodies and thins out to the east. The distance between the overlapping lenses gradually decreased upward (toward the NNW), and these ore bodies eventually join in the upper levels.

The walls of the ore bodies have generally sharp contacts and are conformable to the schistosity of the enclosing country rocks, but some massive ore cuts and replaces wall rock in the border zone between greenstone and mica-schist. The ore zone appears to be located in a fold hinge along the border between greenstone and the lower sedimentary beds. With depth the ore bodies appear to wedge out as small-scale faults are approached; they might reappear on the lower sides of these faults.

The eastern parts of the ore bodies are surrounded by light mica-schists, the minerals in which are mainly quartz, biotite, and muscovite. An easily identified zone of bleached schist, about 10 to 20 cm thick is at the immediate border between mica-schist and massive pyrite ore; it is composed of quartz, muscovite, and much sulfide, largely pyrite. Wall-rock alteration at the Røstvangen mine is appreciably less widespread than in the other ore bodies so far discribed.

Two main theories for the formation of the Røros ores have been put forward: (1) that the ores were emplaced by upward-moving ore fluids, probably produced in the final stages of the crystallization of palingenic magmas - the emplacement probably occurred during, or possibly, before the major deformation that affected the region or (2) that the ore minerals were deposited by volcanic-exhalations that reached the area during the sedimentary and volcanic processes by which the stratigraphic columns were developed (Vokes, 1973, in Rui, 1973, 1974). The strata-bound and largely stratiform character of the ores and their stratigraphic associations with volcanic rocks suggest (in the magma chamber from which the volcanic rocks came) a source of ore-forming fluids that reached the sea floor during the time period during which volcanic and sedimentary materials were accumulating there.

Rui (1973) lists the following points that must be satisfied by any theory that seeks to explain the genesis of the Caledonian sulfide ores:

(1) the clear structural relationship between the ore bodies and the major and minor regional tectonic structures;

(2) the mineralogical and chemical aspects of the wall-rock alteration that indicate the presence, at some time in the past, of chemically reactive solutions in the wall rocks adjacent to the ores;

(3) the apparent late or postmetamorphic age of the wall-rock envelopes;

(4) the preferred location of Caledonian sulfide ore in general, though not exclusively, in submarine metavolcanic sequences.

Rui holds that, during the late stage of regional Caledonian metamorphism and simultaneous folding and thrust faulting, solutions were generated in the eugeosynclinal supracrustal rocks. Rui thinks of these solutions as not being pure water but containing enough other chemicals (unspecified) that they could extract base metals, iron and sulfur from sulfide concentrations or disseminations in the rocks through which these solutions passed as they were forced upward. In such upward migrations, these solutions became concentrated between the flexed and brecciated beds along thrust planes or in fold hinges. There the sulfides, now found in the Caledonian sulfide ore bodies, were deposited. These sulfides, although Rui considers them to have been emplaced syn-orgenically, were introduced into their present surroundings sufficiently before the end of the metamorphic episode to have suffered the deformation that their examination in polished sections and in outcrops or underground so definitely shows.

Vokes (Rui, 1973, 1974) points out that not all Caledonian sulfide ores are located in fold hinges, and some of the deposition of such sulfides in fold hinges may have resulted from the flowage of sulfides originally deposited in the beds prior to folding into the hinges under the pressures of deformation. However, other sulfides, arriving on the scene in the late stages of deformation, may have been most readily deposited in the more porous hinge zones of these folds from the solutions carrying them. Vokes, however, believes that he has demonstrated that the sulfides must have been deposited in the rocks that now contain them prior to the F_1 folding. He adds that the F_2 folding cannot have produced the deformational and remobilization phenomena because they are large open folds quite different in character from those in which the ore sulfides are contained.

One of the major arguments that Rui advances for the epigenetic introduction of the sulfides in the Røros area ores is the surrounding completely of most of the massive sulfide deposits by haloes of wall-rock alteration. It is difficult to conceive of such zones having been produced syngenetically, but it is possible that they were produced by fluids involved in the mobilization process that moved syngenetically deposited sulfides from their stratiform positions in sedimentary or volcanic beds into hinge zones developed in the F_1 deformation. How it is possible to distinguish between sulfides emplaced in structures created by deformation through plastic flow and sulfides deposited in these structures while the F_1 deformation was still in process and undergoing the plastic flow caused by these forces I do not know. The choice between the two methods depends then, as Vokes says (Rui, 1973, 1974) on which is more "tidy". Vokes believes that the deformation of syngenetically deposited sulfides and their plastic flow (in some, but not all, instances) into zones of hinge folding is more reasonable; I agree with Rui that the introduction of the sulfides into the formations after they had been lithified, but considerably before the deformative process had ceased, would provide the same relationships to the host rocks as Vokes concept would do. At the same time, Rui's method explains in the all-enclosing envelopes of wall-rock alteration in a more convincing manner than Vokes' hypothesis would do.

The presence of some ore bodies in the Caledonian rocks of central Norway in thin, parallel essentially undisturbed horizons,indicates either that these sulfides were emplaced after the deformative forces had ceased or that some areas were less strongly affected by the Caledonian orogeny than others in this section of Norway. Rui (1974) argues that, in the Kjøli area, below the thrust plane, the sedimentary strata were gently folded by the F_2 folding, and the sulfides appear as thin parallel horizons that may locally swell into somewhat thicker lenses. In contrast, the same formations that contain the ores at Kjøli appear in the Kongen mine of the eastern Røros area to have much the same mineralization but, in the Kongen area, the strata have been folded into tight to isoclinal folds. In these Kongen strata, the ores have been concentrated into elongated lenses or extremely rod-shaped bodies that always are parallel to the preferred orientation of the megascopic fold axes. From this, it is apparent that the deformation of the Kjøli area was not sufficiently strong to remobilize the primary sulfides to any significant degree, whereas, in the Kongen subarea, the force was strong enough to engender appreciable remobilization. In neither case, however, is it possible to say with certainty whether the primary sulfides were emplaced syngenetically or epigenetically; all that is certain is that, no matter how the sulfides got into their host rocks, they certainly were deformed and, in many places, remobilized after their deposition by the F_1 deformative forces.

Therefore, I think it more "tidy" to conceive of the ore fluids to have been generated palingenically. During the crystallization of the molten silicate-minor sulfide-minor-water-containing palingenetic magma, the ore fluid was concentrated in the end stages of the crystallization of that magma. When sufficient pressure was exerted on the fluids by the F_1 deformation, this fluid was driven out from many points in the magma chamber or, better, from several magma chambers developed at various places in the subsurface. These fluids moved upward through paths of least resistance, reacting continuously (but at different rates) with the various rocks through which they passed until they reached one (or more) or the five stratigraphic horizons (at a minimum) in which the sulfides of the Røros-Kvikne-Killingdal area are found and were deposited. This accounts for the evidences of high temperature of ore sulfide deposition as is shown by the exsolution textures that so many of the ore sulfides exhibit and for the completely enclosing haloes of wall-rock alteration with which these deposits are surrounded.

The exsolution textures in chalcopyrite, pyrrhotite, and sphalerite indicate that these sulfides were deposited in the hypothermal range, but this deposition probably continued down into the mesothermal range as well. The deposits of this area, therefore, are so classified here. The effect of the F_1 metamorphism of the deposits in this area is, however, so drastic that the category Metamorphic-C must be added.

The age of the introduction of these ores, under the hypothesis here fovored, probably was 400 to 450 m.y. ago and places the deposition, and the metamorphism, as middle Paleozoic, the time of the Caledonian orogeny.

KONGSBERG, BUSKERUD

Late Paleozoic	Silver	Leptothermal

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Notes

The premier silver mining area in Norway was located around the town of Kongsberg (59°42'N, 9°39'E), a small settlement about 80 kilometers southwest of Oslo in the county (Fylke) of Buskerud. The largest mine, which was worked down to a depth of 1076 meters, was Kongens Grube (King's mine) but it was only one of 130 mines that, between 1623 and 1957, produced about 1300 tons of silver in total. Because the Kongsberg area was a center of silver-smithing in Norway for long before 1623, it is probable that silver had been mined on a small scale well prior to 1623 and that the presence of that metal was kept secret from the Crown for as long as possible. The concealment of the presence of silver was more easily done because the Kongsberg silver was so pure that it needed no refining; all it required was melting under charcoal in a smith's forge to be ready for the silver-smith (Neumann, 1944). To obtain the 1300 tons of silver, probably at least 2000 tons of the metal were mined, the missing 700 tons being accounted for by losses in smelting and outright theft. In the Kongens Grube, mining, as it appears to have been in essentially all mines in the district on a lesser scale, encountered rich ore on several levels with each good lode being separated from the next beneath it by as much as 100 meters of barren ground. No less than six or seven times, the ore appeared to be exhausted on a given level, but further work on lower levels always found rich, new ore until the 1067 level was reached. There, the low price of silver during the 1930s and the influx of large amounts of water forced the closing of the mine, apparently for the last time. I should think that the spectacular rise in silver prices would have caused the Norwegians to have considered an attempt to reopen the mines, but, if this has been thought of, word of it has not reached me.

The mines are located in the Precambrian Kongsberg-Bamble (K-B) formation (C. Bugge, 1917; J. A. W. Bugge, 1943), which is separated from the Telemark formation to the northwest by a prominent belt of friction breccia (or mylonite) that extends south-southeast from Modum (where several small bodies of magnesite in serpentinite were mined between 1900 and 1965) on the north and Kristiansand on the south. The rocks are all Precambrian, and J. A. W. Bugge (1978) considers that the Kongsberg-Bamble complex moved northwards as a large block in late Precambrian time. In order of increasing age, the members of the K-B formation are: (1) banded gneisses and dioritic gneisses (gray Kongsberg gneiss); (2) Vinor amphibolites; and (3) granitic gneiss (Kongsberg granite). The Vinor amphibolites occur in long dikes (the rock of which originally was hyperite, and these dikes were associated with more massive olivine norite (hyperite) bodies. These rocks all appear to be of Svecokarelian age.

In the Kongsberg area proper (Gammon, 1966), the K-B formation is composed of narrow bands of light and dark rock (of metamorphic origin). Where these bands are impregnated with sulfides, they are known as fahlbands. The sulfides of the bands are not restricted to any one rock type but may be present in biotite schist, biotite-garnet schist, chlorite schist, chlorite-garnet schist, quartz-feldspar-biotite gneiss, feldspathic quartzite, and quartzite. It is Bugge's opinion that the banded and dioritic gneisses resulted from what he calls dioritization and metamorphic differentiation. The dikes of hyperite that were intruded into the banded and dioritic gneisses themselves were altered to amphibolite. What the banded and dioritic gneiss were before they underwent dioritization is uncertain; Bugge describes the parent rocks of these gneisses as "an older complex of rocks", whatever that may mean. The period of eruption during which the hyperites and other mafic varieties were intruded was followed by a period of what Bugge calls "granitization"; during this period, the Vinor hyperites were converted to amphibolite and, apparently at much the same time, the older rock masses were largely granitized to red-granite gneiss. Some of the red-granite gneiss (as is shown by the intrusive relations it bears to these older rocks) actually was mobilized to the molten state. The granitization affected the banded gneisses and the dioritic gneisses to a greater degree than it did the hyperites; at least the change in the gneisses was more drastic than that of the hyperites.

Within the K-B formation the banded nature of the gneisses was emphasized by the development of alternating light and dark bands on a major scale. Where these bands are mineralized (normally in less than economic amounts) by sulfides, mainly pyrite but also including some chalcopyrite and less pyrrhotite, these mineralized bands are known as fahlbands. Gammon indicates that a fahlband is a concordant layer or lenses of sulfide-rich gneiss, the sulfides having been impregnated at some time after the host rock was solidified, perhaps as early as in the diagenetic stage of the yet-to-be-metamorphosed primary sediment or as late as after all metamorphic action to affect the area had taken place. This matter will be discussed further. In the fahlband, the sulfides should be in such quantity that they are too abundant to be classed as accessory minerals but too sparse to form a massive ore lens. Within any given banded host rock, the distribution of sulfides is irregular, only portions of any particular band fitting into the fahlband category. On the large scale, however, the major part of these sulfides is confined to zones that parallel the north-south strike of the metamorphosed rocks of the region. Some of the sulfide impregnations do cross-cut the structure of the metamorphosed rocks. The two sulfide-bearing zones that are of major importance in localizing the deposition of silver (the silver being located in veins, to be described later, that cut across the fahlbands at about right angles) are known as the Underberget and the Overberget fahlbands. Besides these two, 13 other fahlbands of some importance in localizing silver ore are known; in addition, there is a large number of smaller bands that are of little importance in providing for the precipitation of silver. Any fahlband is made up of hundreds of small sulfide lenses that seldom are more than two meters in their greatest dimensions, and the long axis of each of these lenses is always parallel to the dominant strike of the metamorphosed host rock. Sulfides constitute about eight per

per cent of the fahlband in the central part of any given sulfide lens; before the boundary of an adjacent lens has been reached, the sulfide content has dropped off to nearly zero. Single fahlband zones may be as much as 15 kilometers long but a width of as much as 300 meters is unusual. Within the fahlband rock, the sulfides are disseminated as an integral part of the host-rock fabric. Mica trains flow around small aggregates of sulfide grains in the same manner as around garnet porphyroblasts; the sulfides do not cut across the foliation of the host rock. If the host rock is rich in quartz, small concentrations of quartz grains form in the planes of foliation, and these lens-like concentrations may contain sulfides of larger grain size than normal and, in a few places, sulfides in small veinlets cut across these lenses.

Weathering gives the fahlbands a characteristically rusty-brown appearance that is useful in deciding as to the areal extent of sulfide impregnation in the fahlbands.

Normally, pyrite is the only sulfide in the fahlbands, but chalcopyrite is common in the Underberget fahlband and locally in the Overberget. In places, pyrrhotite is developed and generally alters to marcasite. A few other sulfides and sulfosalts are known but they are present mainly in areas near silver veins.

Gammon suggests that the sulfide impregnations in the fahlbands are due to: (1) hydrothermal solutions, (2) sulfides from injected mafic dikes, and (3) sedimentary syngenesis. These concepts must be tested against the characteristics of the fahlbands. These are:

(1) sulfides are irregularly distributed, occurring in innumerable lenses in a variety of rock types;

(2) zones of sulfide impregnation are conformable to the dip and strike of the metamorphic rocks that enclose them;

(3) the mineralogy of the sulfide lenses is simple, mainly iron and some copper sulfides;

(4) the sulfides textures indicate that they were present during host-rock metamorphism;

(5) sulfides generally are in the vicinity of graphite schists;

(6) the sulfides always are concentrated in and near shear zones;

(7) fahlband zones are common in many other areas of the Precambrian of Scandinavia and in many other metamorphic terranes.

On the basis of these data, it can be emphasized that the shear zones would furnish excellent channels for the movement of hydrothermal solutions through the bands after metamorphism at which time sulfide deposition could occur. The main objection to this idea (Gammon, 1966) is that the texture of the sulfides always is conformable with the metamorphic fabric of the bands, and this suggests to him that both silicates and sulfides underwent metamorphism at the same period. He also points to the lack of gangue minerals contemporaneous with the sulfides; this may be an objection, but it certainly is not an overwhelming one. The silver-bearing minerals are associated with gangue minerals, but these veins and the minerals that fill them are far younger than the fahlband sulfides. It remains possible, however, that the sulfides were introduced before or during metamorphism. Gammon objects that this would require so much hydrothermal activity to mineralize all the parts of the huge number of fahlbands in the Kongsberg-Bamble formation as probably to be impossible.

If the sulfides in the fahlbands were brought in by hydrothermal solutions, these solutions must have come from somewhere. Bugge (1917) thought that the fahlband-sulfides came from solutions given off by the Vinor diabase, apparently at the time of its dike-like intrusion into the area. If the Vinor diabase is not accepted as the source of the sulfide-depositing hydrothermal solutions, appeal might be made to the same source that produced the granitizing fluids (whatever they may have been). If both of these are eliminated by fact or prejudice and a hydrothermal mechanism is preferred for the deposition of the fahlband silicates, appeal must be made to an unknown source; such appeal is considered the greatest argument against assuming hydrothermal deposition where no conceivable igneous rock source for the ore fluids is known.

A syngenetic genesis for the sulfides in the fahlbands would explain the conformity of the sulfides with the dips and strikes of the metasediments that contain them, the metamorphic textures of the sulfides, and the common occurrence of the sulfides in the Kongsberg-Bamble formation. This does not, however, make understandable the normal association of sulfides with shear zones even though it may do so for the close juxtaposition of sulfides and graphitic schists. Gammon, however, suggests that at least part of the K-B formation was laid down in a near-shore, deltatic environment, where, on occasion, organic matter was deposited and covered before its decomposition was complete. H_2S generated in this material may have fixed such iron and copper as were present into sulfides. Later metamorphism could have converted this organic material to graphite and produced the present sulfide textures. Shear zones could have developed preferentially in the graphite schists because of their relative incompetence under applied stress.

In addition to the Precambrian (Svecokarelian) rocks, the area was intruded by diabase dikes that certainly are post-Silurian and probably are closely related in time to similar rocks produced by Permian igneous activity in the Oslo region. What connection these dikes may have with the silver mineralization, if any, will be discussed later as will be the relationship between the fahlbands and the formation of silver-bearing, mineable veins.

Bugge (1917) and Neumann (1944) divide the Kongsberg veins into two classes: (1) first-generation veins and (2) secoond generation. The first-generation veins are designated by both authors as sulfide-bearing quartz-breccia veins. These are present not only in the area of the Kongsberg calcite-silver veins but also well to the north, south, and east of it; in this last direction they have been traced to the vicinity of Oslo. These first-generation veins in the Kongsberg area normally strike more or less north-south, with ENE being the most common direction; the dip in all places is very steep. The strike of these veins is the same as that of the younger diabase dikes, which suggests that the dikes and these veins are genetically connected. In places, quartz veins are intersected by diabase dikes; in others, fragments of what probably is diabase of the dikes is present in the quartz veins, thus tying them very closely in time. In some locations, both diabase dikes and quartz veins occur in the same fracture. The thickness of the quartz veins ranges from essentially zero to more than 10 meters with the average being about one meter. In the strike direction, the quartz veins may be as much as 6 kilometers long. These veins usually contain fragments of their wall rocks, so abundant are these in places as to justify the designation as mineralized breccias. The abundant quartz is compact, gray, and usually intersected by fissures; vugs are frequent and commonly contain small, transparent crystals of quartz. Fluorite is another gangue mineral in modest amounts; it normally is green. Barite is rare. The order of frequency of the sulfides in these breccia veins is: pyrite, pyrrhotite, sphalerite, chalcopyrite, and galena. C. Bugge (1917) says that the chalcopyrite and galena are silver-bearing but obviously in very minor amounts, 0.01 to 0.1 per cent, although locally galena may contain up to 1.0 per cent silver. The amount of gold in these sulfides is so low as to be negigible. In several places, these veins have been mined for fluorite.

In addition to the quartz-breccia veins, some calcite veins of the same strike are present that are of a different variety from the silver-bearing ones. These are wider than the calcite-Ni-Co-Arsenide-Ag silver veins, and they are more widely distributed. In addition to calcite, these veins contain as much as 30 to 40 per cent fluorite, plus sphalerite and galena. At least one of these veins was followed underground for several hundred meters, but no silver was found. These calcite veins probably have no genetic connection with those calcite veins that are silver-bearing.

In places, those portions of the quartz-breccia veins within the fahlbands have been traversed by the same ore fluids that produced the calcite-nickel-cobalt-arsenide-native silver veins, and locally appreciable silver was deposited in them. Bugge designated these veins as "Main Veins"; they in many places are very rich in silver, though they contain argentite in much larger proportions to native silver than do the calcite-Co-Ni-Arsenide-Ag veins. They have been mined to a considerable extent, but they do not appear to have provided more than a minor fraction of the total silver production.

The calcite-Ni-Co-Arsenide-Ag veins, in contrast to the quartz-breccia veins, strike about east-west and are essentially confined to the fahlband zones, extending into the surrounding gneisses only a few meters at most. Further, the individual veins are quite narrow, seldom being more than from a few millimeters to a maximum of 0.5 meters; their average thickness is from 5 to 10 centimeters. These veins in many areas are present as vein-groups in which the parallel veins are cut by fissures that cross them at various angles. The result is a series of mineralized zones or even a single zone, the width of which may be as much as several meters. Although the mineralized portion of these calcite-Ni-Co-Arsenide-Ag veins is confined essentially to the width of any fahlband in which they are mineralized, they never much lengthen in their east-west strike direction; not uncommonly it appears that the calcite vein systems are wider than they are long. Before the deposition of the minerals that filled the calcite-Ni-Co-Arsenide-Ag veins, they constituted a network of open fissures (Fiederspalten). Single veins, however, did exist and in considerable numbers. Some fissures had strikes parallel to those of the fahlbands (i.e. N-S) and were mineralized with calcite and some zeolites and adularia; these "striking veins" seldom carried silver. The silver content of a calcite vein usually is greatest where it is close to a striking vein. Why this should be so is a matter on which there is considerable uncertainty. Neumann thinks that this is due to tectonic factors entirely. This is true because the contact between a striking vein and the Fiederspalten that contain calcite-Ni-Co-Arsenide-Ag veins is wider and more open than anywhere else along the short course of the calcite vein, thereby permitting the ore fluids to flow most freely along the upward-trending channelway so produced.

The silver-bearing veins of the Kongsberg district contain calcite as the principal gangue mineral and native silver as the main ore mineral (C. Bugge, 1917). These veins strike about east-west and cut across the north-south-striking metamorphic rocks of the district (including, of course, the fahlbands) at about right angles, where for all practical purposes, the mineralization in the veins takes place. The silver veins mostly dip between 70° and 80°S, but some dip steeply north and 45°S-dipping veins are common in the northern part of the Overberget fahlband.

The most important mineral in these veins is native silver that occurs mainly as threads, moss-like masses, or plates; crystals of silver are rare. In a few places, large masses of silver (up to 500 kilograms) have been found. In part, the silver is amalgamated with mercury, Hg generally being less than 0.5 per cent, though locally as much as two per cent, of the silver. The silver is essentially lacking in gold, the average amount of gold being 0.002 to 0.005 per cent. Silver seems to contain more gold than the average where the veins contain quartz gangue. Argentite is present to the extent that it provides about three to four per cent of all the silver mined. Pyrargyrite, stephanite, and other sulfosalts are present in only minor amounts and account for no more than a tiny fraction of the silver.

In addition to the calcite, which is the dominant mineral in the silver-bearing veins, quartz, barium-bearing feldspar (in some abundance), barite, axinite, albite, chlorite, zeolite, and bitumen also are present in those veins.

Bugge also mentions "rotten veins"; these veins are mylonitized east-west striking veins, that is veins of the same direction as the ore-bearing calcite veins but ones in the filling is a rotten looking mass of clay, calcite, and, in many places, quartz and the remains of rock fragments. They contain no silver interspersed in their much disturbed filling-material. The effects of tectonic activity have been, in Neumann's opinion, most severe where fissures that would have been potential sites of calcite-Ni-Co-Arsenide-Ag deposition were so badly crushed and broken that they were largely incapable of permitting the upward passage of such ore fluids as may have passed in their vicinity. In either direction from these rotten veins, however, this "rotten" vein-filling quickly dropped off and left these veins open for filling by calcite-vein-material. The strike of the "rotten veins" is that of the calcite-Ni-Co-Arsenide-Ag veins (i.e. E-W), and they dip from 45° to 60°S, perhaps somewhat flatter than the average calcite-Ni-Co-Arsenide-Ag vein.

Although a large number of minerals, at least 40, has been identified by Neumann, most of these can reasonably be categorized as rare. Most of the attention of mineralogists has, understandably, been given to native silver as by far the most abundant ore mineral and definitely the most valuable. Neumann reports that silver is present in three major forms: (1) "enclosed" silver, that is, silver the outer form of which has been determined by the surrounding minerals; (2) as "thread" silver in which the silver threads are long, curled, and twisted with characteristic streaks parallel to the long dimension. The threads may be long and thin or as curled rods or as short thick rods or as balls of fine fibers or aggregates of smaller threads that resemble moss; other forms are reasonably designated as hair or tooth silver. Some masses look like twisted roots or deformed trees, (3) as crystals with well-defined faces; these are very rare with perfect and complete faces but some showing of crystals faces is not uncommon.

It appears, from experiments first made by Hauy in 1801 that heating argentite will cause threads of silver to grow outward from the argentite surface. Neumann suggests that the thread silver was considerably, though not entirely, produced from argentite. Experiments also have shown that thread silver can be developed by exposing stibnite or cobaltite to silver sulfide solution; various combinations of sulfate solutions also used experimentally have produced silver threads from argentite. How far these experiments are applicable to Kongsberg is, of course, uncertain. Neumann reports seeing large masses of thread silver in the mines without finding even the smallest amount of argentite associated with the silver.

Although the amount of silver showing crystal faces appears to be small, a wide variety of crystal faces have been reported, with octahedrons, cubes, and tetrahexahedrons being associated together.

Of the minor impurities in native silver, so minor that little refining of the silver is needed, mercury is the most important with one specimen containing over 23 per cent Hg, but with most mercury-bearing specimens having about five per cent of that element. It is interesting that the thread silver contains much less mercury than the other silver forms. Perhaps this is due to the late stage deposition of thread silver in relation to silver in other forms, a time at which the mercury content of the ore-forming solutions may have been much less than it was in the earlier stages. In a few instances, golden silver was found in Kongsberg in which the gold percentage ranged from more than 27 per cent to as much as 53.1 per cent.

Certainly, the rarity of this silver type was a matter of keen disappointment to a long line of mine managers and to the Norwegian government. Traces of nickel, cobalt, copper, bismuth, iron, and arsenic also have been determined to be present in Kongsberg silver. Some question arises as to whether the silver sulfide in the ores was precipitated above or below the inversion temperature of argentite to acanthite; Neumann thinks that most of it was deposited above that temperature (173°C), but Hurlbut and Klein, in the 19th edition of Dana's Manual of Mineralogy, seem to want to drop the name "argentite" entirely from the mineralogical literature. They admit that an inversion takes place at 173°C, but they claim that what takes place is simply a change in crystal form of acanthite and not a change from argentite to acanthite. This does not matter much as long as the Ag_2S can be shown to have inverted from one crystal form to another, no matter what the higher-temperature and lower-temperature forms are called.

Pyrite is the most common sulfide in the Kongsberg ores, and Neumann believes that it was deposited throughout the mineralization cycle of the main veins. Pyrite is, in places, replaced by calcite, native silver, and argentite, but normally it shows little corrosion. Next to pyrite, sphalerite is the most common sulfide; nevertheless, it is an unusual mineral. Except for iron and cadmium, its impurities are minor in amount. It does, however, appear to have been introduced in two generations, one older than the native silver and the other younger. The older sphalerite contains exsolution blebs of chalcopyrite; the younger does not. Iron is more abundant in the older sphalerite than in the younger. Chalcopyrite also occurs in two generations, separated by the formation of native silver the younger is connected paragenetically with pyrrhotite and nickel and cobalt arsenides, the older with sphalerite. Some older chalcopyrite contains exsolution blebs of sphalerite, but other chalcopyrite, equally old, does not. Where paragenetic relationships can be determined, most of the chalcopyrite is young. Silver sulfosalts are paragenetically close to chalcopyrite but even in total these are rare in the Kongsberg mines. Galena is very rare and most of it occurs in relationships similar to those of the older chalcopyrite; a little younger galena is known. Pyrrhotite ranks after pyrite and sphalerite in abundance but is more abundant than chalcopyrite and galena. Niccolite (NiAs) is, according to Neumann, not uncommon in the ores but is not abundant; it usually is in small crusts on native silver; it also is found in veinlets that cut native silver, so appears definitely to be the younger of the two. The diarsenides (?), rammelsbergite-safflorite [$NiAs_2$ - $CoAs_2$] and chloanthite-smaltite [$Ni,CoAs_{3-x}$ - $Co,NiAs_{3-x}$] are very widely distributed through the ore but are far from abundant. They occur as thin coatings around native silver. Neumann thinks that the arsenic is low enough in these minerals that the formula should contain As_{2-} and not As_{3-} as Hurlbut and Klein suggest. The possibility exists that what Neumann was dealing with was some mineral other than those just mentioned. The sulfosalts, such as polybasite, pyrargyrite, and stephanite, are extremely rare, as are minerals of the sternbergite group.

Neumann (1944, p. 117, Figure 39) gives a paragenetic diagram for the Kongsberg mineralization in which he considers that deposition started between 400° and 500°C with that of quartz, calcite, and pyrite, (in that order); fluorite and axinite also crystallized in that temperature range - two lower-temperature depositions of fluorite also occurred, perhaps at about 300° and around 200°, respectively. Barite began to precipitate just below 400°, followed by galena, sphalerite and chalcopyrite, the sulfosalts of silver, argentite, marcasite, and, after an appreciable gap, native silver, this last mineral depositing at about, perhaps, 250°C. Native silver was followed by pyrrhotite, niccolite, the diarsenides, and finally by the zeolites. Argentite had two lower-temperature generations, but these did not form Ag_2S in anything but minor amounts.

The close association between fahlbands and ore deposition is so common as to indicate strongly that the bands exerted some control on the ore fluids to cause them to precipitate their mineral lodes dominantly in that type of environment. For most of the life of the Kongsberg mines, the so-called fahlband rule has provided a guiding principle for most exploration. This rule is that: "the veins are silver-bearing only where they cross an ore band." This rule does not say that all veins are silver-bearing where they cross an ore band, but it does indicate that they have no chance of being so where they do not cross an ore band. It seems certain, however, that there is no direct connection between the amount of sulfide in an ore band and the amount of ore in the vein that cuts it. A suggested change in the fahlband rule was to say that the greater the ability of a particular fahlband to conduct electricity, the greater will be the content of sulfide in a main vein that crosses it. This rule places particular emphasis on the importance of pyrite-impregnated quartz lenses and veins which the fahlband contains. Nevertheless, the validity of this rule has been questioned by ore geologists at Kongsberg, and one of them said that the rule actually said that a vein is silver-bearing only where it crosses the fahlband at a point where it (the vein) is silver-bearing. Such an interpretation of the rule, of course, would be nonsense. But, almost certainly, a vein must cross a fahlband to be ore-bearing, and it will most likely be ore-bearing where it (the vein) does cross the fahlband. Even if the rule is true, this does not explain why the silver (and other metallic minerals) should have been precipitated under this apparent influence of the sulfides in the fahlband.

J.H.L. Vogt (1899) thought that the pyrite in the fahlband in question had been attacked by the ore solutions, thereby liberating H_2S that caused silver to precipitate as argentite. This theory requires that the amount of sulfide in the fahlband be proportional to the amount of argentite in the main vein; this, of course, is not true. Further, almost all the silver was precipitated as native silver and not as argentite.

Chr. A. Munster (in an 1894 paper to which I do not have access) is reported by Neumann to have suggested that the silver was electrochemically precipitated by stray electrical currents in the earth. It follows from this that the sulfides in the sulfide-impregnated fahlbands acted as electrodes; thus the vein must have furnished the cathode and the fahlband the anode, a most unlikely situation. Neumann thinks that the main cause of precipitation of the main vein minerals was decreasing temperature in the ore fluid and that the decrease in temperature was most marked where the ore fluid in the vein was in contact with the fahlband because the fahlband conducted heat away from the ore fluid more rapidly than did the normal wall rocks of gneisses. It hardly seems that any of these suggestions solves the problem, although the last one might help if it also is considered that an increased amount of hydrogen ion might mingle with the ore fluid as it passed by, thus aiding precipitation. Nevertheless, the problem as to why the silver (and the other vein minerals) were deposited in the vicinity of the vein-fahlband crossings is not solved, and it is unlikely to be except serendipitously by someone working on another deposit having similar characteristics, if one exists.

The formation of the Kongsberg deposits almost certainly was carried out by hydrothermal solutions, but the source of the solutions is far less certain. Bugge considers that, without question, the ore fluids came from the young diabase (Permian). The sulfide-bearing, quartz-breccia veins are very closely related in age to the diabase, but the diabase dikes definitely were solidified before the silver-bearing (calcite) main veins were introduced. Neumann believes that the calcite (main) veins are too young to have been supplied with their mineral content from hydrothermal fluids given off by the magma chamber in which was produced the magma that became the diabase dikes. He thinks that the post-diabase granitic intrusives of the Drammen area, nearer Kongsberg than Oslo, supplied the ore fluids that deposited the Kongsberg ores. The problem at Kongsberg is much the same as that at Cobalt and Gowganda in Ontario where the rich silver ores of those camps are located in an area with a major quota of diabase dikes; yet the most recent opinion on the Cobalt-Gowganda area is that granites later than the diabases came from the same magma-chamber source of the Cobalt ore fluids. The principal basis for this opinion is that silver is found in fractures in the diabase, fractures that could have been produced only after the diabase had solidified. Thus, the situation in both districts is such that the ore is definitely later in both than is the diabase and in both granitic magma chambers are available to which appeal for hydrothermal fluids can be made.

The only important ore mineral in the Kongsberg deposits is native silver, and it is Neumann's opinion that the Kongsberg silver was deposited at temperatures of slightly above 300°C. The lack of appreciable amounts of base-metal sulfides makes it difficult to place the deposits in the modified Lindgren classification, but native silver, associated with Ag_2S (to avoid the argentite-acanthite controversy) that must have been deposited above 173°C, makes it no great risk to categorize the deposits as leptothermal.

SØVE, TELEMARK

Early Paleozoic	Columbium, Iron as Hematite	Cb-Magmatic-Fla Fe-Hypothermal-2 to Mesothermal

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Notes

The center of the Fen area of southeastern Norway is located about 110 kilometers southwest of Oslo and about 3.0 kilometers southeast of the industrial center of Ulefoss (59°17'N, 9°15'E) on the southwest shore of the Norsjø (lake). The area is a small complex of carbonatites and peralkaline rocks. The carbonatite center of the area is rudely circular, with a diameter of some two kilometers and composed of a variety of carbonate rocks that will be described below. The entire Fen vent is about five kilometers^2 in area and contains, in addition to the carbonate and silicate-carbonate rocks, four additional groups of rocks: (1) metasomatic rocks (fenite and pulaskite); (2) silicate rocks (urtite, ijolite, melteigite, malignite, jacupirangite, vibetoite, and tinguaite); (3) transitional rocks (juvite, kamperite, tveitåsite, hollaite, kåsenite, and ringite); (4) kimberlites or peridotites (damkjernite = kimberlite and sannaite). Most of the transitional rocks contain (the only exception is kamperite) more or less calcium carbonate (from 66 to 3 per cent) and are mixed silicate-carbonate rocks (Barth and Ramberg, 1966).

The economic importance of the area does not lie in the unusual silicate rocks, but in the hematite in the rødberg (a red-colored carbonate rock in which hemataite is finely divided), but in which the mineable ore is in northwest-southeast-striking veins) and in søvite (normally a marble-appearing rock in which, in addition to calcium carbonate, ankerite, and dolomite occur in varied amounts). The ore-mineral of columbium is reported to be koppite (a name not recognized by Fleischer) which Bergstøl (1960) says is a Ca-Na niobate, poor in Ti and Ta - some or all koppite has been partly converted to columbite but retains its original octrahedral form - the original formula apparently was $[(Ca,Na)Nb_2O_6]$ with some of it now being $FeNb_2O_6$.

The iron mines in the Fen rødberg operated from 1652 to 1952 to a depth of nearly 200 meters; the total production was some one million tons in which the average content of iron was 50%; P ran 0.45%; the Mn 1-2% and S 0.3%. This ore is present in veins and lenses in the rødberg; the rødberg proper did not contain enough hematite to be economically mineable.

The columbium was recovered by mines in søvite that operated from 1953 to 1965; the annual production increased from 1230 kilograms Nb_2O_5 in 1953 to 2730 kilograms in 1962. Mining began in søvite in the Cappelen mine near the shore of Norsjø (lake); the ore was cut off on the southwest against a fault. This ore was later discovered to have been downdropped to about 60 meters below the surface, and a shaft was sunk to develop and mine it. Several other, but smaller, faults were found in the course of this operation.

The major minerals in this mine were pyrochlore (koppite) 0.5 per cent; apatite 7.0 per cent; magnetite 3.0 per cent; and calcite 80 per cent. Minor silicates present in the søvite included alkali amphibole, tremolite, mica, and zoisite; accessory minerals were fluorite, topaz, zircon, and barite. Another mass of søvite (the Hydro deposit, also near the north edge of the søvite body) was too low-grade in columbium for commercial exploitation. A drift was run from the Hydro area some 900 meters south into the central portion of the søvite body where additional ore was found. This mine (known as the Tufte) contained a larger proportion of dolomite than in the Cappelen area, and in one portion of the Tufte mine a zone containing uranium-rich pyrochlore = uranpyrochlore $[(U,Ca,Ce)_2(Nb,Ta)_2O_6(OH,F)]$ was discovered.

During the last years in which the mine was operated, annual production was 30,000 tons from the Tufte mine and 100,000 tons from the Cappelen.

The columbium content of the carbonatite decreases in the order from søvite to rauhaugite (90 per cent ankerite; eight per cent apatite; two per cent feldspar; with minor barite and magnetite) to rødberg. On the contrary, the content of yttrium, the rare earths, and thorium increases in the same direction. The rødberg contains 2.5 to 4.0 per cent of rare-earth minerals of which the most important are: monazite ($CePO_4$), synchysite $[Ca(Ce,La)F(CO_3)_2]$ and parisite $[(Ce,La)_2CaF_2(CO_3)_2]$. The rødberg area contains large volumes of that rock in which there are 200 ppm Y, 2000 ppm La, 300 ppm Sm, and 1400 ppm Th. The hematite extracted from veins and lenses in the rødberg contains 0.2 per cent of Th. Because of the fine grain of the rare-earth minerals (on to 50 microns in diameter) laboratory research has not developed a successful method of concentrating the rare-earths and yttrium-thorium minerals, but further work may yet solve this problem. Work by Amli (1975) suggests that the Fen area carbonatites may be a possible economic source of scandium. He believes that thorveitite $[(Sc,Y)_2Si_2O_7]$ is present as rather widely disseminated crystals in rauhaugite and rødberg. His analyses show that the maximum amount of scandium in the samples he took is 83 ppm in rauhaugite and 131 ppm in rødberg. Scandium-bearing columbite (Amli's term) also is present in both of these rock types. His analyses for scandium in columbite in rauhaugite range between 0.64% and 1.32%; considering the small content of columbite (koppite, according to Bergstøl) in this rock (perhaps 0.5%), the amount of scandium in the rock certainly is not economic under currently imaginable conditions.

The localization of the Fen area rocks is governed by faults that strike parallel to the margins of the Precambrian Baltic shield; the carbonates of both Alnö in Sweden and in the Kola Peninsula (USSR) are associated with similar faults. The faults that govern the location of the Fen are Eocambrian or Early Paleozoic (565 m.y.) as opposed to the 240 m.y. faults that are the latest tectonic events in the Oslo area. Barth and Ramberg, however, believe that the Fen and Oslo fault systems mark the beginning and the end, respectively, of a long epoch of faulting. The Oslo rift valley is a northward continuation of the Rhine graben and is similar to the African rift valleys in that it originated in the Precambrian and had a long continued history of earth movements and, in the Oslo and African areas, of the production of magmatic fluids that were high in carbon dioxide and other volatile materials. The culmination of this tectonic and magmatic activity was the introduction of the Oslo (Permian) magmas. As is also characteristic of the Alnö area, the Fen area has numerous breccia zones and concentric radial faults, on some of which major vertical displacement occurred.

The Fen area faults radiate outward from the center of the ring complex and extend into the surrounding gneiss for several hundred meters. Where the faults pass through carbonatites, the carbonate material reacted in a plastic manner and was strongly foliated parallel to the strikes of the faults. The concentric faults have their centers in the Fen area mines and, in many places, mark the contacts of the rocks of the Fen complex, but some of these faults are apparently confined to the carbonatites. The structure thus produced is a typical cauldron subsidence with the central parts having sunk relative to those rocks farther out.

Numerous occurrences of breccia bodies are found in the Fen area, and those of granitic composition are present not only in, but well beyond the limit of, the peralkaline suite of the Fen explosion vent. Those breccias outside the Fen area proper are more sparsely distributed than those within it. The granite breccias are made up of fragments up to 0.5 meters across in a matrix of quartz and alkali feldspar that is aplitic in texture.

In extreme contrast to the granitic breccias, the matrix of the other breccias is damkjernite (better spelling damtjernite). This is a dark, porphyritic rock with phenocrysts of biotite (1-5 centimeters long) plus lesser ones of hornblende, pyroxene, and olivine; the phenocrysts normally are broken, bent, and rounded. The hornblende usually is barkevikite that is covered by a green amphibole that also appears in the groundmass. The damkjernite is composed of phenocrysts of biotite and titan-augite in a fine-grained groundmass of pyroxene, hornblende, biotite, perovskite, and magnetite with interstitial nepheline, microcline, and calcite (some primary and some secondary). The damkjernite is classed as a kimberlite. Inclusions in the damkjernite are: (1) exogenic; gneiss, fenite, the urtite to jacupirangite series, and søvite and/or rauhaugite and (2) endogenic; peridotite, pyroxenite, and hornblende. In certain damkjernite breccias, the fragments are so abundant that the damkjernite is a quite minor constituent of the material. Endogenic fragments, in many places, are surrounded by aureoles of damkjernite that give a nodular structure to the rock. Inclusions of both types of fragments normally are angular but locally are in well-rounded shapes. Dawson concluded that kimberlite is the result of interaction between inclusions of forsterite, enstatite, diopside, and ilmenite, on the one hand, and a water-rich fluid containing carbon dioxide, alkalies, lime, and, in some cases, silica on the other.

The damkjernite has essentially no effect on the rocks through which it passes nor are they visibly brecciated or shattered. Some of the small damkjernite pipes have thin rays that cut the surrounding gneiss for up to 10 to 12 meters. Barth and Remberg conclude that the emplacement of the damkjernite was not explosive but was achieved by "gas coring" with the location of the pipes being governed by the existing fissure pattern.

Bergstøl (1960) is quoted by Barth and Ramberg as saying that damkjernite is the only truly igneous rock in the Fen area; this statement may be somewhat exaggerated, but contains, as will be seen later, at least a germ of truth.

Barth and Ramberg (1966) agree with Bergstøl and Svinndal (1960) that the granite and damkjernite breccias grade into each other; if this is true it argues against the two being products of separate episodes of volcanic eruption and subsequent subsidence. Both the granites and the damkjernites are encountered beyond the Fen area proper, but the granite extends much farther from the ring complex than does the damkjernite. This suggests that, at the very least, the two may not have been essentially simultaneous and almost certainly that they were not from the same magma source.

Most of the damkjernite in the Fen rocks is in irregular bodies that have vague contacts with the rocks that enclose them; in the gneisses, the damkjernite occupies dikes and pipes that are arranged along fault lines and shatter zones. The dikes normally are from 20 centimeters to five meters wide and may be as much as 200 meters long.

Dikes of tinguaite [a textural variety of phonolite = a fine-grained extrusive rock, mainly alkali feldspar (anorthoclase or sanadine), and nepheline, that is the extrusive equivalent of nepheline syenite] occur thickly distributed around the Fen. Barth says that they are ubiquitous and range in size from small spatters to rather large dikes and irregular bodies. This tinguaite is a nepheline syenite porphyry with phenocrysts of alkali feldspar, nepheline, and some biotite and pyroxene; the ground mass is made up of feldspar laths, nepheline, aegerine (acmite) needles in radial or criss-cross textures, and perhaps sodalite. The dikes appear as much as 10 kilometers from the Fen rocks and definitely are older than the rocks of the Fen complex; xenoliths of tinguaite are present in damkjernite plugs and breccias, and one damkjernite dike cuts a tinguaite body and a plug of damkjernite penetrates another tinguaite dike.

Several silicate-carbonate rocks which are highly varied both mineralogically and chemically, were designated by Brøgger (1920) as (1) hollaite [55% pyroxene, 16% calcite, 10% nepheline, 6% garnet, 4% apatite, and 9% biotite]; (2) kåsenite [25% pyroxene, 57% calcite, 7% apatite, 5% nepheline (muscovite), 6% chlorite, plus feldspar, sphene, and ore], and ringite [20% acmite, 66% calcite, 6% alkali feldspar, 5% apatite, plus biotite, sphene, and ore]. These hybrid rocks are poorly defined and pass into each other and into fenite, søvite, and damkjernite almost imperceptibly; they cannot be true igneous rocks but must have been produced by magma-generated fluids that probably derived from the kimberlite magma. The compositions of these transitional rocks differ widely from the estimates given above for the averages of the three types. The Glossary of Geology does not recommend the usage of any of these terms. Hollaite and kåsenite and ringite are better referred to as pyroxene søvite or søvitic melteigite.

Fenite is the contact metamorphic aureole around the eruptive rocks of the Fen area. The name was given to it by Brøgger in his pioneer work on carbonatites, and essentially all later-studied carbonatites are surrounded by it. On its outward margins, the fenite grades most gradually into the gneisses around it. The first effect of fenitization is the conversion of the biotite in the gneiss to aggregates of acmite or iron oxides. As the temperature of the rocks involved is raised, a homogeneous alkali feldspar is formed that inverts, on cooling, to microperthite. Quartz and plagioclase are replaced by this alkali feldspar, and the texture of the gneiss is, at the same time, converted to a "mortar structure". In the most usual fenite, microperthite and some albite are present as the sole light minerals; some types, however, contain some nepheline. The actual conversion necessitates only a small exchange of material between gneiss and the fenitizing solutions.

Most of the fenite has been produced in the southwestern portion of the Fen area. Barth and Ramberg mention two varieties: (1) hornblende fenite (soda amphibole instead of acmite) and (2) pulaskite fenite (in which nepheline is altered to muscovite). The composition of the feldspars indicates to Barth and Ramberg that the fenite was developed above 650°C. Apatite, sphene, and zircon are present as accessories.

The cause of fenitization almost certainly is solutions from the carbonatite magma; an urtite-melteigite magma (if one exists) could not have produced fenitizing solutions because, around low-carbonate alkalic rocks, wall-rock gneisses are not fenitized. Nearly all carbonatite bodies in gneiss are surrounded by zones of fenite. In the one place in the Fen area where an urtite-melteigite mass is surrounded by fenite, the urtite-melteigite must have been intruded into fenite that was produced earlier by solutions from the søvite magma. Barth and Ramberg think that the first intrusion was a rather pure carbonate magma; this later was followed by the introduction of an alkali-silicate magma. As the carbonate magma could not have developed by fractional crystallization, it is probable that it and the alkali-silicate magma initially were all present in one melt, from which the carbonatite material was separated as an immiscible phase. From studies of other carbonatite-containing areas, it appears that the carbonatite magma can be introduced before the alkali-silicate one or after it; in the Fen area, however, the søvite appears to have been driven from the magma chamber before the alkali-silicate material. It appears the fenitization was a long-continued process that was in effect after the emplacement of all the main rock types. Brauns (1926) points out that it covers the primary rocks as with a "veil". Allowance must be made for this "veil" or erroneous conclusions may be drawn about fen-rock genesis.

The urtite-melteigite series of peralkaline silicate rocks that are of subhedral texture and are feldspar-free. The size of the grains ranges from one to five millimeters in diameter, but, in some places pegmatitic varieties are known. They are confined essentially to a body of irregular shape in the southwest portion of the complex. This mass of rock is not sharply divided among the three varieties (urtite-ijolite-melteigite), but these grade so gradually into each other that definitive boundaries cannot be drawn. This mass of rocks is, in all places, surrounded by fenite with a one-meter zone of transitional rocks always obtaining between fenite and the particular peralkalic peralkalic rock adjacent to it. Immediately north of this peralkaline body are a few outcrops of nepheline syenite and malignite (nepheline 20%; acmite-diopside 49%; orthoclase 21%; apatite 6%; biotite 4% plus sphene and ore on the average). Near the south edge of the complex is a small body of vibetoite (hornblende plus titan-augite 64%; calcite 14%; biotite 10%; apatite 9%; sphene 3%; plus or ± albite and nepheline). This rock is surrounded by søvite with which it has a gradational contact; it is coarsely crystalline, with minerals five to 20 millimeters long that have curved and jagged outlines. Parallel growths of pyroxene and amphibole are common, and the rock is transitional to kimberlite. In all of these rocks, except vibetoite, the pyroxene is diopsidic with some acmite and with some zoning that results in rims richer in acmite than the cores.

In the urtite-ijolite-melteigite series, urtite is mainly nepheline with 0% to 30% mafic minerals of which the clinopyroxene acmite $[NaFe^{3}Si_{2}O_{6}]$ is most abundant. Ijolite is nepheline 40% to 70%; mafic minerals 30% to

60% (mainly a clinopyroxene) plus sphene, apatite, and melanite (Ti-andradite = $Ca_3Fe_2^3Si_3O_{12}$). Meltigeite is composed of nepheline and 60% to 90% of mafic minerals, particularly green pyroxene. Jacupirangite, the most mafic of the series, in mainly titan-augite and magnetite with small amounts of nepheline. Barth and Ramberg say that this series of peralkaline silicate rocks can be explained easily as having formed from a cooling magma that was undergoing fractional crystallization. On the other hand, they point out that von Eckermann has shown in what seems to be a conclusive manner that a similar series of rocks was developed metasomatically. As I read von Eckermann's 1948 work, I believe he says that the peralkaline suites at Alnö were formed by the fractional crystallization of anatectically formed magmas and not that they were produced by metasomatism. So it may not be necessary to be in opposition to such a renowned authority as von Eckermann, and it can be agreed that the characteristics of the urtite-melteigite series in the Fen area are such as to make it the more probable that they were formed by fractional crystallization from a single magma than metasomatically. The transitional rocks, kåsenite, hollaite, ringite, juvite, kamperite, and tveitåsite, however, are the result of mixtures between carbonatite and peralkaline silicate magmas, perhaps when both were still molten, perhaps when one had at least partially crystallized.

The series of søvite, rauhaugite, and rødberg are all carbonate rocks with the søvite and rauhaugite being essentially monomineralic, the former being principally calcite and the latter ankerite. Rødberg consists of calcite (mainly), ankerite, and much finally disseminated hematite with dikes and veins of almost pure hematite (from which at least one million tons of iron ore were mined over 300 years). Much of the volume of transitional and peralkaline rocks is cut by numerous dikes of søvite; the søvite dikes partly have sharp contacts with the silicate rocks and breccia bodies which the søvite invaded. The breccias and larger dikes are largely in transitional rocks where much fragmental and diffuse material is incorporated in them. Smaller dikes are filled with nearly pure søvite that shows sharp contacts against the country rocks; some traces of crushing and boudinage-like flow lines can be seen in such søvite.

The søvite dikes have the same circular strike as the main søvite mass and more or less concentric dips. Although these dikes are not regular cone sheets as they are at Alnö, they seem to suggest a vaguely defined focus some 2000 to 3000 meters below the present surface.

The high content of sodium carbonate in the carbonate lavas at Katwe in Uganda, if it also obtained in the søvite magma in the Fen area, could have provided the necessary sodium for fenitization and other later metasomatism.

The other minerals in the Fen søvite, beyond the 67.0% calcite, include 7.7% fluorapatite; 22.0% silicates, 2.0% magnetite and hematite, and 1% pyrite. The silicates are phlogopite, amphibole, and some acmite. Locally, biotite, usually altered to chlorite, is the main silicate mineral. Additionally, minor microcline, albite, and more rarely quartz are found. Nepheline may be converted to sericite, which is pseudomorphous after the feldspathoid. Magnetite contains exsolved blades of ilmenite and assays TiO_2 30% and V 0.4%.

In addition to these minerals, søvite also contains rather regularly distributed barite and pyrochlore rich in columbium; the latter mineral is designated as koppite, although the name is not recognized by Fleischer. The formula for koppite is $(NaCa)_2(NbTa)_2O_6(OH)$. The composition probably is not constant as the color of this mineral ranges from colorless through red and brown to black; it normally is zoned. The koppite analyses range from Nb_2O_5 58% to 72%; Ta_2O_5 0.5% to 2.0%; TiO_2 0.5% to 0.6%; FeO 1.0% to 20%; CaO 0.5% to 20%; and ThO_2 + rare earths 0.5% to 20%. Locally, in the Cappelen mine, U_3O_8 is from 15% to 20% in the pyrochlore.

In the type locality, søvite is a rather coarse-grained calcite, the grains being two to four millimeters across; euhedral crystals of magnetite and koppite are as much as five millimeters long. In other places, the grain sizes are less. Three types of petrofabrics are known in the søvite where it is tectonically deformed. Barth and Ramberg believe that these show that the calcite was intruded in the solid state or, more likely, was later recrystallized by plastic flow.

Rauhaugite was divided by Saether (1957) into types I and II. Type I is a marble-like rock that weathers yellowish and is composed of 95% ankerite; it has the same additional minerals as søvite and the same fabric. Only subordinate amounts of carbonates transitional between søvite and rauhaugite are known.

Type II also contains ankerite as the most abundant mineral, but it also included appreciable chlorite and lesser biotite as dark dust (evenly distributed) or as irregular spots and veins. Magnetite and apatite locally may attain rather high concentrations. Hematite and pyrite are common. Quartz, albite, microcline, and particularly barite are seen in tiny clusters and veins; fluorite locally occurs in the same manner.

Rauhaugite II forms an irregular mass in the center of the sub-circular core of the carbonatite mass, but it has only vague and diffuse borders with the rocks that surround it. On the north side of the carbonatite body, Rauhaugite II has a broad transitional zone between it and fenite.

Barth and Ramberg say that dikes of rauhaugite are common, but they do not say if these dikes are rauhaugite I or II or both. If these dikes are all one type of rauhaugite or the other, they most probably would have been formed at the same time as the main mass of type I or II material was being produced from the søvite in these dikes and by the same solutions. If dikes of both types of rauhaugite exist, then it is probable that type I solutions affected those that are now rauhaugite type I and type II solutions those are now rauhaugite II. The rauhaugite dikes have the same manner of occurrence as those of søvite, but they may be more irregular in detail (as would be expected if they had formed by the replacement of søvite dikes); these rauhaugite dikes do, however, parallel the flows lines and margins of the søvite, again as would be expected if they originally were søvite.

Røberg is an ancient mining designation for a fine-grained red hematite-calcite rock, almost certainly of carbonatite lineage. It is principally present in the eastern part of the Fen complex; no Rødberg is known in any carbonatite area other than the Fen. This carbonate rock is composed of tiny grains (0.01 to 1.0 millimeters across) of carbonate that are essentially all composed of calcite; these grains are poikilitically flooded with tiny tablets of hematite (0.003 to 0.01 millimeters in the long dimension), and the calcite grains also are outlined by hematite grains of the same size. Hematite makes up 20% to 50% of the rødberg; its minor constituents are chlorite, apatite, and magnetite with quartz, albite, and barite as accessory minerals. Although it normally contains 0.05% to 0.5% of columbium, the columbium-containing mineral is not known; it also includes 0.2% of thorium and 1% of cerium rare-earths.

In rødberg, as well as in rauhaugite, Barth and Ramberg see exogenic relics of gneiss and of the Fen rocks. A metasomatic transition zone lies between rødberg and damkjernite, but the contact between rødberg and rauhaugite is knife sharp. Saether (1957) suggests that the rødberg was formed at rather shallow levels by the same "perimagmatic" solutions that produced rauhaugite (by replacement of søvite?) deeper within the crust.

Within the rødberg are dikes and veins of massive hematite that are not found outside the red-rock; they were economically mineable for over 300 years.

In manner of formation of the three carbonate-rich rocks it is not clear, but it appears most probable (to me) that the original carbonate rock of the Fen complex was a single, intruded mass of molten carbonate-rich magma that crystallized to form not only the søvite mass (then much larger than it is now), but also sufficient additional søvite to allow for its later conversion to rauhaugite and rødberg. This carbonate intrusion probably was separated as an immiscible melt from the kimberlite magma (from which the damkjernite also came). Some time after the intrusion of the molten carbonates and their solidification, part of the søvite body was invaded by iron-rich solutions from the carbonate magma chamber that converted the calcite of that portion of the søvite body to the ankerite that is the essential mineral of rauhaugite I. This process left the fabric and the minor and accessory minerals of the søvite almost entirely unchanged.

At much the same time, or, more probably, somewhat later, solutions from about the same source as those that produced the rauhaugite I invaded another volume of the søvite to form an irregular core in the sub-circular carbonatite body. The rauhaugite is without definite borders to the rocks that surround it; on the north side of the complex, it passes through a metasomatic transition zone into fenite. Where rauhaugite II is in contact with damkjernite, the transition is gradual. The main mineral of rauhaugite II is the same as in type I - ankerite, but the fabric has changed and new non-carbonate minerals - chlorite, biotite, magnetite, hematite, and pyrite plus quartz, albite, microcline, barite, and fluorite - were developed. This suggests that the rauhaugite II-forming fluid was moderately different from that that formed type I rock but not spectacularly so.

Still later than the formation of rauhaugite II was that of the rødberg, and the solutions that caused the change of another volume of søvite to rødberg were appreciably different from those involved in the production of the two types of rauhaugite. They did not convert the calcite of the søvite mass to ankerite, but they did add huge quantities of iron as inclusions in the calcite grains and around their borders. In both the production of rauhaugite and of rødberg impressive amounts of iron had to be added, though far more in the case of the rødberg than in that of the rauhaugites. In what different way the iron was transported or what were the parameters of the rødberg-forming fluid different from those of the fluid that produced the rauhaugite, the field evidence does not indicate, but some appreciable difference there must have been. Certainly, the retention of calcite as the carbonate mineral in the rødberg strongly suggests that the development of rodberg was a separate event from the formation of rauhaugite in both time and character.

Presumably all of the original søvite contained a columbium-bearing mineral and columbium is reported not only from the remaining søvite but also from the rauhaugite and rødberg, but the amount decreases in that order. This suggests that the solutions that created the two rauhaugites and the rødberg perhaps may have been capable of affecting the mineral character of the columbium mineral and were able, to some extent at least, to remove it from the new rocks. The solutions did, however, change the columbium mineral sufficiently that it is not economically recoverable from either rauhaugite or rødberg. The mineable columbium mineral in the remaining portions of the Fen complex that still are søvite (whether it is koppite or some other mineral) almost certainly was emplaced as an early separation-early crystallization component of the carbonatite magma and should, therefore, be classified as Magmatic Fla.

The hematite in the veins and dikes in the rødberg and flooded into the calcite crystals of which that rock mainly is composed very probably was deposited from hydrothermal solutions that had been generated in the same magma chamber from which the molten carbonate material of the remaining søvite and søvite now altered to rauhaugite and rødberg came. The hematite in the mineable dikes and veins should be classified as hypothermal in carbonate rocks (Hypothermal-2) to Mesothermal, and this is done here.

The "exogenic relics" of gneiss and other Fen rocks that Barth and Ramberg see in the rauhaugite and rødberg would appear to argue against these two rocks having been formed by replacement of søvite, but these may be remnants of such rocks engulfed in the original søvite and preserved in the transition from søvite to rauhaugite or rødberg.

In the course of his pioneering work on the Fen complex, Brøgger (1920) decided that the rocks of the complex were Jotnian (1400 million years) or Eocambrian (600 to 700 million years) or much older than the intrusives in the Oslo area. Saether (1957) produced age dates by the Th/Pb method of 420, 570, and 590 million years. Faul, and others, (1959) determined a K/Ar date on biotite from biotite-søvite of 565 million years. This is accepted as the most reliable date so far made and places the age of the Fen complex as earliest Paleozoic or, perhaps, as latest Precambrian. Early Paleozoic is given preference here.

FLÅT NICKEL, AGDER

Late Precambrian Nickel, Copper Magmatic-2a

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Notes

The Flåt nickel mine is in the parish of Evje on the northern slope of a hill (Mykleåsen); the mine is about 6 km NE of the town of Evje (58°36'N, 7°51'E) and is 65 kilometers north of the port of Kristiansand on the southern coast of Norway; the elevation of the mine is about 400 meters above sea-level. A smelter was operated near the Evje railway station (Bugge, 1978). About 10 additional nickel-bearing deposits of small size are located in the general vicinity of the Evje mine, but none of these is large enough to be of any economic importance. The Flåt mine production of nickel was at least twice that of all the other nickel mines in Norway. This mine was discovered in 1870 and operated from 1872 to the end of 1945 with two periods of idleness (1894-1899 and 1920-1927). In the latter years of the operation of the mine and for a while thereafter, much exploration was carried out in the area of the Flåt mine, but no success was achieved. Before 1917 (Bjørlykke, 1947), the ore was concentrated to give a smelting ore with a grade of 1.25 to 1.5 per cent nickel, and one per cent copper. From 1932 through 1940, the flotation plant provided a concentrate for smelting that contained three to four per cent nickel and nearly three per cent copper. The raw ore was concentrated by about 5.7 times, and the mill recovered about 82 per cent of the nickel and 90 per cent of the copper with the amount of the concentrate being about 30,000 tons a year. As mined, the ore averaged 75 per cent silicates, 12 to 13 per cent sulfides, about eight per cent magnetite, and four per cent apatite. The total mine production during the period 1927 through 1944 was about two million tons of raw ore in which the average grade was 0.69 per cent nickel and 0.475 per cent copper. The total production from 1872 through 1945 was 2,700,000 tons with 0.72 per cent nickel and 0.48 per cent copper. At the beginning of the shorter period, the nickel grade was 0.95 per cent; by 1944 it had dropped to 0.55, but copper content in relation to nickel rose somewhat. The ore differed appreciably in appearance from one part of the mine to another; ranging from nearly 100 per cent sulfides to dioritic rocks that contained only a slight impregnation of sulfides. The relation of copper to nickel differed considerably from one part of the mine to another, but the relationship of pyrrhotite to pentlandite was quite constant. In the raw ore, the ratio of nickel to cobalt averaged about 8:1, but rose to 12:1 in the concentrates, indicating an appreciable loss of cobalt in the concentration process; why is not stated.

The rock that contains the sulfide concentrations at Evje is designated as an "amphibolite" by Barth (1947) and by Bjørlykke (1947) as a "quartz diorite" or as the "ore diorite". Barth says that the amphibolite "swims in a sea of congealed granite ichor". If the definition of "ichor" as given by the "Glossary of Geology" is correct, then the amphibolite is surrounded by a congealed fluid that, at least when the term was first used, carried the connotation that it had been derived from a magma. If Barth means that the ichor was originally a fluid (presumably largely water), what he says about congealed ichor can hardly be sensible since such an ichor if congealed would mainly be ice. However, be that as it may, Barth believes that the rocks surrounding the Evje amphibolite are gneisses that have been more or less granitized; for the present here, this problem can be left at that.

All of the rocks in the Evje area are Precambrian in age and probably part of the Telemark group; this group appears to be oldest late Precambrian. Barth considers all the Evje rocks to have been considerably to largely metamorphosed. Presumably, by amphibolite Barth means what the Glossary of Geology defines as amphibolite, that is, a crystalloblastic rock made up mainly of amphibole and plagioclase with little or no quartz; if the quartz content becomes appreciable the rock becomes hornblende-plagioclase gneiss. The amphibolite also seems to be considered by Barth as a metanorite or perhaps a metahyperite (hyperite being a rock composed of orthopyroxene, plagioclase, olivine, and clinopyroxene and being intermediate between norite and gabbro - the Glossary of Geology does not recommend the use of the term). Vogt (1893) says that there was no olivine in the Evje rock, whatever it may be otherwise, so it must have been a gabbro or a norite, more probably the latter. In his map of the Evje area, Barth's legend described the Evje mass as gabbroidal rocks and amphibolites, so this would seem to require the Evje rock before metamorphism was a gabbro.

On the contrary, Bjørlykke calls the rock containing the ore "ore diorite" but says that it differs from the surrounding rock only slightly in chemical and mineralogical composition. This surrounding rock he calls the Mykleås type (or diorite or quartz diorite or amphibolite). Of course, as the principal femic mineral in diorite is amphibole, particularly hornblende, the principal reason for designating the Evje rock as an amphibolite and not as a diorite would be if it had a crystalloblastic texture. Apparently Barth considers the texture to be crystalloblastic and Bjørlykke does not.

If the Evje rock was a gabbro (or norite) before it was metamorphosed, the pyroxenes in the gabbro were converted to amphibole (hornblende) and the primary plagioclase was converted to a less calcic variety. This conversion

Barth refers to as "granitization", this having been effected by "ichors" derived from some source at depth, and the depth at which this entire process took place must have been considerable. In some places in his paper, Bjørlykke refers to his quartz diorite or ore diorite as "ore gabbro" and to the Mykleås type of quartz diorite as "the older gabbro of the Mykleås type". All this is rather confusing, but it may be summarized (and simplified) by saying that the amphibolite is, in composition, essentially somewhere between gabbro and quartz diorite and that it either has been metamorphosed from gabbro (or norite) to amphibolite (of much the composition of quartz diorite) or Evje rock was intruded in two stages, one of an earlier gabbro or quartz diorite (Mykleås type) and the second of the ore diorite (or ore gabbro) that indicates its later intrusion by the fine-grained or porphyritic texture shown by the contacts of the ore diorite against the Mykleås-type quartz diorite.

The greenish-black hornblende in the ore diorite gives its characteristic dark color to that rather fine-grained rock. The hornblende crystals are anhedral and contain poikilitic minerals (quartz and possibly plagioclase?). The chemical composition is quite complex, with the cation groupings being [Na,Ca], [Mg,Fe,Al], and [Al,Si] and the anions [O,OH,F]. The hornblende makes up, on the average, 33 per cent of the rock. Plagioclase averages 30.5 per cent of the ore diorite and is in polygonal grains in which curved lamellae are known and zoning is very irregular. The composition of the plagioclase is about An_{35}, but lesser An has been noted. Biotite constitutes 14.5 per cent of the rock and has dark to light brown pleochroic colors; quartz averages 8.5 per cent; ore minerals, including magnetite, about 9.5 per cent; and apatite four per cent. The apatite is in large (?) crystals (one to 10 mm long). Titantite has been seen, and zeolites, laumontite and apophyllite, have been found.

The ore diorite gradationally passes into the schistose Mykleås type of quartz diorite (or amphibolite) that, in most places, has a gneissic texture. The plagioclase composes 54.5 per cent of the Mykleås rock and always is andesine; it has sharp polygonal outlines and curved lamellae. The hornblende is in elongated angular crystals with curved outlines and, as in the ore diorite, is poikilitically penetrated by quartz. The hornblende tends to grow in clusters of several individuals and these in many instances are pulled out into long streaks that produce banding on a microscopic scale. In some thin sections, a different type of hornblende is seen that is porphyroblastic and is accompanied by zoisite; the pleochroism of this hornblende is much paler than the main hornblende. This lighter hornblende also contains poikilitic quartz. Locally, biotite is appreciable; garnet is rare, apatite is seen in every thin section, titanite also; ore minerals are minor in amount and present only irregularly. The percentages of the minerals other than plagioclase in the Mykleås-type rock are: biotite, 11.5; hornblende, 10; quartz, 6.5; muscovite, 5.5; apatite, 4; titanite, 2; and ore minerals 6. The Mykleås-type hornblende is slightly richer in iron and a little lower in aluminum than that in the ore diorite; it does not appear to contain fluorine.

The Evje amphibolite is well-known for the many pegmatites that are present throughout its extent. These pegmatites are famous for their unusual minerals in addition to the quartz and feldspar that are the major minerals and for which the pegmatites are mainly mined. Among the uncommon minerals are thortveitite [$(Sc,Y)_2Si_2O_7$], beryl, gadolinite [$(Ce,La,Nd,Y)_2Fe^2Be_2Si_2O_{10}$], euxenite [$(Y,Ca,Ce,U,Th)(Nb,Ta,Ti)_2O_6$], thalenite [$Y_3Si_3O_{10}(OH)$ (?)], and xenotime. These minerals are hand sorted from the broken pegmatite rock and sold abroad as mineral specimens. Barth believes that these pegmatites are the result of: (1) fracturing of the amphibolite, (2) migration of dispersed ions in the amphibolite (those ions not readily accepted in the amphibolite minerals) into these fractures and consolidation of these materials into pegmatites. A more reasonable explanation, I think, would be partial melting of the amphibolite to produce a parent melt for the production of the Evje pegmatites and their crystallization in place within the amphibolite where they were developed. An alternate approach, though less likely, would be the production of melts in the gneiss and their migration into the amphibolites. If this explanation is correct, then it is difficult to understand why what small amounts of pegmatite are in the gneiss are always near the gneiss-amphibolite contact and never appreciably removed from amphibolite. The lack of any feeder channels for the pegmatites in the amphibolites would seem to confirm their development by partial melting of the amphibolite, essentially in situ.

Barth (1947) is convinced that essentially all of the original parent of the amphibolite was completely recrystallized during post-intrusion metamorphism. Only a few relics of primary hypersthene have been found in the amphibolite, and these are not near the Flåt ore body. Thus, Barth considers that all the minerals now in the amphibolite were produced by metamorphism and were not the result of a minor recrystallization of the primary mafic rock. Most of the amphibolite is cleavable and resembles a gneiss in splitting along wavy surfaces. This is caused by the minerals of the rock arranging themselves in thin streaks parallel to the schistosity; this is aided in places by the feldspars and hornblende segregating into alternate, locally wavy, bands. Barth believes that this banding was due to differential movements (metamorphic differentiation) that mechanically separated the various minerals on the basis of the gliding properties. Certainly there could have been no such primary banding in the source rock of the amphibolite so this explanation of banding being due to metamorphic differentiation must stand until a better is produced.

In essentially all places Barth observed, the mineral parallelism and the grooves on the schistosity planes point in the same direction, but he thinks that the grooving and wavy corrugation of the amphibolite developed after the metamorphic recrystallization. Further, appreciable portions of the amphibolite, particularly in the peripheral areas of the rock, were plastically deformed. Despite this, Barth says that his observations show that the central parts of the amphibolite were more rigid than the surrounding gneiss, being deformed by gliding, faulting, and slippage, nor is true folding found in the central area of the amphibolite. He is convinced that the amphibolite is usually gneissic with the strike of the body differing considerably from place to place although normally it is northerly. No crests of folds are to be seen in the central amphibolite, but, from the strike and dip of the schistosity, and the trend and pitch of the linear structures that developed on the planes of schistosity, Barth could construct fold axes on the assumption that all folds are shear folds. Barth believes that the folding that recrystallized the gabbro to develop amphibolite was earlier than the folding that caused the grooving. Barth believes that the original position of the amphibolite was with its long axis northeast-southwest, not the north-south position it now has. During the main phase of conversion of gabbro to amphibolite (Barth calls this granitization) the then plastic mass was compressed in an east-west direction and was completely recrystallized. After a following drop in temperature, the amphibolite became more rigid than the still plastic gneiss that surrounds it. At this time, the amphibolite was rotated into its present position. No further recrystallization (or essentially none) took place during the rotation stage but slippage did occur along the long limbs of the folds, producing striations and grooving on the planes of schistosity. Most of the major tectonic features and mylonitic structures, however, appear to be younger than the granitization and strike generally northeast-southwest, a direction that fits with the regional system of faults in the area. Because the gneisses reacted as brittle rocks during this regional faulting, they must have been much cooler than during granitization and were, when the faulting took place, at a much higher level in the crust than they were during the granitization.

Any discussion of the manner of genesis of the Flåt nickel ores must first summarize, and then attempt to reconcile, the widely divergent views of Bjørlykke and Barth. Bjørlykke believes that the quartz-diorite magma, from which the ore diorite crystallized was intruded into the older quartz diorite of the Mykleås type. The ore-diorite magma not only contained the silicate minerals that define the rock as a quartz diorite but also an appreciable amount of sulfide material that was initially, or soon became, immiscible in the silicate melt and separated from it. In part immiscible sulfide melt settled under gravity toward the bottom and lower surface of the ore diorite magma, concentrating around gneiss inclusions in the amphibolite. The silicate minerals apparently crystallized before the sulfides, so that these latter minerals fill spaces left among silicate minerals or actually developed sizeable masses of essentially pure sulfides.

Barth, on the contrary, argues that the Evje-Iveland amphibolite originally was a gabbro (or possibly a norite) that later was affected by fluids given off from some more silicic and crystallizing magma at great depth. These fluids affected the gabbro during a period when tectonic forces were acting strongly on the gabbro, and converted the gabbro to amphibolite. Still later hydrothermal fluids entered the amphibolite and deposited the sulfides, largely by replacement.

The solution of this disagreement, if one is possible, must follow a more detailed discussion of the ore minerals and their relation to the silicate minerals or the ore-diorite (or amphibolite). The principal ore minerals are: pyrite, pyrrhotite, pentlandite, and chalcopyrite. No low-temperature minerals that might indicate at least a late-stage hydrothermal phase are present (Bjørlykke). No arsenic-bearing minerals had been found according to Bjørlykke. Some secondary nickel minerals are present in the ore; they are alteration products of the primary mineralization process and are not the result of supergene activity. These minerals include millerite (NiS) and violarite ($FeNi_2S_4$). The oxygen-bearing minerals are magnetite and ilmenite, which latter mineral contains exsolved laths of hematite. Magnetite is not directly intergrown with the ore minerals but instead accompanies the silicates; no titanium nor vanadium has been reported to be in the iron oxides.

The pyrite contains considerable quantities of cobalt and the range of cobalt in pyrite is between 1.00 per cent and 0.60 per cent with an average of 0.77 per cent. The cobalt is homogeneously distributed throughout the pyrite. The cobalt content is much higher than that in other Norwegian pyrites. Nickel appears to be essentially lacking in the pyrite.

Pyrrhotite is the most abundant sulfide mineral in the Flåt nickel ores. It normally is present in fine-grained masses, but, where it is coarser, it is darker brown in color. The pyrrhotite always is associated with pentlandite. Most of the pentlandite is in small granular stringers around pyrrhotite grain margins, suggesting that it reaches these loci by exsolution. Pentlandite also occurs as thin, vein-like bodies in the pyrrhotite; these may be the result of exsolution or of replacement, but which is not certain. The pentlandite in pyrrhotite also may be as lenses, oriented blades, and flakes; these also probably are textures produced by exsolution. The pentlandite is so thoroughly mixed in the pyrrhotite and is so finely divided that magnetic separation of the ore mineral from the other is shown to be impossible. The Ni:Co ratio in pentlandite is about 80:1. Chalcopyrite, as is normal in such sulfide ores, appears to have separated from pyrrhotite-pentlandite melt at an early stage in the crystallization of the Fe-Ni melt; this chalcopyrite melt remained molten long enough to have been the latest mineral to crystallize; it appears to have replaced sulfides, particularly pyrrhotite, and it also fills fissures in the surrounding silicate minerals.

Ilmenite in many places occurs as well-developed crystals in pyrrhotite, and, under the microscope, laths of hematite are seen in the ilmenite. Magnetite, as has been stated, is abundant in the ore diorite in general and in the silicate minerals associated with the sulfides, but is not intimately intergrown with sulfides. The magnetite which made up about eight per cent of the ore, contained about one per cent nickel and very small amounts of cobalt and even less chromium. The ore as mined contained about 1.7 per cent P_2O_5; concentrate containing 28 to 30 per cent P_2O_5 was produced and was used in the making of superphosphate. In the last years of the mining operation about 3000 tons of this concentrate were produced annually.

In plan, the ore body strikes at the surface about north-northeast and dips downward from the outcrop at about 45° south-southeast. The strike length of the ore body is 100 to 150 meters and the horizontal thickness ranges from 13 to 30 meters. Beginning at the 282-meter level downward, the ore body gradually flattens out, and the vertical thickness increases from a minimum of 30 meters to a maximum of nearly 100 meters. The total length of the ore body along the ore axis is about 900 meters. Bjørlykke suggests that the ore body on the lowest levels splits in two, with the more western of the two splits being about 20 meters lower than the eastern; this relationship is not clear from his diagrams.

Bjørlykke says that the ore body is made up mainly of silicate rocks impregnated by sulfides; this means, of course, that the amount of massive sulfide accumulated must be small. The average mined ore was made up of 75 per cent silicates, 12 to 13 per cent sulfides, eight per cent magnetite, and four per cent apatite. The composition of the ore mined did not differ appreciably from year to year. Despite this constancy in composition, some ore masses were nearly 100 per cent sulfides, but these could not have been important in amount or the grade would have been much higher than it was. The sulfide minerals in the smelter concentrate, however, were calculated by Bjørlykke as: pyrrhotite, 70.77 per cent; pentlandite, 14.44 per cent; chalcopyrite, 11.48 per cent, and pyrite, 3.31 per cent; it must be remembered that this smelter concentrate is enriched in nickel relative to the raw ore. The average nickel content of the raw sulfides, however, must have been about 4.5 per cent.

The concept put forward by Bjørlykke (and Carstens) is that the ore diorite was intruded as a magma enriched in sulfides into the older quartz diorite (Mykleås type); during the cooling of this sulfide-enriched silicate melt, the sulfides separated in the molten state (if they were not already immiscible to a large extent when they were intruded) and segregated along the margins of inclusions of gneiss in the ore diorite. From the rather large proportion of the sulfides that are randomly distributed through the ore diorite and a rather few masses of silicate-free sulfides, this separated molten-sulfide material did not have much time to settle toward the bottom of the magma before it was halted in place by the crystallization of the silicate minerals. These authors base their theory on the following points that they believe they have established: (1) the ores show an order of crystallization that is what would be expected in a cooling silicate melt of the composition of the supposed ore-diorite magma; (2) the absence of any low-temperature minerals such as would be expected in the late stages if the ore minerals had been introduced by hydrothermal solutions, (3) the high content of Co that ranges from 0.5 to 1.0 per cent in the pyrite; (4) the distribution of Ni and Co in the different types of ore seems to these authors to indicate that silicate and sulfide crystallization took place from the molten state; and (5) the considerable amount of hydrothermal alteration of the ore where it borders on granite pegmatite indicates that the ore minerals were unstable against attack by the fluids given off from those pegmatitic melts. This suggestion, however, does not necessarily suggest that the pegmatic magmas had a magmatic rather than a palengenic manner of development. This point - the genesis of the pegmatites - probably is the major problem in accepting Bjørlykke's idea that the quartz-diorite ore magma was intruded and crystallized as such rather than having been developed by partial melting of the amphibolite.

Barth believes that the original rock was a gabbro and that this gabbro later was altered to amphibolite (or quartz gabbro) by intense metamorphism at a considerable depth beneath the surface, so deep that solutions generated there converted the gabbro's feldspar to a more sodic variety (andesine) and turned the pyroxenes to amphibole (hornblende). At the same time, essentially, some of the altered gabbro was partially melted and converted to local pockets of magma that crystallized to pegmatite or aplite, depending on whether these pockets were able to retain water or not. Barth thinks that, when all this metamorphism was completed, or in the late stages of it, hydrothermal fluids were somewhere generated that moved into and through the amphibolite (quartz diorite) to deposit the sulfide minerals as random replacements in the quartz diorite.

Bjørlykke, however, points out that the reactions between the pegmatite bodies and the ore sulfides (where there happened to be in contact) shows that the sulfides were in place before the pegmatites were introduced into the area.

It seems possible to reconcile the views of both Barth and Bjørlykke by using something from the ideas of each. In other parts of southern Norway, most of the mafic rocks have been converted to amphibolites (quartz diorites) by subsequent metamorphism. With certain modifications, this history of the production of the amphibolites seems more reasonable than that, in the one place (Evje-Iveland), an igneous magma was intruded that by coincidence happened to be of the same mineral and chemical composition as all the other amphibolites in this portion of Norway.

Thus, I suggest that in the Evje-Iveland area, a gabbro was intruded at essentially the same time as this was done in the rest of southern Norway. What the rock was into which the gabbro was intruded is not certain, but certainly during the period when the gabbro was being converted to amphibolite (quartz diorite) the surrounding rocks were metamorphosed to gneiss. But, before this metamorphism occurred, a second intrusion of gabbro was forced into the region, and this was a sulfide-rich gabbro that was placed more or less in the center of the early gabbro. Even at this time, it appears probable that the second gabbro intrusion was rich enough in sulfides that, had no further metamorphism intervened, the second gabbro would have been ore, if it had been brought to the surface in that condition. Before such upward movement occurred, however, the gabbro was probably buried even more deeply than the elevation at which it was intruded, and there, at depth, it underwent metamorphism to amphibolite, a process that was accompanied by partial melting, encouraged, if not solely made possible, by the introduction of "granitizing fluids" and thereby produced material that was essentially granitic-pegmatic in composition, much of which was later solidified in place as pegmatite but some of which was moved into veins where it formed aplite. Some of this pegmatite crystallized in the vicinity of sulfide ore, and fluids given off from the crystallizing pegmatite to some extent altered the sulfides, forming such minerals as millerite and violarite.

Now the amphibolite is a massive rock, which I presume is why Bjørlykke designated that rock as ore diorite (quartz diorite = amphibolite) and thought it to be a primary igneous rock. Since the silicate minerals have such an equigranular texture, different from the surrounding gneiss, this would be another reason for assuming that the ore diorite was a primary rock. Barth, however, thinks that the lack of schistose structure in the ore diorite must be ascribed to the rock having recrystallized under static conditions; probably, he believes, the rock simply found itself in the protective shadow of tangential forces and was shielded thereby from stress action. The surrounding Mykleås-type of quartz diorite (amphibolite) was not so shielded and acquired the schistose structures it now possesses. The country rock into which both gabbros were intruded seriatim, however, was so strongly metamorphosed that it developed its present gneissic structure. The ores were part of the same geologic event as the original intrusion and not the product of some later hydrothermal phase of unknown source; that is, the sulfides came in with the second gabbro and were little disturbed as to place or composition by the metamorphic processes that converted the gabbro to amphibolite (quartz diorite) and blurred the contact between the older and younger altered gabbro (amphibolite).

Because the metamorphism that the sulfide-bearing quartz diorite underwent after it was intruded does not appear to have changed appreciably either the sulfide content or composition (except for minor development of millerite and violarite), the metamorphic processes should not be included in the classification of the ores. The sulfide melt almost certainly separated in large part from the molten silicates quite early in their career as part of the ore gabbro magma but did not crystallize, particularly the chalcopyrite, until late in that cycle of magma solidification. This means they should be classified as Magmatic-2a, that is early segragation, late solidification.

EGERSUND AREA (ÅNA-SIRA ANORTHOSITE MASSIF), ROGALAND

Late Precambrian	Iron as Magnetite (with Vanadium) Titanium as Ilmenite Minor Iron as Hematite in Ilmenite, Nickel, Copper, Cobalt in Sulfides	Magmatic-3a Magmatic-3b

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Notes

The Egersund (58°37'N, 6°01'E)-Åna-Sira district consists largely of a mass of intrusive igneous rocks about 75 km long in a northwest-southeast direction and about 25 km wide in its maximum dimension. The major igneous rock is anorthosite but appreciable areas of farsundite, mangerite, and norite intrusions also are present. The total area of the complex of these rocks is about 1200 km^2 of which anorthosite makes up about 1000 km^2. Several subordinate anorthosite massifs are recognized but these are differently named and, to some extent, differently bounded by different authors. Bugge (1978) lists five massifs of which four (Egersund-Ogna, Håland and Helleren, Ana-Sira, and Hidra are arranged in order from northwest to southeast. The Orrestad massif is situated almost due north of the Ana-Sira; the Orrestad and Hidra bodies are much smaller than the other three. Between the Håland and Helleren and the Ana-Sira massifs is an area of complex intrusive bodies about 5 km wide at its seaward end (against the North Sea) but some 12 km to the north it widens considerably to have a maximum northwest-southeast length of over 40 km. The core of this mass is composed of mangerite (a hypersthene-bearing monzonite with charnockite affinities). This core is partly surrounded (to the northwest and south) by foliated norite and (on the northeast) by a norite sheet. On the west side of the prong that extends north from the North Sea, is a thin mass of mangeronorite, wider at its northern end, that is about 20 km long. The Hidra massif (at the southeastern end of the complex) is almost entirely surrounded by farsundite (a hypersthene- and hornblende-bearing granite of charnockite affinities). The entire complex is surrounded by gneisses and migmatites derived from granitized metasediments that have been metamorphosed to the granulite facies.

The first ore bodies to be discovered in the complex (Bjåfjell and Koldal) are located in the Ana-Sira and Håland and Helleren massifs, respectively, the first in norite pegmatite and the second in leuconorite. These deposits were discovered in the 18th century but major mining did not begin until 1864 to 1876; most production from these mines (and the Anker mine associated with the Koldal) was in the years just mentioned, but small-scale mining continued until 1918. In that year, modern mining began in the Storgangen deposit that lies along the boundary between foliated norite and the Ana-Sira massif and, farther north, passes into the Ana-Sira massif proper. This mine was worked from 1918 to 1965 and produced 12,000,000 tons of ore that ran 17 per cent TiO_2 and 20 per cent Fe. Diamond drilling has proved a further 60,000,000 in this mine-area.

In 1950, the owners of the titanium-bearing area, A/S Titania, began a program of systematic geological exploration that was supplemented by an airborne geophysical effort beginning in 1954. In 1954, the Tellnes body was discovered for which 300,000,000 million tons of reserves averaging 18 per cent TiO_2 were proved; mining operations began on Tellnes in 1960. In 1975 (Bugge, 1978) production for the year was 840,000 tons of limenite concentrates and 46,000 tons of magnetite concentrates that contained 65 per cent Fe and 0.6 per cent vanadium. The sulfur in the Tellnes ore runs 0.2 to 0.3 per cent, and the sulfides are recovered by floatation, 13,000 tons being produced in 1975 that averaged 4.5 per cent Ni, 1.6 per cent Cu, and 0.8 per cent cobalt.

The anorthosite massifs composed at least two-thirds of the total igneous (South Rogaland) complex and include not only andesine anorthosite but also leuconorites. The remainder of the complex is norites and mangerites. As has been mentioned, the farsundite occupies only a small area at the southeast end of the complex.

Bugge (1978) believes that the rocks of the various massifs are all parts of the same pluton. The Egersund-Ogna massif is considered by Bugge to be the oldest of the series; it has somewhat of a dome shape and dips out under the surrounding rocks. In this oldest anorthosite, the quantities of ferromagnesian minerals increase toward the margins of the body, and the feldspar is an andesine with 44 per cent An. The rock has a medium- to coarse-grained texture, and its accessory minerals are hypersthene, ilmenite, and magnetite.

The sheet of norite (the Lakssvelefjeld-Koldal) lies around much of the Egersund-Ogna massif and is principally norite and leuconorite, but locally contains charnockitic gneisses and banded gneisses with quartzites. Bugge believes that the norite is a differentiation product (later) of the parent magma of the two rocks. Michot and Michot (1969), however, suggest that the norite is a marginal intrusion of material derived from the ocean crust that was introduced along the discontinuity between the anorthosite of the basement and the sedimentary gneiss cover (now removed by erosion). This noritic sheet contains inclusions of both anorthosite and of the original cover rocks. This sheet has a usual width of 100 to 200 m but, in places, may be as much as 1000 m wide. The Koldal portion of the sheet was emplaced along the southern border of the Egersund-Ogna massif and ends against the body of the Håland massif. Prior to the intrusion of this second massif, the Koldal sheet may have curved north to connect with Lakssvelefjeld portion. The noritic sheet is migmatized. Some small lenses and irregular schlieren of ilmenite-magnetite ore are in close juxtaposition to this sheet.

The Michots think that the differentiated series of rocks on the Rogaland complex resulted from simultaneous migmatization and metamorphic differentiation at deep levels in an orogen. In these processes, the leuconorite was changed to anorthosite when the ferromagnesian constituents of that rock were mobilized; the Fe-Mg components rose up as a mafic front that is now represented by schlieren rich in hypersthene or oxide minerals.

The Håland massif is located to the southeast and south of that of Egersund-Ogna. In its northern areas it is made up of anorthosite and leuconorite that the Michots believe was related to magmatic remobilization of leuconorite and norite material; the remobilization cleaned out the Fe-Mg constitutents of the parent rock, leaving an essentially pure anorthosite behind. They have called this process basic (mafic) palingenesis and have called the rock produced para-anatextic anorthosite. The southern part of the Håland massif has been less affected by this process (or the parent magma was less well differentiated) and is much like the upper portion of the Egersund-Ogna rock.

The Helleren massif contains a coarse-grained (para-anatectic?) anorthosite that encloses masses of leuconorite and norite. The Helleren rocks in places cut into both the Egersund-Ogna and Håland massifs. Ana-Sira, the Hidra, and the Orrestad (Outlier) massifs are quite similar to the Helleren; Michot thinks that they also are para-anatectic anorthosites (although again the possibility exists that they are only better differentiated than verging on leoconorite).

The Ana-Sira massif is the largest (Bugge, 1978) massif (although, from Bugge's map, 1978, this does not appear to be so); it certainly is economically the most important since it contains all of the Tellnes and Blåfjell ore bodies and almost all of the Storgangen. The Ana-Sira also is noted for its diversity of rock types and structures. The anorthosites in this massif range from medium-grained rocks (some of which contain porphyritic crystals of andesine of huge size) to rocks made up of aggregates of smaller (but still large) andesine crystals. These anorthosites grade gradually into hypersthene-bearing rocks in which the hypersthene crystals may be as much as 40 cm long. Norite pegmatites are coarse in grain and are composed of andesine, hypersthene, and ilmenite; they cut and enclose major xenoliths of anorthosite in the vicinity of the old, now exhausted, Blåfjell mine. The pegmatite can be traced southward from this mine for several kilometers and contains a number of small occurrences of ilmenite ore.

Despite the opinions expressed by the Michots as to the manner of formation of the anorthosites (their para-anatectic anorthosites), Bugge (1978) remains of the opinion that the entire process from anorthosite through leuconorite to norite to norite pegmatites are the results of a sequential differentiation process. [After two visits to the area, I am convinced of the correctness of Bugge's concepts.]

The Hidra masiff is a small and separate anorthosite body at the extreme southeast end of the Rogaland complex. To the east and south it is surrounded by farsundite; these are charnockites (or charnockite-like) rocks; it is separated from the Ana-Sira massif by a narrow band of farsundite (nearest the Hidra rocks) and by a narrow band of norite-leuconorite (nearest the Ana-Sira massif).

The Orrestad (or Outlier) massif is a second definitely separate anorthosite-leuconorite body. Bugge is convinced that the Orrestad mass is a southern continuation of anorthosites that form the bottom layers of complex intrusive body that separates the Håland and Helleren massif from the Egersund-Ogna; about 12 km from the sea, this complex turns to the northwest and broadens to a maximum width of about 12 km. In structure this complex is a lopolith, the flanks of which dip inward at quite steep angles (50° on the east limb, 80° on the west limb). Surface mapping places quartz-mangerite in the core of the synform, but drilling appears to have found anorthosite at depth. This complex mass is called the Bjerkrem-Hauge series as the town of Hauge lies near its southern extremity and Bjerkrem at its northern.

Bugge (1978) describes the Bjerkrem-Hauge complex as a layered eruptive lopolith, the succesion of rocks in which is (from the bottom upward): (1) anorthosite with disseminated ilmenite and a few ilmenite-bearing schlieren; (2) a banded zone made up of alternate layers of leuconorite and norite - it is as much as 5000 m thick in the axial zone but thins to less than 1 km on its margins; west of the Storgangen ore body, the banded norite of the Hauge norite lies conformably on the Ana-Sira anorthosite and grades down into it; in other places, the boundary is discordant as the result of local faulting or other disturbance that occurred when the Hauge norite was intruded. Hauge norite contains not only norite but also ilmino-norite, leuconorite, and anorthosite all of which are gradational to the rocks above and below them; (3) mangero-norite (known as the Juton-norite) that is a microcline, quartz-bearing norite that is normally conformably intercalated in the lower part of the Hauge norite or beneath it - in places the Juton-norite definitely intrudes the Hauge; (4) mangerites and quartz mangerites that are the central rocks of the synclinal (lopolithic) body; these rocks locally contain Fe-rich olivine and diallage, minerals that are minor in the anorthosite-norite series - the plagioclase contains 30 to 35 per cent An and usually is antiperthitic.

Although opinion differs among the various writers who have discussed the genesis of the Rogaland igneous complex, there is agreement that the rocks were developed partly by magmatic melting, partly by anatexis (used here apparently in the sense of partial anatexis), and by metasomatic reactions. Bugge appears to favor the majority of the rocks of the complex having been derived from magmas that were completely molten at some stage of their history, the various relations of the different rocks of the complex being most readily explained in his mind by the sequential differentiation after intrusion as a leuconorite magma. The age of the host charnockitic migmatites and gneisses probably is much older than that of the rocks of the differentiation sequence. Ages obtained by various isotopic methods for the rocks of the complex range from 1500 to 850 m.y.; the youngest date is that of the final consolidation of the last of the Rogaland units, the farsundites and the mangerites. The orogenesis which produced these rocks is considered by Krause and Pedall (1980) to have been the equivalent of the Grenville orogeny in North America. Thus, the age of the ores of the Egersun-Ana-Sira area is Late Precambrian as, of course, is true for practically all anorthosites in the world.

The most important tectonic element (Krause and Pedall, 1980) on a regional scale in the area is a system of folds with more or less north-south axes that are, however, not strictly parallel. Within the intrusive complex, Michot, (1960) distinguishes two additional phases of folding; in the gneiss area around Flekkefjord, Falkum (1972) was able to separate out five systems of folding. Presumably, however, the gneisses at the south-eastern end of the igneous complex had undergone more deformation than had the younger intrusions. Falkum also pointed out that the form and structure of the whole southern tip of Norway could be considered as a megatectonic synclinorium with the north-south axes plunging to the north.

Krause and Pedall (1980) say that all the rocks and ore bodies of the South Rogaland complex can be assigned to one cycle of magmatic-tectonic development. This cycle they divide into three sections on the basis of geological and petrographic criteria that ranges through. This sequence - anorthositic, noritic, and mangeritic - follows the chronological order of their differentiation. These authors further divide the noritic and mangeritic portions of the intrusion by their spatial-structural relations to the anorthosite series. They term those parts of the noritic and mangeritic rocks that lie outside the Ana-Sira anorthosite, the Bjerkrem-Sokndal lopolith or its eastern equivalent, as external units, whereas those within the central portion of the anorthosite of Ana-Sira are called internal. Connecting links between the mangeritic or noritic externides and internides are formed by dikes of these rocks, respectively.

In all of the rock series, local concentrations of iron-titanium oxides can be found, but economically mineable ore bodies are present only in the noritic or mangeritic rock series.

The form, location, and extent of the noritic and mangeritic bodies are the result of the dome-shaped structure of the anorthosite series. The structures of these younger intrusions, however, depend largely on the form elements of the anorthosite massif. The deformation of the three series was determined by their depth of burial; the more deeply they were buried, the more likely they were to have deformed plastically, the less deeply, the more probable this was to have been by fracturing. Different fabric types in the various rocks depend on their spatial position within each series and upon the composition of the rocks. To a greater or lesser extent, orthomagmatic structures were obliterated by secondary deformation. With minor exceptions, the secondary foliation was superimposed conformably on the primary structural features. Gneissosity developed in the rocks of the complex with wide areas having been secondarily foliated only to a minor degree. Deformation by plastic flow is characteristics of the leuconorite layers in association with anorthosite, with rock deformed by fracturing. Where the two interface, small-scale crumpling was produced; during the doming of the anorthosite, this crumpling changes locally to shear folding. In such deformation, anorthosite layers may have been separated into single boudins. Krause and Pedall (1980) say that this competent behavior of the anorthosite, in contrast to that of the leuconorite, in syn- to postconsolidative deformation causes the widely developed migmatite-like structures seen on the maps and in the field.

Essentially all of the known ilmenite deposits are located in the Håland, Helleren, and Ana-Sira massifs and in the Hauge norite. These rocks are, in Bugge's (1978) opinion, separate parts of the same pluton with the anothosite layer, as represented by these massifs, being overlain nearly conformably by the banded and layered Hauge norite. The ilmenite-rich portions of these rocks simply are produced by the last stages of magmatic differentiation.

The 90,000 tons of ore that came from the Koldal and Blåfjell mines from 1864 to 1876 was produced by an English company that attempted to use it in the production of pig iron; it is surprising that this effort continued for as long as it did. The opening of the Storgangen mine coincided with the first use of TiO_2 as a pigment, but production was only a small scale until 1930 when the annual output soared. Between 1935 and 1947, ilmenite-concentrate production ranged between 40,000 and 70,000 tons a year, and magnetite concentrates were produced at between 7000 and 15,000 tons. In 1947, the production took another upward surge, in 1959 reaching a total of 720,000 tons of ore from which 225,000 tons of ilmenite concentrates were obtained and 40,000 tons of magnetite concentrates as well. By 1975, the annual production of ilmenite concentrates had risen to 840,000 tons and of magnetite concentrates to 46,000.

The Storgangen mine is an irregular, but dike-like mass of ore, in part banded with foliated norite and in part as dense bodies that are almost entirely composed of ilmenite. In the western part of the deposit, the ore was concordant with the foliation of the norite, but, in the eastern part, the ore breaks away from the formation boundary and cuts through the Ana-Sira anorthosite. In places, the Ana-Sira rock is brecciated next to the ore vein, and ore penetrates the anorthosite of the hanging wall as veinlets. Within the Ana-Sira massif, the ore body strikes in an easterly direction and dips off to the north at 45° to 55°. Dybdahl (1960) reported that diamond drilling determined that the ore flattened out to about 30° to the north with depth.

The area of the Storgangen ore is about 80,000 m^2. Owing to the banding of rich ore with much lower grade ilmenite-norite, the mineralogical composition within the dike-like ore body and its metal analyses differ over a wide range. No analysis of this ore later than 1958 is available to me, but what follows should give a good idea of its chemical and mineralogical composition.

SiO_2	27.9%	V_2O_3	0.11%
TiO_2	18.58%	Cr_2O_3	0.02%
FeO	19.50%	P_2O_5	0.042%
Fe_2O_3	10.77%	plagioclase	33%
FeS_2	0.55%	hypersthene	20%
Al_2O_3	7.85%	ilmenite	39%
MgO	7.08%	magnetite	6%
CaO	3.26%	biotite	1%
MnO	0.17%	accessories	1%

The ilmenite contained about 13 weight per cent of exsolution lamellae of hematite. The magnetite concentrate of 65 per cent Fe contained 0.60 per cent V_2O_3 and the ilmenite concentrate 0.2 per cent V_2O_3.

The length of the Storganen dike is about 2 km. The ore is composed of alternating layers and lenses of ilmenite norite, leuconorite, and anorthosite, each layer or lens is from 0.5 to 10 m wide and from 200 to 400 m long. It was impossible to mine the ore selectively, but fortunately the grade over the entire width of the ore was sufficiently high to permit economic recovery of the entire ore body. The footwall contact with the anorthosite is sharp, but the hanging-wall contact was gradational. Where the western part of the ore body is in the contact plane of the anorthosite (east) and the foliated norite (west),it is parallel to the banding of the norite and, following that banding, the ore passes gradually into the norite. In the anorthosite, the ore body appears to have intruded the host rock and to have followed a prominent shear zone that had been developed in the solidified (or nearly solidified) anorthosite. Branches from the veins of ore are not uncommon, and where the hanging-wall rock is brecciated, the breccia is crossed by networks of small viens of ore.

The Storgangen dike system provides a connective link between the norite dikes internal to the Ana-Sira massif and the norites external to it along its northeast margin. These norites external to the Ana-Sira massif are located in the southern arm of the Bjerkrem-Sokndal lopolith. The structure of the main Storgangen ore body shows direct relationships with the rock and ore layers of the external norite bodies. The Storgangen ore body, in its lower reaches, is rather poorly structured massive ore and ilmenite-norite; this arrangement changes gradually upward to regularly bedded layers of norite, leuconorite, and anorthosite, whether the sequence is richly or poorly mineralized. The titanomagnetite part of the ore ranges from 0 to 30 per cent; the titanium in such ore appears to be present as ulvöspinel (or ulvite) (Fe_2TiO_4)- a spinel that is quite readily soluble in magnetite at high temperatures and that separates as exsolution bodies at lower ones. Krause and Pape (1975) report that at its western end, the dike material exhibits a definite differentiation that is marked by a continuous increase in magnetite content and in the $FeSiO_3$ molecule in the pyroxenes and by a corresponding decrease in the An content of the plagioclases.

The Tellnes deposit is contained in a large intrusion of ilmenite-norite in the Ana-Sira massif; the intrusion has the form of an arc of a circle with the convex direction facing northeast. The length of this arc is nearly 7 km. The thin western end of this intrusive arc cuts through the center of the mass of norite-pegmatite that contains the Blåfjell ore body at its northern end (see below). The surface area of the Tellnes ore body is about 570,000 m^2; 200,000,000 tons were proved by the diamond drilling and another 100,000,000 is inferred by a projection down dip. Of the 7 km arc-length of the intrusive, 2.7 km constitutes the ore body. The width of the ore body is up to 450 m. The intrusive character of the ore is shown by its texture, by apophyses of ilmenite-norite that cut into the surrounding rocks, by the xenoliths of anorthosite contained in the ilmenite-norite, and by the eruptive breccia produced by the act of intrusion in the wall rocks. The presence of bent and deformed crystals of plagioclase demonstrate that the magma was moving during the solidification process. The main and central part of the ore body is quite homogeneous but the margins against the anorthosite are more leucocratic.

Average chemical and mineralogical analyses of the Tellnes ore follow:

SiO_2	30.37%	P_2O_5	0.20%
TiO_2	18.40%	K_2O	0.60%
FeO	17.43%	Na_2O	2.40%
Fe_2O_3	7.25%	Plagioclase	36%
FeS	0.60%	Hypersthene	15%
Al_2O_3	11.70%	Ilmenite	39%
MgO	6.13%	Magnetite	2%
CaO	4.39%	Biotite	3.5%
MnO	0.18%	Accessories	3.5%

The ilmenite of the Tellnes ore contains 12 per cent hematite in exsolution lamellae.

Southeast of Egersund, in a narrow zone, more than 75 small prospects and mines (of which the Anker and Koldal were the most important) are known. The surface area of the Anker is 400 m^2 and of the Koldal 250 m^2; these ores appear to have been developed at, or, near, the boundary between the Egersund-Ogna massif to the north and the Håland and Helleren massif to the south where the rock containing them appears to be norite injected along this boundary of leuconorite and concentrated there during the crystallization of the younger (the Håland and Helleren) anorthosite or injected there from depth along this available line of weakness.

In the vicinity of the old, and now exhausted, Blåfjell mine the anorthosites range from medium-grained rocks, some of which are porphyritic due to the presence of huge crystals of andesine, to rocks made up of aggregates of large andesine crystals. These anorthosites grade into hypersthene-bearing rocks with the hypersthene crystals being up to 40 cm long. The coarse-grained norite pegmatite is composed of andesine feldspar, hypersthene, and ilmenite. Near, and presumably in, the Blåfjell mine, the pegmatite cuts and encloses major xenoliths of anorthosite derived from the host rocks of the pegmatite intrusion. This pegmatite can be traced southward from the Blåfjell mine for several kilometers in which small ore prospects have been found. Pegmatites of the same type are present in the vicinity of the ilmenite deposits in the Koldal zone. Bugge (1978) considers [and my observations agree with his] that the pegmatite and the ores it contains are the late stage melts developed during the differentiation of the leuconorite magma, thereby producing the anorthosites, leuconorites, and finally ilmenite-rich pegmatites.

The Blåfjell ores are present in the pegmatite mass as lenses or schlieren of high-grade ilmenite ore; they exhibit quite a range of sizes and seem to form an interlacing network in the silicic minerals that make up pagmatite, and these lenses and schlieren extend even into the anorthosite into which they were, of course, intruded.

An analysis of the Bjåfjell ore dating from 1945, was:

SiO_2	0.50%	FeO	32.69%
TiO_2	44.16%	MnO	0.26%
Al_2O_3	2.00%	MgO	5.28%
Fe_2O_3	14.34%	CaO	-----
V_2O_3	0.30%	S	0.00%

The composition of ilmenite from this source was (from a total of 1):

$FeTiO_3$ 0.80; $MgTiO_3$ 0.19; $MnTiO_3$ 0.01; Fe_2O_3 0.14.

In addition to the pegmatitic Bjåfjell ore, ilmenite and magnetite locally are concentrated in certain horizons of the banded norite. The largest of these no-more-than low-grade deposits will average 5 to 10 per cent TiO_2 and are, at least for the present, uneconomic. The deposits of Bakke-Ørsland [in the banded (foliated) norite] are about 2 km northeast of the Storgangen mine, while the Bø deposit is about the same distance southwest of Storgangen. The Bakke-Ørsland deposits are in a zone that can be traced for about 2 km (NE-SW) and is 2 to 10 m thick. Almost all of these minor iron-titanium deposits outside of the norite bodies internal to the Ana-Sira massif must be genetically connected with the norite pegmatites into bodies of which rock such dikes locally terminate.

Although the bulk of the mangerite (or mangeric-norite) is concentrated around the western and southern margins of the south prong of the Bjerkrem-Sokndal lopolith, much larger volumes of quartz-mangerite form the core of that lopolith and largely surround the north and east margins of the Ana-Sira massif. These mangerites (Krause and Pedall, 1980) are a complex association of rock types, the general modal composition of which causes them to be called the mangerite series. They are intruded rocks that make up the final stage of the magmatic-tectonic evolution of the South Rogaland igneous complex. They show gradual transformations to the underlying norite rocks external to the anorthosites and conformable with them, this suggesting they are comagmatic with the rest of the Rogaland complex series. These rocks are divided (Krause and Pedall, 1980) into rocks external to the anorthosites and those internal to them. The rocks of both these types are connected by a network of dikes. These authors consider that the Tellnes ore body, the major representative of the internal mangerites, shows a more mafic stage of differentiation than do the external mangerites in the Bjerkhem-Sokndal lopolith. These various types of mangerites were introduced into the area through a complex system of major and minor fractures and fissures that were produced through the final stage of the uplifting of the Ana-Sira antiform. This stage of uplift resulted in the development of fractures only, plastic flow was absent. The major fracture containing mangerite was that formed on the crest of the antiform that marks the eastern margin of the summit of that structure and provided the locus for the emplacement of the ilmenite-norite deposit of Tellnes that is the prime example of mangerite of the type internal to the Ana-Sira massif. This crest fracture is directly connected with a system of dikes that are arranged both radially and concentrically to the dome of the massif. This internal dike system thus is, further, connected in all directions with the structures containing the external mangerites (Krause and Pedall, 1980).

The mangerites of the Rogaland complex have a complicated mineralogy. The main constituents are plagioclase (An 20 to 50% and averaging 27%) and hypersthene as the orthopyroxene with quartz increasing the more youthful the mangerite in relation to other rocks of that species. Potash-feldspar and perthitic feldspar intermediate in composition between perthite and anitperthite, (that is mesoperthite), K-feldspar and albite (possibly oligoclase) in about equal proportions. Even in the most silicic varieties of this rock, plagioclase is markedly dominant over K-feldspar. Amphiboles, muscovite, and biotite are present only in minor amounts or are entirely absent even though the rock is rich in ilmenite. Most of the mangerites, however, have ilmenite and magnetite present as no more than minor constituents. Graphite is a common accessory mineral in the mangerites except where biotite occurs; in such cases, graphite is lacking.

The various rocks of the mangerite sequence are ilmenite-norite (as a special type), mangeronorite, quartz-mangerite, quartz-pegmatite, and pure quartz veins. All of these rock types may occur in one rock volume and there grade gradually into each other; more normally, however, the sequence is arranged concentrically within the anorthosite of the Ana-Sira massif. The crest fracture (Tellnes) and the broader dikes or sections of dikes contain the more mafic varieties, whereas the more silicic differentiates are in narrower dike sections with the quartz pegmatites and veins in even narrower fractures.

The study of the Tellnes deposit has shown that the ore body probably narrows down dip but no drilling has penetrated beyond the bottom of the ore. On the surface, toward the southeast, the ore body breaks down more and more into dike zones and offshoots that include many anorthosite xenoliths. At depth, in this area, however, the ore becomes more compact with a thickness of several hundred meters having been determined by drilling. To the northwest the ilmenite-norite ore gradually thins out and is converted to a mangerite dike, the strike of which is west-northwest.

In the ore-bearing part of the Tellnes mass, ilmenite, magnetite, and sulfides become less abundant toward the margins and the Cr_2O_3 content of the ore does likewise. Where apatite is present, it is most common in the outer 10 m of the ore body. Magnetite and sulfide lessen with depth but apatite and the Cr_2O_3 content increase.

Krause and Pedall (1980) suggest that the anorthositic, noritic, and mangeritic rock series, in all their subunits, definitely belong to the intrusive group. The original intrusive material in the area was introduced into the deep katazone, material that is now migmatitic charnockite gneiss. It was synkinematic in nature and was developed during the formation of a fold system that dipped to the east and must the present nonlinear north-south axes date from that time. This fold system seems to fit with that of the main folding of south Norway.

These authors believe that the main mechanism for the formation of anorthosite was the result of the gravitative separation of the primary andesine crystals and mafic orthocumlates. Probably due to differences in relative density and probably aided by convection currents, the plagioclase crystals rose and/or the mafic minerals sank. If the remains of this magma are assumed to have been covered by a layer of anorthosite, the accumulated mafic minerals below may have been remelted so that this magma became more mafic (noritic). This concept seems to be confirmed by finding xenoliths of olivine-magnetite-norite in the norites internal to the anorthosite masses now known at the surface. This idea of non-exposed basal units also draws confirmation from the presence of labradorite and bronzite inclusions and adcumulates and even more mafic xenoliths in the anorthosites of the Rogaland complex.

Krause and Pedall speculate (1980) that continuous east-west shortening resulted from the principal folding; thereby, the magma chamber remaining was narrowed. This caused, in the largely consolidated anorthosite series that overlay the magma chamber, the formation of the transverse fissures that are characteristic of the anorthosite, of which the Strogangen fissures are the quintessential example. At the same time, molten material from the upper parts of the internal magam was forced along the contact between the gneiss cover and the anorthosite since this contact was an already present zone of weakness. This melt was, in their opinion, concentrated principally in the synform that was developing between Sokndal and Bjerkreim and also along those mechanically appropriate points (as the result of east-west compression) on the northeast and southwest flanks of the Ana-Sira massif, thus producing the external examples of norite emplacement.

Through progressive crystallization differentiation, less mafic norite and norite-pegmatite magmas were produced as well as intercalated and conformable external examples of mangerite.

The intrusion of noritic-mangeritic melts above the lighter anorthosite caused isostatic adjustments with folding along north-south axes being discontinued in favor of vertical movements. These movements produced a further lowering of the Bjerkrem-Sokndal synform and a coincidental doming of the neighboring, east-dipping antiforms in both the Ana-Sira and the Håland Helleren massifs. At this stage, plastic deformation was superimposed on the structures developed during the orthomagmatic stage, subordinate folding was produced, the axes of which were arranged radially around the crests of the upfolds. As the upfolding of the Ana-Sira massif continued, three major crest fractures split open. In the oldest of the synkinematic intrusions, plastic deformation predominated and obliterated their contacts with the anorthosite. In the younger of these intrusions, such as the Blåfjell-Måkevatn (or Laksedal) norite pegmatite, however, the rigid framework in which they were enclosed caused them to be broken rather than to flow. Thus, the youngest and most easterly of the crest fractures, the one that now contains the ilmenite-norite Tellnes ore body and its associated dike system, were produced by the fracturing of a completely rigid rock framework.

The layered intrusion of Bøstølen, that was introduced to the west of, and before, the Blåfjell-Laksedal norite-pegmatite body, was a typical mafic intrusion that had a one-sided internal development that took place from its bottom to its top. Such a situation did not develop in the Blåfjell-Laksedal more than indistinctly and did not occur essentially at all in the ilmenite-norite body of Tellnes. The Tellnes body shows a direct differentiation relationship to the older norite bodies, but the various rock units of the mangerite series (mangerite, quartz-mangerite, and quartz-pegmatite) show direct evidence of having been differentiated in place. The internal mangerite sequence, therefore, occupies a cupola-shaped volume within the Ana-Sira antiform and shows a definite dependence on the width of the dikes it occupies. Mafic ilmenite-noritic to mangero-noritic material occupies the crest fractures and wide places in the dikes, whereas the narrow parts of the dikes contain more silicic differentiates and quartz pegmatites were emplaced in off-shoots and the branches that accompany them.

The formation of the external mangeritic bodies marks the end of the magmatic-tectonic development of the South Rogaland complex. The vertical movement after this stage, tore the Ana-Sira massif from the rocks that had surrounded it. Into the opening so created (essentially with the form of a ring-dike, the external mangeritic-quartz-mangeritic material was introduced.

The scenario, as developed by Krause and Pedall (1980), is far more satisfactory than that of the Michots, singly or together (Michot and Michot, 1969) and comes as nearly to being an outline of what must have taken place as it is possible to produce at this time.

It is apparent from this summary of Krause and Pedall's concept that the ilmenite-norite ores of the Rogaland complex were products of late-magmatic reactions. The deposit at Bjåfjell definitely is a norite pegmatite in which one of the crystallization products was ilmenite. In some instances, the ilmenite appears to have been crystallized from the melt along with the silicic minerals, in others the ilmenite (and its associated magnetite) may actually have become immiscible in the silicic magma, so that pods of ilmenite-magnetite separated from the norite magma and finished their crystallization as distinct entities. It is quite possible that some of the high-titanium pods in the Storgangen structure were separated from the main magma mass in the same manner, but, if this is so, appreciable amounts of silicate material remained in solution in the dominant ilmenite. In the Tellnes deposit, the ore is quite fine-grained and the rock appears to have crystallized rapidly; the ilmenite is uniformly distributed throughout the ore mass. The magma from which the Tellnes ore crystallized probably was a still-miscible solution of ilmenite-magnetite in the typical minerals of norite. The rock, however, is fine-grained enough to be classed as a norite aplite, a norite-aplite rich enough in ilmenite to be of ore grade. The classification of the Tellnes ore as a norite aplite is based on my own observations in 1963.

Thus, the ores in the Rogaland complex can be categorized as magmatic-3a and magmatic 3-b.

SWEDEN

SWEDEN (GENERAL)

Precambrian Middle Paleozoic	Iron, Manganese, Base Metals as Sulfides Precious Metals	Magmatic, Hydrothermal, Volcanic Exhalative

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Notes

Sweden is composed largely of rocks of Precambrian age that make up part of the Fennoscandian shield. The oldest of these rocks are definitely Archean, so old that they were subjected to major orogenic activity about 2800 m.y. ago, not to mention previous tectonic stages of greater age and less well-defined effects. The Archean rocks are confined to the eastern and northern parts of the Baltic shield of which the Fennoscandian is the westernmost extension. In Sweden proper, Archean rocks are exposed only in a very small portion of the northern part of the country. The boundary between Archean and younger rocks is not well established in Sweden, although in Finland, conglomerates mark the unconformity between Archean below and Proterozoic rocks above quite clearly. In the Archean of northern Scandinavia, only a few ore deposits are now (1982) known, and none has been located in Sweden. The nearest Archean ores to Swedish territory are the Norwegian iron ore body of Bjørnevann, near Sydvarangar, immediately adjacent to the Soviet border.

In contrast to the Archean, the Proterozoic rocks of Sweden are well and highly mineralized. Most of these Proterozoic deposits are concentrated in three main areas: (1) the Bergslagen which area forms a rude three-quarter circle that centers around the town of Sala (59°55'N, 16°38'E). The deposits in this area are quartz-banded and skarn iron ores and largely massive sulfides and range in age from as much as 2600 m.y. ago to perhaps as young as 1700 m.y. How much of the character of these deposits was determined early in the middle Precambrian and how much later under the influence of various stages of igneous activity from 1800 to 1700 m.y. ago remains to be firmly settled. Although the quartz-banded iron ores probably were developed as banded sediments, how these were, in part at least, converted to skarn ores and when remains uncertain. The sulfide deposits, such as Falun, probably were produced during the Svecokarelian era of igneous activity, being more probably related to the synkinematic intrusions that to those of post-orogenic age.

The second major area containing Proterozoic mineralization is in the Skellefte district that extends more or less northwest from the vicinity of Luleå on the west coast of the Gulf of Bothnia for over 100 km with an extension to the southwest from the Rakkajaur mine to the Ravliden-Kristineberg area. The age of these deposits appears to be approximately that of the palingenic Revsund granite that has an age of about 1800 m.y.

The third major ore district is that of Kiruna-Gällivare that includes not only these two major iron-ore deposits but also the huge, albeit low-grade copper ore body of Aitik and the newly discovered greenstone-contained copper deposits of Viscaria in the immediate vicinity of the Kiruna ore body. Several less large, but still important, iron-ore deposits are located within the general Kiruna area; including: Luossovaara, Tuolluvaara, Rektor, Nukutusvaara, Lappamalmen, Henry, Svappavaara, Leveäniemi, and Mertainen. The age of these deposits appears to be late middle Precambrian, that is, between 1800 and 1700 m.y.

A fourth district, about which the literature still is too scant to provide information for a detailed discussion of any single deposit, is northwest of the Skellefte district and is designated as the Arjeplog province. The initial discovery here appears to have been of molybdenite bodies north of Arjeplog in 1967 to 1968. These ores are contained in granites that are related to the 1565 m.y. old Lind granite. Deposits at Haurok and Skarjaviken consist of fissure fillings and disseminations of molybdenite with pyrite and chalcopyrite in lesser amounts; the actual host rock is sericite-altered porphyry. Deposits at Björntjärn and Munka appear to have better economic possibilities; they are aplite or quartz veins in which molybdenite is disseminated (Frietsch, and others, 1979). Some scheelite-bearing skarns also have been encountered in the area.

In addition to the molybdenite deposits, several uranium ore bodies have been found near Arjeplog. Frietsch and his colleagues suggest that the area may contain 20,000 tons of uranium, half of which is in the Pleutajokk body. The remainder of the uranium is divided among more than 20 occurrences of a similar different nature in silicic volcanics that are older than 1900 m.y., and a few have been discovered in granitoides in the Jörn granites also about 1900 m.y. old. The minerals in these deposits are pitchblende, quartz, chlorite, and some calcite and fluorite as fissure fillings and disseminations. The host volcanics and granitoids have been metasomatically altered to produce albite and riebeckite. The pitchblende is between 1740 and 1850 m.y. old, indicating that the deposits were formed at the height of the Svecokarelian orogeny. Frietsch and his colleagues, however, think that the ore fluids were of metamorphic origin and were introduced through joints and fault zones.

In addition to these Arjeplog deposits, the area north of that town includes some low-grade, quartz-banded iron ores in silicic volcanics and sedimentary rocks. Frietsch and his colleagues believe that these iron ores were volcano-sedimentary in origin and amount to 100 million tons (of ore?)

A number of more or less isolated deposits of rather unusual type occur elsewhere in the Proterozoic of Sweden. One of these is the carbonatite of Alnö Island (the town of Alnö is at 62°20'N, 17°30'E); the carbonatite contains columbium, barite, apatite, and lime. Another uncommon occurrence is the Varuträsk pegmatite that was emplaced 22 km southeast of the Boliden mine in the Skellefte district; it has no genetic connection with the sulfide ores, despite the presence of small pegmatite masses in the Boliden ores. The probable age of the Varuträsk pegmatite is about 1800 m.y., and it probably was genetically connected with the Revsund granite. It was mined for lithium, cesium, rubidium, and feldspar. Still a third unduplicated deposit in Sweden is the iron-titanium-vanadium mineralization at Smålands Taberg which is located about 10 km southwest of the town of Jönköping at the southern end of Lake Vättern and is the most southerly of the major ore deposits of Sweden. Smålands Taberg and Alnö Island are the only two late Precambrian deposits discussed in the Swedish section of this Volume.

Still another variety of unusual deposit in the middle Precambrian of Sweden is the nickel ore body of Lainijaur of which a fuller description is given in this Volume. Although the Lainijaur deposit is in Boliden district, lying on the general strike of that district about half way between Rakkejaur and the deposits of the Adak dome, it is directly connected with an igneous body. In the instance, the intrusive is a rather unusual gabbro phacolith and shrinks downward into a dike from which wing-shaped sills of gabbro extend outward into the synclinal structure made up of well-bedded sediments that include dacitic tuffs. In the central portion of the phacolith and dike are rich disseminations of nickel-bearing pyrrhotite. The disseminated ore is about 15 per cent of the total volume of the ore-bearing portion of the ore-bearing gabbro, 70 per cent is Ni-bearing pyrrhotite, and 15 per cent is nickel-arsenic ore. The average grade of the ore, of which 125,000 tons had been proved (Grip, 1978), 0.2 per cent Cu, 0.2 per cent nickel, 0.02 per cent cobalt, 2.5 per cent sulfur, and 0.2 per cent arsenic. The Ni-pyrrhotite ore runs 50.2 per cent Fe, 0.4 per cent Cu, 2.6 per cent Ni, 0.03 per cent Co, 32 per cent S, and 0.1 per cent As. The arsenic ore is in dikes that cut the Ni-pyrrhotite ore bodies and radiate out from them. The arsenic ore contains 18 per cent Fe, 1.9 per cent Cu, 10.1 per cent Ni, 1.1 per cent Co, 16 per cent S, and 18.5 per cent As. It appears that the intrusion of both the gabbro and the ore occurred in several stages. The Ålgliden nickel deposit, slightly north of the trend of the Skellefte ores and contained in the major Jörn granite, is much like the Lainijaur deposit. A major gabbro dike, at least 3.5 km long and as much as 100 m wide, cut a granodiorite; the dike is a multiple intrusion that ranges in character from ultramafic to silicic. The gabbro strikes NE-SW and dips vertically; although the generally uniformly disseminated chalcopyrite and pentlandite-bearing pyrrhotite, plus intergrown magnetite-ilmenite, sphalerite, pyrite, bravoite, and millerite, a greater concentration of these minerals is located in a fold near the mid-point in the strike of the dike. The average grade of this ore is 0.2 to 0.5 per cent Cu and 0.06 to 0.25 per cent Ni but with locally much higher concentrations.

In addition to Lainijaur and Ålgliden, a definite nickel-bearing belt was found south of Skellefteå (town 64°45'N, 21°00'E), about 110 km SSW of Luleå. Frietsch and his colleagues (1979) report that the belt is at least 10 km long and includes three separate ore bodies (Lappvattnet, Brännorna, and Mjövattnet). It is different from the two ore bodies mentioned at the head of this paragraph in that it is associated with ultramafic rocks (peridotites and minor pyroxenites); these are present as elongated parallel bodies in migmatites and, locally, in graphite-bearing metasediments. The sulfides are located in the ultramafics and in the neighboring gneisses as disseminations, veinlets, and breccia cements. The ore minerals are the typical pyrrhotite, pentlandite, and chalcopyrite. Frietsch and his colleagues consider that these ores are older than the 1785 m.y. old Revsund granite that migmatized the host rocks of the ultramafics. The breccia zones appear to connect the three deposits and suggest that the ultramafics were introduced along fault lines. A somewhat similar deposit has been found at Kukasjärvi, north of Luleå.

The most recent discovery in the Kiruna district has been the copper property of Viscaria about 3.5 km northwest of Kiruna. The Kiruna greenstone complex, in which the Viscaria deposit is contained, makes up a series of splitic lavas with which are intercalated thick beds of sediments. This Kiruna greenstone strikes NE-SW and dips about 80°SE. The sediments include tuffs, graphite schists, cherts (albitites), and limestones; these rocks have been metamorphosed to the greenschist facies in which primary structures are well preserved.

The copper mineralization in some of these beds, mainly limestone and graphite schist, are present as rich impregations and, in the limestones, as distinct strata-bound bands. These bands may be bordered by 1 cm to 1 m of massive chalcopyrite ore. In the graphite schists, similarly distinct strata-bound bands are present as are lenses and fracture fillings. Magnetite is a common mineral, appearing as impregnations in the limestone, as thin layers in the tuff, and locally as distinct beds of that iron oxide several meters thick. Copper and magnetite maintain a positive correlation with each other, and pyrrhotite is a common sulfide mineral in all of the greenstone series; pyrite is rare, and sphalerite is minor (Forsell and Godin, 1980).

The reserves at Viscaria are estimated to be as much as 30 million tons at a grade of 1.1 per cent Cu, the ore being contained in three separate beds designated as A,B, and D; their respective lengths are 3.5, 1.8, and 1.0 km and their respective grades are 1.4 per cent, 1.0 per cent, and 1.0 per cent Cu. The beds, that stand nearly vertically, range in width from 5 to 25 m and have been followed (by drilling) to depths between 150 and 400 m (Forsell and Godin, 1980). Although these authors do not say so in their 1980 paper, it seems almost certain that they consider that the Viscaria copper ores were formed by volcano-sedimentary processes and are, therefore, syngenetic.

The amount of Phanerozoic platform cover in Sweden is small and is confined to islands in the Gulf of Bothnia, a few patches in south-central Sweden, and the southern tip of the country. The only ore mineralizations in these rocks is limited to uranium ores in upper Cambrian alum shale.

The Caledonides in Sweden contain a considerable number of valuable ore deposits and cover nearly all of the western edge of Scandinavia, the belt being about 1700 km long and as much as 250 km wide. Although most of the Caledonides of Scandinavia are in Norway, the eastern edges of the variously and complexly associated nappes are quite sharp. The rocks of the nappes, in both countries, are locally highly metamorphosed, have moved considerable distances in a generally eastwardly direction and overlie a narrow belt of authochthonous sedimentary rocks that range in age from latest Precambrian through Silurian. These authochthonous rocks rest unconformably on the Precambrian of the Fennoscanian subdivision of the Baltic shield. Within the nappes of the Caledonian orogenic belt, the thrust slices are made up, wholly or in part, of Precambrian rocks. Within Sweden, three main geological and structural units can be recognized. These are: (1) farthest to the east, a thin and poorly exposed zone of authochthonous late Precambria Cambrian, and Orodovician sediments. In an antiformal zone along the boundary between the two countries, similar (though highly deformed) rocks surroun Precambrian windows. (2) this authochothonous zone was overridden by parauthochthonous to allochthonous nappes containing Eocambrian and lower Paleozoic beds; these rocks have been considered as miogeosynclinal but this designation is not certain. (3) the major portion of the Caledonides in

Sweden, however, is a metamorphic allocthon in a complicated nappe structure in which the rocks range from greenschist to upper amphibolite facies, these rocks having been thrust over 100s of kilometers from the west-northwest. The Seve-Köli nappe, important as a host to ore bodies, contains mica schists, gneisses, and amphibolites. This nappe and those overlying it, are characterized by a penetrative schistosity and a multi-phase fold history (Grip, 1978). In the Seve-Köli nappe, the Köli is the uppermost of the two rock sequences; the metamorphic grade of the Köli is appreciably less than that of the Seve, but the Köli structures are equally complex. The rocks of the Köli are principally sedimentary - quartzites, phyllites, conglomerates, limestones, and graywackes. The included volcanics are mafic and quartz keratophyres. The major Caledonide strata-bound sulfide deposits are located in this Köli segment. Although the nappes that overlie the Seve-Köli are more greatly metamorphosed, they are of minor extent in Sweden, in contrast to the situation in Norway to the west. Ore deposits in the eastern (frontal) portion of the Caledonides are less well mineralized than those in the central part. In the frontal zone deposits, the grade of copper is negligible and that of zinc is almost so; lead, however, averages nearly 4.0 per cent and sulfur and iron are low (not much over 1 per cent). In the central Caledonides, gold runs 0.2 ppm; silver 44 ppm; copper 1.3 per cent, zinc 3.2 per cent, lead 0.3 per cent, sulfur 20 per cent, and iron 18 per cent (Grip, 1978). The simple mineral suite in the miogesynclinal (frontal) deposits, in which galena is the main mineral, is in stark contrast to the eugeosynclinal (frontal) deposits in which the sulfide suite is complex and much more abundant.

The most important of the miogeosynclinal deposits was Nasafjäll that was worked between 1637 and 1806 and produced about 500 tons of lead and 1000 kg of silver from 20,000 tons of ore. Grip says that the deposit originally contained 1,600,000 tons of ore that averaged about 30 ppm silver, 1.2 per cent zinc, 1.5 per cent lead, and 3.3 per cent pyrrhotite. Why, if 20,000 tons of ore could yield 500 tons of lead and 1000 kg of silver, more ore was not recovered is not clear unless one set or the other of Grip's figures is incorrect. The Nasafjäll deposit is located at the south end of the Nasafjäll window in Norrbotten province and is in a narrow zone of Eocambrian and Cambrian quartzites and shales. The quartz-sulfide veins are localized by late-east-west folding in sedimentary rocks with the sulfides being concentrated in joints related to the veins. The main minerals are pyrrhotite, sphalerite, and galena; chalcopyrite in minor and boulangerite and pyrite are sparsely and erratically distributed. The normal occurrence of each mineral separate from the others allowed the ores to be beneficiated by hand-cobbing; the silver was present almost exclusively in the galena.

The exposed late-Precambrian to lower Paleozoic rocks in the inner Caledonides in Sweden contain a major Paleozoic sulfide province that extends for about 1500 km from Kvaenangen in the northeast to the south coast of Norway. Although most of the Caledonide rocks are in Norway, the remainder in Sweden include several major deposits. These are of two types: (1) those such as Stekenjokk that are richer in zinc than lead and include appreciable copper (the grade at Stekenjokk is 1.5% Cu, 3.0% Zn, and 0.3% Pb); other deposits of similar character include: Ropen, Västra Storbäcksdalen, Allak, Tjåter, Rikarbäcken, Gaisar, and Remdalen. (2) those such as Laisvall in which lead is much higher than zinc (the grade at Laisvall is about 4% Pb and minor Zn in some 60 million tons of ore); other deposits of similar character are Maiva (close to Laisvall), the Löstrand and Bellviksberg deposits in the Dorotea area (about 200 km SSW from Laisvall), and Vassbo (about 325 km SW of Dorotea). The Stekenjokk and Laisvall deposits are described in some detail later in this volume. Many other deposits of both types have been prospected to a greater or lesser degree, and some actually may be mined.

Allak is the most northerly of the deposits of the Stekenjokk type, but the concentrations of sulfides are rich, but small. Other, apparently even less valuable deposits are known in the Allak area. Rikarbäcken (some 115 km NNE of Stekenjokk) on the north side of the Joefjället (mountain or dome) is a massive sheet of ore that is about 260 m long and dips 30°NE; the strike, therefore must be about NW-SE; a number of smaller ore sheets parallel the main ore body; the grade is: 44 ppm Ag, 0.8% Cu, 4.3% Zn, 1.1% Pb, and 35% S. The tonnage now known (1978) is between 100,000 and 200,000. The principal mineral is pyrite with smaller amounts of sphalerite, galena, and chalcopyrite and even lesser quantities of arsenopyrite. The ore sheets are contained in a belt of graphitic phyllites some 10 m in thickness; these are over- and underlain by gray phyllite. Wall-rock alteration consists of sericitization of the graphitic phyllites next to the ore.

The Västra Storbäcksdalen ore deposit is about the same size as Rikarbäcken but is on the south side of the Joefjället. The ore is in a massive sheet about 1 m thick that dips to the NNE at a low angle; the principal minerals are sphalerite, galena, chalcopyrite, and pyrite, and the ore sheet is surrounded by sericite schist; the grade is 49 ppm Ag, 0.8% Cu, 6.3% Zn, 1.1% Pb, and 16% S. The third of this trio is Tjäter; it is 8 km ESE of Västra Storbäcksdalen and is composed of a large group of small lenses that are enclosed in sericite schist. These schists have been isoclinally folded, thus forming a quite thick unit in which the average grade is 49 ppm Ag, 1.0% Cu, 4.8% Zn, 1.9% Pb, and 13% S.

Farther north than any of these deposits but Allak is Gaisar; the ore tonnage is at least 1,000,000 tons of a grade of: 0.8% Cu and 30% S. It is composed of three sheets that dip 20° to 30°WNW, and these sheets are 1 to 3 m wide. The ore is massive pyrite containing some chalcopyrite, but it grades laterally into a banded magnetite-chalcopyrite-pyrite ore. The ore sheets are contained in greenschist, it follows the sedimentary banding, and it has a distinct alteration zone around it in which carbonates are nearer the ore and epidote farther away. The Remdalen deposit is located about 20 km SSW of Gaisar (if Grip's map is to be believed instead of his text which says 10 km) and contains a massive body of 600,000 tons of pyrite and pyrrhotite; the grade is 1.5% Cu, 2.5% Zn, and 25% S. The body stands nearly vertically and is contained in greenschists and phyllites. In all Grip and Frietsch's map shows 22 named deposits in the Stekenjokk district of which that deposit and its associated Levi and Tjokkola deposits are far the largest.

The Laisvall deposit is the largest lead deposit in Europe and has two major counterpart areas - Dorotea and Idre. The total distance from Laisvall to Idre is well over 500 km.

In the Dorotea area, nappes of Ström quartzite overlie authochthonous and parauthochthonous Cambro-Silurian beds; these primarily were shales but basal arkoses lie between the shales and the Precambrian rocks. These quartzite nappes moved from a WNW-direction toward the ESE and covered about 30 km. In this nappe complex, the shale rocks were folded into narrow anticlines, the axes of which parallel the Caledonian trend and which are cut by faults and joints that are parallel to that trend. These major folds divided the quartzites of the nappes into blocks and within these blocks minor folding and imbrication were developed. Much mineralization took place in the quartzite blocks, but most of these instances are uneconomic. Two well-mineralized areas within the Dorotea district were found at Bellviksberg and Lövstrand where a major number of mineralized bodies, both large and small, are separated by barren material or volumes of sparse disseminations. The zone in which these mineralized volumes occur is no more than 5 km wide, trends NNE-SSW, and lies on the eastern border of the nappe complex.

Although Grip reports that the lead-zinc relationship differs from one deposit or one volume in a deposit to another, both in a given area and over the entire nappe region in the deposits of the Laisvall-type, lead essentially always is more abundant than zinc. The location of any volume of mineralization is connected with tectonically developed structures, such as breccias zones or more simple fractures but also was produced as homogeneous impregnations in the more porous quartzites where these existed. Of course, not all apparently favorable sites, so far as channelways for solution movement were concerned, actually were mineralized. The largest deposit in the Dorotea area is the Lövstrand deposit that contains more than 10 million tons of ore in which the average lead content is 2.4 per cent; of course, both higher and lower assays can be found. The quartz nappe in which the ores were emplaced is about 100 m thick. The galena is located, in part, in fractures in the quartzite but also occurs as very fine-grained disseminations that exist as diffuse patches or layers parallel to the bedding. These quartzites contain many belts of imbricate faults and dip 45°SE; folding took place around two perpendicular fold axes that trend northwest and northeast.

A result of structures of this type is that the richer parts of the ore bodies have been elongated in a northwest-southeast direction and have a northwest dip. The more folds there are in a given area, the more volumes of valuable mineralization will be contained in a given volume of rocks.

The Bellviksberg deposit is much like that at Lövstrand, but it is a few kilometers farther to the south. Although Bellviksberg contains only one million tons of ore, compared with the 10 million at Lövstrand, the ore is appreciably better, containing 21 ppm Ag, 0.3% Zn, and 5.0% Pb. In this area, there are two nappes separated in the vertical direction by a thin zone of black shale, with the nappes being thrust over shales of Cambro-Ordovician age; the total thickness of the shales, due to thrust-induced repetition is about 280 m. The lead mineralization is almost all contained in the lower nappe, which is appreciably more strongly deformed.

The area around these two major deposits includes numerous small deposits, some of which are of rather high grade. In these deposits also, galena is the main mineral, but more sphalerite (of a light color) is found locally in much greater quantity than in the two large deposits.

The Idre area, the main deposit in which is that of Vassbo, has rocks of the Caledonian thrust structures lying on a peneplaned surface of Precambrian rocks; these ancient rocks consist of silicic volcanics intruded by diabases and dip toward the west at low angles. Grip (1978), quoting Tegengren, gives the stratigraphic sequence, from top to bottom, as:

1. nappe of Eocambrian quartzite and sparagmite with minor remnants of black shale.
2. Thrust surface
3. Mid-Cambrian black shale with a 12 m limestone horizon
4. Conglomerate with black-shale fragments, (± 0.5 m)
5. Lower Cambrian quartz sandstone with layers of shale or sandstone (4.5 to 15 m)
6. Discordance
7. Calcareous sandstone (5 to 8 m)
8. Lower Cambrian shale (1 to 4 m)
9. Conglomerate and arkose (3 m)
10. Discordance
11. Precambrian peneplane

This sequence remains essentially the same throughout the district, although facies and thickness differ considerably from place to place. The sandstone reaches its maximum thickness in a northwest-trending zone that passes through the Vassbo area, thinning out in both directions away from the mineralized volume. Cross-bedding of the ore-sandstone suggests that it was moved into the area from the southwest; the already solidified calcerous sandstone beneath it provided a solid base for the ore-sandstone. The ore-sandstone probably was a deltaic deposit.

Almost all the galena in the Idre district is contained in the ore-sandstone, even though a little disseminated lead sulfide is locally present in the uppermost part of the underlying calcareous sandstone in the Vassbo deposit proper. The concentration of the galena appears to have been controlled by both tectonic and sedimentary features.

The Vassbo deposit was another of the many Scandinavian ore deposits found by tracing mineralized boulders, in this instance, a 500 kg boulder with much galena and sphalerite was the first clue. The drilling program in the area indicated by the boulder train was successful very quickly, and mining began at Vassbo in 1960. The original estimate of the ore contained was 3 million tons with a grade of 18 ppm Ag, 0.3% Zn, and 5.7% Pb. Over 260,000 tons were mined in 1973.

The Vassbo mineralization was controlled by two systems of joints that strike N80°W and N30°E; these directions appear to have been produced by stresses resulting from the eastward-directed thrusting. The net result was a Z-shaped pattern formed by the joint intersections. Grip is convinced that the upward-moving ore-bearing solutions followed the joint surfaces and penetrated outward into the sandstone to form galena impregnations. Nearest the joints, lead is more important in relation to zinc than farther out into the quartzite.

It should be noted that the deposits of the Stekenjokk type lie consistently farther to the west than those of the Laisvall type. The Stekenjokk deposits almost certainly have been affected by higher temperatures (as have similar deposits still farther west in Norway) than those of the Laisvall type. The lower-temperature character of the Laisvall type is suggested by the presence in most of them of appreciable amounts of fluorite and barite, typically associated with telethermal deposits all over the world. The ores of Laisvall and Vassbo contain 1 to 2 per cent barite and 0.2 to 0.4 per cent fluorite; considerably richer deposits of these two non-metallics, however, are found in the country surrounding these deposits. North of Lake Storumen (located about halfway between Laisvall and Dorotea), in particular, Eocambrian and lower Cambrian sandstones are present; in these proved reserves of 13 per cent fluorite have been reported.

Thus far, it seems certain that exploration has never been carried out so thoroughly and efficiently as in Sweden and Finland. Factors favoring this condition have been highly intelligent indigenous populations, effective scientific education since the middle ages (earlier in Sweden than in Finland), and the development and application of progressively more sophisticated techniques in such operations. The primary condition militating against the success of exploration in these countries has been the almost complete cover of glacial drift over a large portion of the land. Nevertheless, glacial drift has not been entirely disadvantageous to exploration; many years ago it was realized, in both countries, that mineralized glacial boulders mark well-defined paths that can be followed back with considerable success to the location(s) from which they came. Even this technique of boulder-tracking would not be of much use if the lay people of both countries did not recognize the meaning of the boulders they found and did not bring them to the attention of the various geological exploration organizations available to carry out a search for the point of origin of the boulders.

In Sweden and Finland, the number of highly competent and successful exploration geologists is so large that any list of the best of them undoubtedly would do injustice to at least as many geologists by the omission of their names as it would give credit to those named. No one, therefore, to my knowledge has attempted to compile such a list for either country. Nevertheless, any student of the ore geology of Sweden cannot but realize that one name that would be on any list made for that country would be Erland Grip. Certainly, this summary of the general ore geology of Sweden that precedes this paragraph could not be written without access to the published work of Grip in which much of what he has not been able to publish, for one reason or another, is summarized.

CENTRAL SWEDEN IRON ORES

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Precambrian	Magnetite, Hematite	Sedimentary-A3

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Notes

The Central Sweden iron-ore area (Bergslagen) is home of at least 131 iron- and sulfide-ore mines, plus another 20-odd mines for various non-metallic minerals. The area forms a rough circle the radius of which averages about 270 km. It is, however, somewhat longer (E-W) than it is wide (N-S), and it also appears probable that the eastern rim of the area is covered by the waters of the Gulf of Bothnia (north) and the Gulf of Finland (south). Thus, the complete area is somewhat more oval than circular and has a total area of over 200,000 km^2.

A cover of supracrustal rocks (both of sedimentary and volcanic origin) lies to considerable depths on the relatively unknown Archean beneath. The age of these post-Archean rocks is probably more than 2000m.y., and the Svecofennian cycle was begun with their deposition. Volcanic rocks are more abundant than sedimentary, and the former are composed of silicic and intermediate lavas, tuffs, and agglonerates; they contain locally effusive rocks that are now greenstones. These volcanics have been appreciably metamorphosed. The finest-grained silicic volcanics, where they are crystalline are known as "hälleflintas"; the coarser-grained varieties are designated as "leptites". These leptites range from potassium-rich to soda-rich (quartz keratophyres); intermediate types are not common.

Within the volcanic rocks are intercalations of calcareous beds that range from magnesium-poor to magnesium-rich. Where the carbonates were impure, calcium-magnesium skarn minerals were developed during the metamorphic phases.

The leptite group is some 2000 m thick with the ores being found at several different horizons within the leptite sequence; in the western part of the region (where the great bulk of the ores are located), each ore horizon normally is marked by a particular type of iron deposit. After the completion of the period of volcanism, sedimentation, and ore formation (?), the erosive period that followed produced a marked discordance with the later sedimentation that lies above it. This later sedimentation consisted mainly of graywackes and shales. Locally, in the Bergsland area (particularly Uppland and Södemanland in the southeast) volcanic rocks were extruded above the later sediments, and these also contain iron, manganese, and sulfide ores.

The Svecofennian orogeny strongly affected both the ore deposits and the country rocks of the Bergslagen province. The folding was so intense that the dips normally are quite steep and the sediments in many places have been converted to schists. Much of the metamorphism was associated with the intrusion of the late synorogenic granites (or granitoids) and of the palingenic (and younger) granites. The magnesia metasomatism that developed in connection with the synorogenic granites was more impressive, particularly as it affected the sulfide ores and the country rock around them, producing skarn minerals in the limestones and dolomites and in the carbonate constituents of the ores themselves. The present nonstrata-bound forms associated with many of the ore bodies may have been formed by deformation and remobilization or may have resulted from the reactions of magnesium-rich solutions developed during the crystallization of the granites.

Prospecting in the Bergslagen region has been essentially continuous since the middle ages, and the prospectors preferred the small, high-grade deposits rather than large, low-grade ones because of the primitive character of the methods of concentration then available. Thus, Grip gives a table that shows how improvements in mining and particularly in concentration techniques resulted in a steady lowering of the numbers of operating mines and progressive raising of the tonnages produced. This table is reproduced here:

Year	No. of Iron mines	Total Production of Iron ore, 1000 tons	No. of Sulfide Mines	Total Production of Sulfide Ore 1000 tons
1861	511	425	23	ca 130
1873	646	823	16	ca 90
1910	229	2037		
1911	242	2153	26	278
1960	68	5510	11	933
1965	57	6442	9	837
1970	27	8979	12	1003
1973	22	8802	9	1216

Still earlier, in the 14th century, iron production from Bergslagen was no more than 13,000 tons, rising to 27,000 tons by the beginning of the 17th. The increase in the next century brought production to 153,000 tons annually by 1800. At the beginning of the 19th century, iron ore production rose markedly again; from 1833 through 1910, it averaged 773,000 tons a year. Much of such increases in tonnage resulted from new techniques in open-pit and underground mining methods but even more was owed to advances that permitted the removal of larger proportions of impurities.

Most of the mines, both for iron and for sulfur, are concentrated in a belt along the west-central margin of the Bergslagen province. Nevertheless, to simplify and clarify the relationships and differences among the various types of iron ores present in the province, Grip (1978) has divided these iron ores into eight categories, as follows:

Type of Iron Ores	Ore Reserves in million tons for 35 working mines 1967	Mean Iron Content in per cent
(1) Fe-Mn ore in carbonate rock with more than 1% Mn (youngest)	(I) 252	37
(2) Fe ore in carbonate rock with less than 1% Mn		
(3) quartz-banded, Mn-silicate iron ore with high Mn content		
(4) quartz-banded iron ore	(II) 288	34
(5) skarn iron ore, quartz-rich	(III) 42	31
(6) skarn iron ore, quartz-poor		
(7) Apatite-bearing iron ore	(IV) 282	46
(8) Iron ore of Lomberg type (oldest)		

For further simplification, he condenses these eight groups into four. Then, as will be summarized later, he takes the 154 mines he recognizes as worth study in the district and places them in 21 districts. These he further consolidates into eight major groupings that he designates as: (1) Western Ludvika; (2) Eastern Ludvika; (3) Ljusnarsberg; (4) Nora-Stråssa-Riddarhyttan-Norberg belt; (5) Filipstad; (6) Falun-Garpenberg-Sala sulfide belt; (7) Northeastern district; (8) Southern district. Next he eliminates the 23 mines from which only non-metallics are mined, but still there are 131 mines to be considered. Of these, 35 can be removed from consideration because only sulfide ores are produced from them. Of the remaining 96 mines, 5 more can be disregarded as they are mined more for sulfide ore than for iron ore. Of the remaining 91, 2 are mined more for iron than sulfide ore but must be considered for that reason. Grip's class (7) also is not discussed here because the major representative of that class (Grängesberg) is taken up separately later in this Volume. This eliminates the first ore field listed above - Western Ludvika, even though that field includes four mines that are not of the apatite-bearing iron ore type. The Eastern Ludvika field, however, contains three mines that are entirely or dominantly sulfide ore mines, but there remain six mines of the quartz-banded, iron-ore type, two wholly or mainly of the skarn iron ore, quartz-rich type, and two that are entirely or largely of the Fe-Mn ore type with more than 1 per cent Mn. The Ljusnarsberg field includes an equal division between type (1) and type (9).

Major grouping (4) [not type of iron ore] includes five of the 21 districts and 29 mines, only one of which is more a sulfide mine than an iron ore mine [Håkansboda in the Stråssa-Stripa district]; the remaining 20 include five types of iron mines. Major grouping (5) [Filipstad] contains three sulfide mines out of 14; 10 of the remaining 11 mines are iron-ore mines, and one of the 11 produces some sulfide ore.

Of the sulfide mines, Falun and Garpenberg are discussed later in this Volume, so sulfide ores in the Bergslagen province are not discussed further in this Central Sweden Iron ores section.

Even groupings (6), (7), and (8) contain a few examples worth summarizing here. The mines selected for this purpose are those that are, or have been, major producers. The mines discussed below are:

Grouping	Mine or Mine Group	Type of Iron Ores	Approximate Total Production, 1925 thru 1974
1	Tuna-Hästberg Mines	(1,2,6)	4,505,000
2	Håksberg	(4)	12,375,000
2	Nyberg	(6)	2,585,000
3	Ställberg	(1)	5,538,000
3	Bästkärn	(1)	3,038,000
4	Pershyttan & Striberg	(4)	8,732,000
4	Stripa, Ingelsgruvan, etc.	(4)	10,494,000
4	Stråssa & Blanka	(4)	19,359,000
4	Ridderhyttan	(6)	10,977,000
4	Norberg	(4,6)	22,050,000
5	Persberg, Långban, etc.	(6,4)	8,214,000
6	Intrånget	(6)	4,711,000
7	Vintjärn	(6)	4,717,000
7	Dannemora	(1)	11,840,000
7	Ramhäll	(2,4)	3,203,000
8	Kantorp & Stav	(4,1)	5,588,000

The Tuna-Hästberg ore bodies are nearly 40 km NNE of Grängesberg and are made up of six or more layers of calcareous and skarn iron ore that is interbedded in leptite, limestone, and amphibolite. The zone containing these iron-ore layers is some 75 m thick, and it dips 30° to 40° to the SE; at depth the dip is appreciably flatter. By the time mining was stopped in 1968, the length of the ore layers was 1300 m. The manganese-rich ores, in which the Mn-content normally is 1 to 2%, may have as much as 8 to 9% Mn. This manganese, instead of being included in the iron oxides, is contained in knebelite, a manganoan fayalite [$(Fe^2,Mg,Mn)_2SiO_4$], a typical skarn olivine in manganese-rich iron deposits. The high-manganese concentrations are in the carbonate ores [iron-ore type (1)]. The Tuna-Hästberg ore field has been considerably faulted, but these have only a few meters of displacement. The diabase dikes in the area appear to have made use of these faults to enter ore-bearing rock volumes, (Grip, 1978). The ore types present are: (1) Fe-Mn ore in carbonate rock with more than 1 per cent Mn, (2) Fe-Mn ore in carbonate rock with less than 1 per cent Mn, and (6) skarn iron ore, quartz poor. The basic material in these layers appears to have been deposited by chemical sedimentation with each pair of layers having been precipitated at the same time but having settled at different rates (Magnusson, 1960, 1970). As for the skarn minerals, this same author believes that they were developed from the thermal metamorphism of leptite and carbonate bands to produce calcium-bearing silicate marbles. The manganese-rich skarn layers, however, resulted from a later magnesium-manganese metasomatism to produce such minerals as the knebelite mentioned above. Just where the Mg-Mg metasomatizing solutions were developed and how they entered the quartz-banded layers will be taken up in more detail later.

The Håksberg series of deposits is located a few km north of Ludvika and is composed of quartz-banded iron ore [iron-ore type (4)]. These quartz-banded layers have been deformed to a major degree and now make up partly folded and partly lenticular stocks that extend to great depths. The essentially isoclinal folding has produced the repetition of the ore in many more parallel bands than can be accounted for by the original sedimentation. The ore contains both magnetite and hematite with all degrees of transition between the one and the other, but magnetite is the more abundant of the two. The planes of the schistosity of the ores, developed during the folding, are marked by mica and hematite scales. Skarn minerals are present, both in bands or as scattered grains; the exact arrangement apparently depends on the original composition of the interbedded layers.

The Håksberg field is dominated by what is called the Central ore body that is located in a syncline with parallel limbs with its trough resting on the 275 m level. The ore is mainly magnetite with a dominantly quartz gangue in a host rock of mica schists. The parallel ore body on the footwall side is a plate-like structure that dips 60°ESE and reaches at least to a depth of 600 m. The mica schist is surrounded by leptite layers. The total area of ore in this field is 25,000 m^2, and the ore assays 34 per cent Fe and 0.09 per cent P.

The Håksberg belt is separated from the Nyberg belt by Hilläng and Stolberg belts (not summarized here) nor are any deposits of the same character as the Håksberg belt of quartz-banded iron ores found east of the Stolberg belt. Instead, a long, thick zone of carbonate rocks is encountered, and, east of the carbonate belt, are the Nyberg (and Kärrgruvan) skarn iron-ore deposits. The Nyberg ores are contained in horizons of carbonates and skarn; Grip (1978) thinks that these originally were a single ore-bearing layer that was, through folding arranged so it is repeated in a series of more or less shallow folds. The ore-containing carbonate rocks are generally bordered by soda-rich leptites. The Nyberg field (with the smallest total production of the 16 discussed here) is about 2 km long, strikes north-south, and dips 25° to 70°E. The ore bodies are mainly magnetite with included skarn schlieren; the ore bodies have the form of irregular lenses with many offshoots that cut into the skarn and carbonate layers. The total area of the ore is about 25,000 m^2.

The Ställberg deposits are located centrally in a zone west of the Ställdalen syncline and some 13 km SSW of Grängesberg. In this structure, graywackes and gray schists make up the upper horizons; these are underlain on the western side by a zone rich in agglomerates, slaggy greenstones, and amphibolites. Under the amphibolites are potassium-rich leptites and manganese-rich iron ores; these, in turn, are underlain by more leptites. Still further west are two massifs of synorogenic granite. The bounding feature of the east side of the syncline is a major fracture zone and beyond this are leptites of the same stage as the lower leptites in the west side of the syncline. These leptites were intruded by pegmatites and granites that derive from a large massif of a younger palingenic granite. Magnusson (1940) considers the iron ores to be contemporaneous with the volcanic activity from which the leptites derived. The concentration of the ore materials took place during the Svecofennian folding and the intrusion of the synorogenic granites. Magnusson believes that the iron in the Ställberg (and related) deposits resulted from remobilization of volcano-sedimentary primary iron ores. Post-orogenic granites develop dome-shaped intrustions that cut off and deformed the older tectonic structure, and a large massif of the younger (post-orogenic?) granite underlies the whole district (Magnusson, 1940). Diabase dikes are the youngest rocks of all in the area. The Ställberg iron ores are of type (1), although the district contains appreciable bodies of type (2) ores. The Ställberg deposit proper lies centrally in a zone west of the syncline, and the grade runs about 50 per cent Fe and 5 per cent Mn. The ore bodies are contained in a complex of leptites that include numerous bands of carbonate rocks. The width of the iron-bearing formation averages about 4 m; it is divided into two parts that are arranged in an en echelon pattern. The two fractions of the Ställberg deposit have been followed to a depth of 1050 m down a dip of 70°SE. The magnetite in the Ställberg ores has more than 2 per cent Mn but about half the manganese is in knebelite. Leptite dikes are common and separate the ore into numerous ore bodies of steep dip; they are consistently surrounded by skarn that is younger than the dikes.

Incidentally, sulfide mineralization is quite widespread in the rocks of the Ljusnarsberg ore grouping with the major development of the ore-type being the Kaveltorp-Ljusnarsberg area. Their character matches that of such deposits as Falun and Garpenberg in that they are stratiform and are preferentially associated with iron ore and limestone (Grip, 1978). The main sulfide minerals in these deposits are pyrrhotite, pyrite, chalcopyrite, sphalerite, and galena with the proportions of these minerals differing widely from one deposit to another. They were developed mainly in skarn masses of various mineral compositions. If the deposits are well mineralized with sulfides, the surrounding leptites are appreciably altered. Magnetite may be one of the ore minerals in these sulfide deposit, and it locally is replaced by pyrite.

The Bastkärn ore field is located in the same 14 km long zone on the western side of the Ställberg deposit with both being in the same stratigraphic horizon. Movement along the faults that now contain the granite and pegmatite dikes broke the ore horizons into en echelon lenses. Where the major faults cut the ore zones, soft ore has been formed. The Bastkärn field is about 5 km northeast of Ställberg. The Bastkärn ore has a maximum width of 50 m but it divided into layers of carbonate rock, ore, skarn, and leptites with the layering becoming less impressive with depth; thus, the ore bodies are not generally well defined and actually are simply the richer parts of ore-bearing skarn layers. Much, if not most, of the Bastkärn manganese is in knebelite and manganiferous cummingtonite (dannemorite). Most of the original carbonate has been converted into silicates of the skarn, so the actual carbonate content is low. The ore ranges between 32 and 51 per cent Fe and has 3.9 per cent Mn. The soft ore is west of a crush zone down which surface waters moved in major volumes; this ore has quite a wide range of manganese content. The total area or ore in the Bastkärn field is 11,800 m^2; the average grade of the 215,000 tons of ore mined in 1973 was: 42 per cent Fe, 0.006 per cent P, 0.25 per cent S, and 4.2 per cent Mn.

The Nora-Stråssa-Riddarhyttan-Norberg belt is about 100 km long and parallels the Grängesberg-Idkerberget belt (though the former is much longer than the latter) about 20 km to the southeast. The Ljusnarsberg ore grouping lies between the Grängesberg-Idkerberget belt and the Nora-Norberg belt. The Nora-Norberg belt has about one-third of the iron ore reserves of central Sweden (the adjacent apatite-bearing iron ores have about the same proportion). The Nora-Norberg belt curves irregularly in a NNE-northeast and is synclinal in form, a form better developed in the southern portion. The central zone of the belt is composed of schists and the younger leptites and is bordered by the older (stratigraphically lower) leptite beds; older granites were concordantly intruded into this structure. The iron ore deposits of this type (4) ore grouping in many places are accompanied by sulfides that locally are dominant over the iron. No large sulfide deposits free of iron are known; the small ones are essentially always strata-bound and may follow the same horizons as some of the iron ores.

The first ores of this belt to be considered here are those of the Pershyttan and Striberg fields; both are located near the town of Nora and some 55 km south of Grängesberg. Pershyttan field contains banded-quartz ores that are southeast of the carbonate zone that is in the central part of a 16 km long ore-bearing zone. The deposit has been strongly folded in a Z-shaped form around axes that plunge 45° at S55°E; the field is cut by two large faults. The Pershyttan ore is fine-grained and locally schistose; it is made up of quartz and hematite with various amounts of hematite. The intrusion of the younger granites has caused the beds to have been recrystallized and (Grip says) remobilized. Some of the ores are strikingly well banded with alternate layers of hematite (with some quartz) and red jaspilite; this banding, however, has been largely destroyed by regional metamorphism. Although nearly 150,000 tons of ore were mined in 1966, mining ceased in 1967. The grade of the last ore extracted was 35 per cent Fe, 0.02 per cent P, and 0.2 per cent S.

The Striberg field is located a few km northwest of the long carbonate zone mentioned in the Pershyttan description. Its ores are quartz-banded hematite, generally with small amounts of magnetite; in places, the hematite has been completely replaced by magnetite. The most important skarn minerals are garnet and epidote. The entire field has been strongly folded, the last movements of which have broken up the initially continuous layers into lenses, stocks, or ruler-shaped bodies. One of these ore bodies, the Åsboberg, consists of scaly, schistose hematite with some quartz and mica or chlorite that have been arranged into a tight fold around an axis that dips 45°ENE; this ore has been explored down to more than 600 m beneath the surface (ca 1975). The ore area is small - 500 m^2 - and becomes smaller with depth. The ore is closely bordered by leptites that are micaceous and schistose. The total area of the Striberg field is 15,000 m^2. Although 157,000 tons of ore were mined in 1966, mining was stopped in 1967; Grip gives no reason for this or for the discontinuation of mining at Pershyttan in the same year; the most logical conclusion must be that the ore was exhausted, but any one of several other possible reasons may be the correct one.

The Stripa and Stråssa mines are located 22 km N10°E and 28 km N27°E respectively from Nora (59°32'N, 15°02'E) and the rocks that contain them are in a well-developed syncline that is a northerly continuation of a more southerly syncline (the Nora-Viker). The youngest beds in the syncline are schists; beneath these are hälleflintas and dolomites that contain Mn-rich iron ores. Still lower, and on both sides of the syncline, are many layers of quartz-banded iron ore, some of which exhibit well-preserved banding. The Stripa mines are located at almost the exact center of the northwest side of the syncline. Quartz-banded ore occurs at two stratigraphic levels and provides the best preserved examples of such banding in the entire Bergslagen. The upper of these two ore bodies, Huvudmalmen, is up to 17 m thick; the second, the Parallelmalmen, is 50 m further down. Both ore bodies have been complicatedly and isoclinally folded, folds that plunge ENE at 25°. During this folding, the ores also were appreciably faulted, and later the whole complex was cut by major faults. The ores have been intruded by younger granites and pegmatites, especially in the eastern and deeper parts. The dominant ore type is a quartz-banded material in which the main ore mineral is hematite accompanied by magnetite porphyroblasts. The quartz is a jaspilite, and actinolite, pyroxene, and locally epidote are scattered through the ore; the hematite bands, in the main part of the ore, are 4 to 7 cm thick and the quartz bands range from 3 to 4 cm in thickness. In 1973, 108,000 tons of ore were mined with a grade of 51 per cent Fe, 0.007 per cent P, and 0.004 per cent S. The mine was shut down in 1975.

The Stråssa mine is 7 km northeast of Stripa and is at the northern end of Nora-Viker the syncline, and it includes one of the largest deposits of non-apatitic ore in central Sweden; the ore is rather low grade, averaging 35 per cent Fe and 0.013 per cent P. It is composed mainly of quartz-banded ore that locally grades into skarn ore of various degrees of intensity. The deposit contains two ore types: (1) magnetite and (2) hematite. The magnetite ores normally are homogenous, have only faint banding, except for local quartz bands; hematite is only very sparsely developed, but quartz, micas, amphibole, and diopside are present. The hematite ores, on the contrary, are both banded and heterogeneous and are accompanied by varied amounts of skarn minerals. Although magnetite is present ubiquitously, it is an alteration product of hematite. Of the two ore types, hematite ore contains quartz banding more generally than does the magnetite. Locally, and rather strangely, barite layers containing disseminated magnetite are interbedded with the normal ore. Chalcopyrite, pyrite, and fluorite occur as intercalations or fissure fillings. In the walls of the ore zones, alteration has produced cordierite gneiss on the footwall and recrystallized rhyolites (leptites) on the hanging wall. The ores in the Stråssa deposits occur in four stratigraphic horizons, and the ore type is somewhat different in each; these ore layers are separated by cordierite and sillimanite gneisses and by leptite layers. Folding took place in at least two stages; the earlier stage is located around nearly horizontal fold axes and the later is folded around steep axes to produce a most complex arrangement in plan and section. The syncline that contains the Stråssa ore field is located as a partial structure in the larger Nora-Viker-Ridderhyttan synclinorium; both the syncline and the synclinorium are overturned to the west. The axis of the Stråssa structure dips 55°SSE on the average, but range from 70° at the surface to appreciably flatter depth where several of the ore bodies unite into one. Grip reports that, since 1858, the Stråssa mine has produced 21,600,000 tons of ore.

The Blanka mine is 2 km SSE of Stråssa and is separated on the surface by a sizeable lake (Dammsjön); it lies at the same stratigraphic level as the larger mine. The Blanka ores are composed in part of coarsely crystalline magnetite that is intergrown with quartz, chlorite, and mica and in part of rich hematite ore that contains magnetite porphyroblasts. Both ore types contain relics of quartz banding; uranium ore (much younger than the iron) is in the gouge (sköl) formation of the mine area.

The Riddarhyttan ore field is made up of at least nine separate mines and covers an area of 2 km by 7 km and is enclosed in a synclinal structure that is surrounded by anticlines in which the rocks are cordierite-mica schist. In the ore field, the supracrustal rocks are potassium leptites;

some of these contain quartz phenocrysts and others are fine-grained. In one of the mine areas (Högsfors), older granites are present in small bodies that cut the ore. On the south, the ore field is cut off against younger granites. The fold axes and lineation normally have a southwest plunge. In the even-grained K-leptites, quartz-banded ore is present in association with skarn silicates and hematite layers, the one intercalated in the others. The stratigraphically highest ores are of the manganese-bearing skarn and calcareous dolomite varieties; they also are associated with the K-leptites. The large fraction of the ores in this field are aligned in two parallel zones that strike NE-SW. The more northeastern zone is enclosed in an anticline of cordierite-mica schists and the southwestern in K-leptites of the syncline. Several ore types are present in the ore field, and transitions from one to another occur throughout the field. Grip considers these transitions to be the result of changing conditions of sedimentation during what he thinks to have been the volcano-sedimentary environment of deposition. Both the NW and SE ore zones probably were each a single ore horizon and have been folded together in places and broken apart in others. In the NW zone, the ores are dominantly skarn and calcareous iron ores, the total area of which is about 12,000 m^2. Those of the SE zone are, in part, manganese-rich and, in part, quartz-banded iron ores. The total area of the SE zone is about 19000 m^2.

Of the nine Riddarhyttan ore field mines, three are on the northwestern zone; these are, from SW to NE, Källfall, Sjögruvan, and Persgruvan. Six are on the southeastern zone; these are, from SW to NE, Rödgruvan, Höjdgruvan. Myrbaaka, Korphyttan, Nya Bastnäs, and (a few km northeast of Bastnäs) Högfors.

The Norberg ore field is 10 km NW of Fagersta [(59°59'N, 15°49'E)] and is at the northeast end of the Viker-Stripa-Stråssa-Ridderhyttan-Norberg belt (about 100 km long). Continued along strike for about 20 km, the trend reaches the Garpenberg area (with which this iron ore belt has no apparent genetic connection). The major rock of the ore field is leptites; these are bordered on both sides by older granites, red to the southeast and gray to the northwest. The first folding of the leptites is gently arranged around horizontal fold axes, the second was more of less strongly draped around steep secondary axes. This second stage of tectonic activity produced intense minor folds, mainly in the center of the ore field. This stage was so late that it also affected the red granites. The youngest rocks in the ore field are diabase dikes; no younger granites or pegmatites are known.

In the southern part of the field, the fine-grained potassium leptites have been folded into a complex syncline. These leptites are intercalated by quartz-banded iron ores, carbonate rocks, and Mn-poor skarn iron ores. To the northeast, at Stripåsen, the syncline wedges out into a fold that plunges SSW. An irregular anticline of Na-leptites and cordierite-mica schists lies northwest of the syncline; these anticline-forming rocks are interbedded with Mn-poor skarn and calcareous iron ore. This zone can be followed to the northeast from Norberg (60°04'N, 5°34'E) as far as Stripåsen (a distance of about 8 km). To the northwest of the syncline, plagioclase leptites (meta-dacites and meta-andesites), with intercalated agglomerates, are present; this rock sequence decreases in age toward the northwest.

The deposits of the Norberg field, as reported to Grip by Back, occupy a number of parallel zones; the field contains all of the ore types of Central Sweden except apatite-bearing iron ores. The iron ore normally is present in definite horizons in which the ore-bearing portions may be separated by considerable distances of barren or low-grade rock. Grip considers the ore to have been of volcano-sedimentary origin (much of which was appreciably changed by later metamorphism). The major ore type is the quartz-banded variety in which jasplite is dominant. The locally occurring sulfide also are confined to specific horizons.

Beginning in the far southwest, and almost due west of Norberg, the most important ore zones number eight. The first of these is the Kolningberg-Klackberg-Bålsjön zone; it has a length of approximately 6 km and locally contains workable lengths of manganese-rich ore that is made up of fine-grained magnetite in a carbonate rock rich in manganese and iron; the silica content is low, Mn is 4 to 5 per cent, and iron is 45 per cent. The total area of the ore in the zone is about 19,000 m^2; the persistent carbonate bed can be followed throughout most of the length of the zone, and graphite occurs in parts of the deposit. Folding in the Klackberg field was intense and the ore horizon fragmented. The Kolningsberg deposit is divided into two major fractions by a strong fault that dips NNE whereas the old and new Kolningsberg fields were appreciably less strongly folded. The second zone, Hedberg-Kallmorberg, is the next to the southeast with an approximate length of 3 km and lies essentially parallel to the northeastern part of zone (1); this zone contains skarn with hematite ore, and the quartz-banded hematite and carbonate rock are beautifully banded; some of this ore type has been altered to massive skarn with magnetite. In the Kallmora part of this zone (in the southwestern part of the Kallmorberg field), there are concentrations of sulfides (PbS and ZnS), apparently economic for the silver included in the galena. The total area of zone (2) ore is 15,000 m^2, with 11,000 of that being in the Kallmorberg mine (at the extreme north of the zone) where the Fe content runs 40 to 45 per cent. The third zone is the Assessorskan-Kallmora zone, immediately adjacent to zone (2) to the southeast; it is traceable for 5 km, but most of the ore is in the Assessorskan area. The ore is hematite with rhodonite, yellow garnet, and yellow pyroxene in a zone of quartz-banded ores. The fourth zone is that of Gustav-Adolf; it is a zone of quartz-banded iron ore almost 3 km long and lies between zone (3) and zone (5) (the Hülpher-Bygruvan zone); the largest ore concentration is in the Gustav Adolf mine proper; this body is ruler-shaped and plunges 15°NNE; the iron content is 38 per cent, and the ore area is about 25,000 m^2. The fifth zone is the Hülpher-Bygruvan zone that is parallel to zone (4) and only a few meters southwest from it. It continues farther south than does zone (4) where the Hülpher-By, and Kärr mines make up the Morberg field and work quartz-banded iron ore. The sixth zone is nameless, so far as Grip's text of 1978 or Geijer's map of 1936 are concerned; it is southeast of the southern part of zone (5) and is another quartz-banded iron ore. Zone (7), also nameless, is manganese-silicate skarn in hematite ore and is located near the contact with granite and slightly north of zones (6) and (8). It includes no major concentrations of any type. The eighth zone - the Nygruvan-Bondgruvan-Miner zone - is slightly under 100 m east of zone (6) and is made up of strongly folded quartz-banded ore that has been intensely folded. In this ore, hematite is more common than magnetite, and the whole has the look of jaspilite with regular banding and sharp boundaries. The hematite bands are wider than those of quartz, and the banded ore alternates with layers of leptite. Even in the leptites, thin layers of hematite emphasize the bedding. Small layers of skarn and relics of carbonate rocks locally are present. In the southern part of this zone (Mimer and Ragvald fields) the ore horizons are minimally folded; this also is true of the Morberg (sic) field to the north. Between these two fields (Mimer-Ragvald and Morberg) folding becomes more intense and affects the neighboring red granite to an equal degree. The Risberg field, between the Ragvald field (south) and the Morberg field (north), is stongly folded and is molded around axes dipping about 65°SW. In Risberg, this folding has converted the original bedded ore into thick ore bodies oriented as are the fold axes. The total area of zone (8) is more than 70,000 m^2; the ore averages about 38 per cent Fe. 57,000 of the 70,000 m^2 lies in the strongly folded Risberg and Norrberg (sic) fields; the latter is about 700 m NW of the Risberg. Sulfides are disseminated in the Risberg ore, 300 m west of Bondgruvan (on the contact between leptite and red granite);

these consist of chalcopyrite, chalcocite, and molybdenite, plus scheelite and fluorite. These sulfides appear to have been introduced at the same time as the magnesium metasomatism; similar, but weak copper mineralization is found throughout zone (8). To the southwest of the eight zones of the Norberg area, particularly southwest of the Risberg field, is an area in which skarn and skarn iron ore have been folded into a Z-shape; here are the Åsgruvan and Ostanmossan mines. In these mines, magnetite is present as grains or aggregates in the skarn all variations are found between magnetite-rich and magnetite-poor skarns in which relics of carbonate rock are included. The main skarn minerals are diopside and actinolite with garnet present where magnetite is absent. Against the boudaries with leptite, hornblende and epidote occur. Where dolomite is present, a second skarn type developed with cerium minerals being locally associated with humite minerals. Disseminated chalcopyrite, scheelite, and molybdenite are found in the Åsgruvan mine which has an area of 16,500 m^2 and 38 per cent Fe.

Southwest of this skarn zone is the Eskilbacken field with a length of more than 600 m; it contains four tabular, steep-dipping ore bodies that have a total area of 12,000 m^2 and contain 38 per cent Fe. Both the quartz-banded iron ore and its host rocks have been appreciably altered, the leptite now being cordierite-mica schists.

In 1973, only three mines were being operated; Eskilbacken and Mimer produced over 101,000 tons and Kallmorberg over 53,500 tons.

The Filipstad district contains a considerable number of mines of which only Persberg and Taberg are considered here: Långban is treated elsewhere in this volume. The Persberg mine is located 7.5 km in a N55°E direction from the town of Filipstad (59°44'N, 14°10'E) on a peninsula extending southwest into the northern end of Lake Yngen, the largest lake immediately northeast of Filipstad. East of Filipstad, a 35 km long belt of leptites reaches from Nordmark in the north (59°52'N, 14°05'E) to Kroppa to the south. Grip (1978) reports this to be the most westerly zone of leptites in Central Sweden; the leptite is surrounded by the Gothian Filipstad granite except for the eastern side where the granite is older Svecofennian (i.e. synorogenic) granite. The stratigraphy in the Filipstad district is reported to be, top to bottom:

gray and black schists
graywackes and spilitic greenstones
sills of greenstone
dolomites and limestones with manganiferous iron ores
potassium leptites
sodium leptites including many horizons of manganese-poor skarn iron ores and iron ores in carbonate rocks

The igneous rocks in the district are, top to bottom:

Gothian Värmland granite
Late Svecofennian palingenic granites
Synorigenic Svecofennian granites.

The characteristic structure of the area is gentle folding about horizontal axes; farther east the folding is appreciably more complex.

In the Persberg area, however, Magnusson (1925) described five strongly folded ore horizons; these are not continuous through the whole area, and their thicknesses differ appreciably from one place to another. Where the ore bodies are largest, the amount of carbonate rock is greatest. One of such ore bodies is that of the Persberg-Odal field, and it is the largest iron-ore deposit in the entire Filipstad area. The ore body is composed of carbonate rock, skarn, and iron ore and is about 400 m wide and nearly 1 km in length. Within this rock mass, two parallel synclinal folds have been deciphered; these are divided by a ridge of leptite. Within this rock volume, the ore bodies have quite irregular shapes and equally irregular are the skarn masses in which they are contained.

The Persberg-Odal area includes several different varieties of ore, and the major one is a skarn iron ore in which the iron mineral is magnetite and the skarn is garnet-bearing pyroxene skarn. Where the skarn approaches leptite, hornblende and epidote are important. Another type of ore is contained in carbonate rock on the western side of the Odal field; the amount of skarn in this ore is minor, and, therefore, locally the bedding is perfectly retained. On the eastern side of the field, in the Alabama mine, the iron ore is a gouge-rich, talc, anthophyllite type with relics of garnet-pyroxene skarn.

The total area of the Persberg-Odal field is more than 13,000 m^2, and the iron content is between 35 and 40 per cent, phosphorus is about 0.01 per cent and sulfur 0.005 per cent.

The Odal field is only one example of the ore types in Persberg, the four lowest of the five ore horizons, of which the Odal field is one, contain similar kinds of ore with the garnet-pyroxene skarn type being most important. The fifth (and uppermost) horizon is composed of skarn ore that is rich in quartz; this horizon occurs in part to the south of Persberg and in part in the zone that extends north to Långban.

The Taberg field is 1 km north of the Nordmark-Odal field and just over 15 km N10°W from Filipstad; there an ore-bearing skarn is tightly wrapped around an axis that dips about 30°NW. In the upper levels, several distinct ore bodies are known, but these join at depth to form a rather integraged ore body. The Taberg field had a total area of about 4000 m^2; mining there was stopped in 1962.

An iron-ore deposit south of Taberg (at Värmlandsberg) was discovered by airborne geophysical prospecting; later drilling found the ore body to be a flat, elongated lens, the downward plunge of which is 15°NNW. This ore body probably is at the same stratigraphic level as that of Taberg; it follows the contact zone between carbonate rock below and leptite above. The upper level of the ore body reaches no higher than the 280 m level, and the lower limit has been determined as the 420 m level. On the 390 m level, the area of the ore was 8700 m^2. In 1973, over 87,000 tons of ore were mined with a grade of 58 per cent Fe, 0.003 per cent P, 0.2 per cent S, and 0.2 per cent Mn.

In the Garpenberg portion of the Fala-Garpenberg-Sala belt, ore bodies were found that were part iron ore, part sulfide ore (as was true of the Ryllshyttan mine) only a few kilometers WSW of Garpenberg, another was the skarn iron-ore deposit of Intrånget, 3 km north of Ryllshyttan.

The Garpenberg area consists of a leptite belt, 15 km long in an NE-SW direction; it starts from the Dalälven (river), runs through Garpenberg, and ends some 10 km beyond that town. This leptite complex is made up of silicic and intermediate volcanic rocks with two important horizons of carbonate rock intercalated among the bedded leptite. Between these two carbonate beds is an amphibolitic greenstone. The structure of the belt is synclinal, and the synclinorium is surrounded on all sides by synorogenic Svecofennian granite. The leptites are overlain by arenites, and the entire rock complex has been deformed into isoclinal folds developed around horizontal axes. A secondary stage of folding was produced around steep axes that are parallel to the plunge of the ore bodies. The leptite complex includes several small, oval-shaped granite bodies that also were involved in the folding.

The horizons of iron ore, in contrast to the form of the sulfide bodies, can be traced for considerable distances; the former are skarn-carbonate ores, and these, in the southwest part of the area, are Mn-bearing. In this area, the manganese-bearing ores are in rocks low in the stratigraphic sequence in contrast to those in the Filipstad district where the ores are stratigraphically high. One exception to this rule is the Småltarmossen iron deposit that is appreciably higher than the other Garpenberg ores.

In the Ryllshyttan mine, the deposit contains mineable amounts of both iron and copper ores; this also is true of the Intrånget mine. Ryllshyttan is a short distance WSW of Garpenberg and is close to the granite massif that cuts off the leptite belt on the west. Intercalated with these leptites are numerous calcareous beds that have been converted largely to skarn. The limestones, the skarn, and the ore contained in them have been strongly and steeply folded. From the Ryllshyttan mine, 85 per cent of the 1 million tons of ore mined were of the sphalerite-rich sulfide variety; the remainder was iron ore. The ore area in this field was 1000 m^2.

The ore at Intrånget is located about 3 km north of Ryllshyttan and is a skarn deposit in a leptite xenolith; this xenolith is entirely enclosed by red granite gneiss and has been cut by bodies of diorite and gabbro. The leptite that contains the skarn ore, on the surface, has been folded into a right-angle structure that opens out at depth, where the thickness of the leptite decreases. The largest area of ore is 4150 m^2 on the 245-m level. The ore includes quartz-banded material as well as skarn ore and most of both varieties is enclosed in limestone. The content of pyrrhotite, pyrite, and chalcopyrite as disseminated grains is high; certain small parts of the iron ores were so rich in sulfides that they could be designated as copper ore. 1968 was the last year of mining when nearly 80,000 tons of ore were produced; it averaged 25.2 per cent Fe, 0.023 per cent P, and 0.56 per cent S.

In the Northeastern district (surely better designated as the Northern district), the principal mine is at Vintjärn, about 65 km just west of north from Garpenberg (and with Kölen) the northernmost iron mine in Central Sweden; Vintjärn is about the same distance, slightly north of west from the port of Gävle and is some 18 km NE of the Svädsjö copper mine. The ore at Vintjärn is of the skarn and calcareous-iron types and is contained in a narrow belt of metamorphosed leptite in which the carbonate rocks are interlayered. This belt is bounded on both north and south by large masses of synorogenic Svecofennian granite; all the older rocks were intruded by pegmatites and fine-grained granites related to them. The actual ore is contained in two parallel layers that strike E-W and dip 30° S; these layers are gently folded about axes that dip 30°WSW. The footwall layer, although it is made up of magnetite- and skarn-bearing limestones, normally is too low-grade to be mined; the hanging-wall, however, is mineable along its entire length of 1500 m. Each layer is about 6 m thick. The plunge of the ore layers is parallel to the dip of the fold axes. The ore is strata-bound but is broken up by faults with displacements of as much as 10 m; meta-diabase dikes are present in at least some of these faults.

In 1973, just over 300,000 tons of ore, 30 to 35 per cent Fe, were mined from this field.

In the Uppland area, northeast of Uppsala, an irregularly shaped mass of leptite (with a generally NNE strike) contains the major iron-ore deposit of Dannemora. This leptite is surrounded completely by synorogenic Svecofennian granite. More or less east-west wings extend outward from this general NNE trend, and these were developed by an early folding that was later distorted by the NNE folding. The potassic, sodic, and intermediate leptites contain intercalations of other rocks, hälleflintas, dacites, metadacites, and extrusive greenstones as well as thick limestones; with these last are associated several horizons of iron ore.

The largest deposit of iron ore is that at Dannemora but the deposits at Ramhäll and Herräng are well worth mining. Both the iron ores and the minor sulfide deposits in the area are strata-bound and are confined to calcareous rocks, although some of the sulfide deposits are thought by Grip to have been somewhat displaced during the intrusion of the granite.

The Dannemora mines appear to be about 40 km NNE of Uppsala and are contained in the limestones of a series of slightly altered volcano-sedimentary layers of hälleflintas and limestones. The mines are spaced along a length of about 3 km; the greatest depth of mineralization is 1150 m at the northern end of the line. The dip of the beds at the surface is 80° to 90°WNW, but the beds flatten out considerably with depth. The ore-bearing group of rocks is confined to a tight syncline. Most of the ore is in the west limb of the syncline; the limbs diverge appreciably with depth. The hälleflintas form the footwall of the ore-bearing complex; farther to the north, the hälleflintas are coarser-grained and are essentially leptites. The zone of hälleflintas and leptites is surrounded on both sides by synorogenic granites; to the south, dikes of granite are present and become more abundant with depth; dikes of felsite porphyry and greenstone are older than the granite but younger than the ore-bearing complex. Faulting has broken up the ore horizon into several separate ore bodies; the major examples of these faults are quite flat and appear to have been thrusts; the hanging-wall blocks appear to have moved several 100 meters to the south. Later faults are much steeper and strike about N-S.

The ore at Dannemora is finely crystalline material, which normally is enclosed in skarn, but may be in direct contact with limestone; skarn minerals also are present in the ore. Dannemorite $[Mn_2(Fe^2,Mg)_5Si_8O_{22}(OH_2)]$ is the major mineral in the skarn; knebelite, pyroxene, actinolite, garnet, and serpentine also were developed.

The area of the Dannemora ore is about 12,000 m^2; in 1973, 580,000 tons of ore were mined; these had to be separated magnetically from the mixture of ore and rock recovered from the mine. The element content of the concentrated ore was 48 per cent Fe, 0.003 per cent P, and 3 per cent of both Mn and S. In 1973, another 310,000 of ore of a lower-grade were mined.

Sulfide minerals are present throughout the ore bodies, but small ore bodies of sulfide ore are found in the southern part of the field; the ore minerals are pyrrhotite, pyrite, sphalerite, galena, and arsenopyrite. The last of the sulfides ores was mined in 1920.

The Ramhäll field is located some 13 km SSE of Dannemora and is contained in a narrow band of leptite that strikes E-W and cuts across the Dannemora leptite zone. The principal ore body is a manganese-bearing limestone-banded magnetite material that is 400 m long and is as much as 4 m thick; it has been followed to a depth of 580 m. This ore field follows the strike (E-W) of the leptite body, has a vertical dip, and plunges east at 40°. Around the main ore body are a few smaller and parallel bodies; the entire ore group is contained in a somewhat dolomitic manganese-bearing limestone. This limestone is contained within a mass of sodic leptites. A zone of quartz-banded hematite ore is parallel to the main ore body, a short distance north of it, the host rock of which is a potassium leptite. The total area of the Ramhäll ore bodies is 3000 m^2. Nearly 150,000 tons of magnetically concentrated ore were recovered in 1973 from a total tonnage mined of just over 200,000; the grade of the concentrate was 30 per cent Fe, 0.02 per cent P, 0.3 per cent S, and 1.5 per cent Mn.

In the Southern district, the two major mines are Kantorp and Stav; these are located 10 and 12 km southwest and WSW, respectively, of the town of Flen (59°04'N, 16°39'E) - 45 km NW of the major city of Nyköping. These mines are centrally located in a WNW-ESE-striking ore zone. Both quartz-banded and skarn ores are present, side by each, in both limestones and dolomites; some Mn-bearing ore is known from the Stav mine. Deformation in the ore zone has consisted of gentle folding around WNW axes and cross folding around NNE axes. The ore zone is intercalated in leptites and limestones and has been intruded by the ubiquitous synorogenic Svecofennian granite. This folded complex was further affected by the formation of veined gneisses; this formation of the gneiss disturbed the first-stage fold pattern and, at the same time, the ore bodies developed during that first-stage folding were buckled and even fragmented. Although the Kantorp ore body was strongly deformed during the gneiss formation, a nearby ore body (the Wilhelmina) has the form of a highly elongated strip that has been double folded, dips at 30°E, and did not exhibit any signs of fragmentation until a depth of 450 m was reached. The Kantorp ore originally appears to have been quartz-banded iron ore, but it was impregnated with feldspar and mica during the formation of the gneisses. During the same process, some parts of the original ore body were changed to skarn-pegmatite ore. The total area of Kantorp ore was slightly over 10,000 m^2, and the grade (after concentration?) was 45 to 50 per cent iron.

TheStav mine (3 km NW of Kantorp) contains both quartz-banded and skarn-type iron ore, both enclosed in limestone and dolomite. The ore was penetrated by pegmatite, as were the carbonate rocks; this intrustion altered some of the original ore minerals and produced some new ones. The ore body plunges to the northeast at 70°. The 170,000 tons of crude ore recovered from the Kantorp mine in 1966 contained 32 per cent Fe, 0.04 per cent P, and 0.01 per cent S; in the same year, nearly 50,000 tons of crude ore was produced at the Stav mine; this ore had a grade of 34 per cent Fe, 0.04 per cent P, and 0.01 per cent S. All mining was stopped in the area of the two mines in 1967.

Obviously, the mines, the geology of which, following Grip (1978), is summarized here, are far from all the 100 plus mines that are recorded by Grip in his legend to his Figure 3, but enough have been given, I hope, to have indicated the wide variety of ores and the appreciably different environments in which they were originally produced and later changed. Nevertheless, it is possible to develop a general picture of the stages in which the various ore types were formed and something of the reasons for their being so consistently located in leptites in close proximity to Svecofennian granites.

No more than a glance at the map of the Bergslagen region in Grip's (1978) paper shows that the bulk of the iron ores are concentrated around the western margin of the circular area of Svecofennian granites. This appears reasonably to follow because by far the larger areas of leptites (and hälleflintas) are to be found there and because the original Bergslagen iron ores and leptites invariably are associated in time and space. Presumably, the entire Bergslagen region once consisted of early middle Precambrian supracrustal rocks - volcanics and sediments. The volcanic-rock suite, composed leptites and hälleflintas in the main, contains potassium-rich and sodium-rich varieties (types intermediate between the two are minor). The leptites and the hälleflintas are essentially rhyolites, and both commonly contain intercalations of limestone and dolomite as layers or lenses. Thermal and hydrothermal metamorphism have converted these carbonate-silica layers, depending on the ratios of silica to carbonate, to skarn-type rocks or to quartz-banded iron ores or any variation between the two. The skarn silicates are the result of reactions between the carbonate and silica (and any impurities originally in the ore lenses or added from later hydrothermal fluids). The quartz (originally chert) required for the formation of these skarn silicates probably was precipitated largely or considerably as chert at essentially the same time as the carbonate sediments, but some (perhaps much) silica was brought in later by hydrothermal fluids.

It was the intrusion of the synorogenic granites into the Bergslagen area that eliminated much of the volcano-sedimentary rocks (mainly leptites), and the granitic magmas that assimilated the pre-Svecofennian sediments probably produced, through differentiation, the hydrothermal fluids that were responsible for much of the metamorphism of the leptites and of the various types of lenses of sediments that they contained. Since this assimilation affected the leptites in the west of the Bergslagen area least intensely, more leptite was left there than in the other parts of the area. The concentration of the iron ores in these leptites indicates that at least much, if not most, of the iron ores were introduced into the supracrustal system at the time when the leptites were being formed. The ore in the leptites definitely is in chert-bearing carbonate and carbonate-bearing chert, a result obtained because the conditions obtaining when these sedimentary types were being formed was more favorable to ore deposition than when the leptites proper were developing.

It appears to be agreed among Swedish geologists that the iron was introduced during the volcano-sedimentary process that produced the leptites and their intercalations and not, as James (1955) suggested for the Mesabi-Range ores, by surface waters entering shallow arms of the sea. The main question remaining to be solved, apparently is: was all the iron now in the Bergslagen deposits introduced with the formation of the supracrustal rocks or was more iron added later?

Some appreciable time after these supracrustal rocks had been laid down in an essentially horizontal attitude, they were intensely folded, the folds being tight and, in many places, isoclinal. The tectonic forces responsible for the folding also developed more or less strong schistosity throughout much of the lepates, whereas, the higher the original proportion of carbonate sediments intercalated into the primary leptites, the greater would have been the skarn-mineral content of the metasomatically produced gangue minerals.

Of course, much (perhaps appreciably more than half) of the remnants of leptite containing quartz-banded iron ore probably would not have been reached or affected by the synorogenically generated hydrothermal fluids. Such ores, although not hydrothermally altered, would still have undergone major changes due to dynamic metamorphism alone such as that which affected the Bergslagen during the intrusion of the synorogenic granites.

The supracrustal rocks, of which the original leptites constituted the major fraction, were laid down in an essentially horizontal attitude. By the time they were attacked by hydrothermal fluids and affected by dynamometamorphism, this attitude was markedly changed. The folding that resulted from the tectonism was intense and, in many places definitely isoclinal. The forces responsible for such folding also developed more or less strong schistosity throughout much of the leptite.

In addition to the direct effect of the hydrothermal fluids on the iron oxides and quartz of the primary leptites, converted much of the leptite to mica schists and quartzites rich in such minerals as cordierite, almandite, andalusite, gedrite $[(Mg,Fe^2)_5Al_2(Si_6Al_2)O_{22}(OH)_2]$, anthophyllite, and cummingtonite. The synorogenic hydrothermal fluids also appear to have been responsible for the intense magnesia metasomatism and the deposition of sulfides that locally were responsible for such deposits as Falun and Garpenberg.

I should emphasize that I think that the sulfide deposits associated with the Bergslagen iron ores and unrelated to iron ores were much more probably the result of the activity of the hydrothermal fluids derived from the synorogenic granitic magmas than they were of primary volcano-sedimentary processes much earlier in the geologic history of the Bergslagen rocks. The presence of sulfides in the same rock volumes as the various types of iron ores seems to have resulted from the favorable conditions for sulfide precipitation provided by the silica-carbonate environments in which most of the sulfides are found. The leptites with less carbonate (particularly) or silica-quartz, on the contrary, did not furnish as effective a medium for metasomatic or open-space deposition of sulfide.

After the synorogenic granite phase was completed with its attendant effects on the banded iron ores in the primitive leptites, an intra-orogenic period followed in which the folded supracrustal rocks (including those younger than the leptite sequence - mainly graywackes and shales) were intruded by mafic magmas (mainly diabases, now amphibole-rich greenstones); these magmas do not appear to have accompished any ore deposition.

At the end of this intra-orogenic period, much of the central Swedish region was forced downward to great depths where palingenic processes (encompassing more or less complete fusion) were able to produce a younger generation of granitic magmas at depth. Above the depths conducive to fusion, the overlying rocks were converted to coarse-grained pegmatitic gneisses; these rocks are quite inhomogeneous and include appreciable volumes of coarser pegmatite veins. Locally, fragments of the original supracrustal rocks are preserved in the gneisses. The conversion of the remaining supracrustal rocks, plus some of the synorogenic granites, to palingenic gneisses and granites must have been carried to varied degrees of completion because all gradations between the supracrustal complex of leptites, graywackes, slates, granite gneisses, and pegmatitic gneisses still remain. The late pegmatites are, according to Magnusson (1970), both true intrusive pegmatites and metamorphic differentiates formed in place.

How much of the alteration of the pre-palingenic rocks was due to solutions developed in the palingenic process and how much was caused by solutions from the synorogenic magmas probably never will certainly be determined. It appears, however, that the sulfide bodies were emplaced before the palingenic magmas had been developed and the additions of iron to the iron-ores primary to the leptites occurred even before the sulfides were deposited. This suggests that the palingenic magmas had little to do with the ore-forming process in Central Sweden.

How much of the development of the skarn and carbonate-bearing, quartz-rich deposits truly can be designated as contact-metasomatic (or hypothermal in carbonate rocks) also cannot be stated with certainty. But it is probably true that most, if not all, of the skarn-bearing iron ores certainly owe their present character, if not their entire iron content, to hypothermal-range processes. On the contrary, the bulk of the quartz-rich, carbonate-poor ore bodies probably received additions of iron above the amount that was introduced into them during the volcano-sedimentary processes that produced their initial form and iron content. Thus, it seems reasonable to add the categories hypothermal-2 and hypothermal-1 to the designation Sedimentary-A3 fully to describe the manner in which the Central Sweden iron ores were formed. [For more detail, see Grip (1978) and Magnusson (1925-1970).]

GRÄNGESBERG, KOPPARBERG

Middle Precambrian	Iron as Magnetite, Hematite	Magmatic-3b Hypothermal-1 (minor)

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Notes

The Grängesberg (58°49'N, 15°06'E) iron deposits are located in central Sweden, nearly 200 kilometers slightly north of west from Stockholm. Grängesberg is the most important of numerous iron deposits of the Western Ludvika district, a narrow belt only 40 kilometers long. Other major deposits immediately north-northeast of Grängesberg (though much less notable than that field) include Blötberget, Fredmundberget and, on the north side of Lake Våsman, Lekomberg. Some 20 kilometers, still further north-northeast, is Idkerberget and a few other and smaller deposits.

The apatite-bearing ores of the Grängesberg district are by far the most important. The total area of ore is 90,000 km^2; of this, 55,000 km^2 are concentrated in the Export field that forms the southeast margin of the Grängesberg ore bodies and is 1000 meters long by 20 to 90 meters wide. This field ends in both directions in several wedges; the plunge of the ore is 55° to 70° south-southeast. In 1973, almost 4,216,000 tons of ore were recovered from the Export field; this ore assayed 59.4% Fe, 0.89% P, and 0.12% Mn. The smaller Risberg field on the other side of the iron-ore belt, almost due north of the Export, produced nearly 715,000 tons of iron ore in the same year. Apatite is recovered by flotation from the Grängesberg ore, production being from 25,000 to 30,000 tons per year.

The Grängesberg ore has been known since at least the end of the 16th century and probably for an appreciable time before that. In the early 1960s, production was about 2,000,000 tons per year, so the tonnage recovered has more than doubled in the past 20 years. Until the Bessemer (Thomas) process for the treatment of high-phosphorus ore in the 1880s, no real demand for the ore existed. This basic Bessemer process eliminates the phosphorus in the pig iron through reaction with lime added to the charge and with the dolomite brick lining of the furnace. If the ore is high enough in phosphorus, the slag can be ground and used as fertilizer. If it is not of sufficient grade, calcium phosphate can be extracted from it and similarly used. Until the Bessemer process was perfected, the pig iron from the Grängesberg ore was too brittle for use in any industrial application.

Since Grängesberg was discovered, it probably has produced over 100,000,000 tons of ore and has reserves for at least 50 years further production at the present rate. The ore has been proved to 1000 meters below the surface, and there is no reason to suppose that the ore does not continue below that depth. Grängesberg probably is comparable in size to the Cornwall mine in Pennsylvania, although the two ores were formed in much different ways.

The deposit is the property of the Grängesberg Company, a huge Swedish industrial and shipping complex that, until 1957, operated the Lapland iron ore deposits (in which it had essentially a half interest) for the Swedish government. Grängesberg is about four hours by train from Stockholm in rolling stock that was (at least until 1963) owned by the Grängesberg Company and largely over the Gränegsberg Company's right-of-way. The Grängesberg railroad is by far the largest of the privately owned railroads in Sweden. There are other private roads, but most of the trackage in that country is owned by the National Government.

The Grängesberg ore (and its smaller counterparts) is unusual in the central Swedish area in that it is largely, though not entirely, an iron-apatite ore. Most of the iron ores in central Sweden belong either to the so-called quartz-banded or to the skarn-type ores. It is beyond the scope of this discussion to present arguments as to how either the quartz-banded or skarn ores got where they are or what relationship there may be between them. For a summary of the facts about these ore types, see the discussion under the heading; "Central Sweden".

The Grängesberg area has been highly glaciated, as would be expected in central Sweden, and it is most fortunate that, when at least part of the Grängesberg field was being prepared for mining, all the glacial debris was stripped off and the entire surface so exposed was geologically mapped on a large scale. Thus, a great deal more is known about the geology of this area than would have been possible under normal mining operations.

Two works on the geology of the Grängesberg area, published in the late 1930s, were printed as bulletins of the Swedish Geological Survey, one by Magnusson (1938) and one by Looström (1939). Magnusson, a former director of the Swedish Survey, studied the entire Grängesberg district; his work is in German, but the huge fund of information that it contains makes it well worth translation. Looström's work concerns only the Lönnfallet (southern) portion of the district, and his area does not present all of the problems of Grängesberg geology. Nevertheless, his study was so carefully carried out and his maps made on so large a scale that his efforts have done a great deal to explain the geology of the entire area. Landergren's (1943) geochemical studies of the Grängesberg deposits, although containing only a short English summary, provides vital information as to the concept of the origin of the deposits held by Swedish geologists.

The principal problems of Grängesberg geology are: (1) the genetic relationships between the two ore types (a) hematite-minor magnetite impregnation ore and (b) magnetite-lesser hematite-apatite largely massive ore; (2) the spatial and genetic relationships of the skarns to both ore types and of those two types to each other; and (3) the genetic significance of the paragenesis of the magnetite and hematite in the apatite-bearing ore. These problems will be considered after the general geology setting has been outlined.

The oldest rocks of the area and the material into which all later rocks and ores were introduced are know as leptites. Leptite is a Swedish term designating acid volcanics - pyroclastics, tuffs, and flows - in which the average grain size lies between 0.03 and 0.05 millimeters. Acid volcanics of similar types but having an average grain size greater than 0.05 millimeters are called leptite gneisses, and those with the average grain size less than 0.03 millimeters are known as halleflinta; these two latter types are not present in the immediate Grängesberg area. The leptite probably was emplaced in the late Early Precambrian or early Middle Precambrian times, with the latter being more likely.

There are two principal and one minor varieties of leptite at Grängesberg. The first of the main types is a red potash-leptite, and the second is a gray soda-leptite; the variant is a strongly porphyritic potash-lime-leptite that is closely related to the potash-leptite. According to

Magnusson (1938), the red potash-leptite contains microcline-microperthite, oligoclase, and biotite; there is minor muscovite but no hornblende. The quartz is in the ground mass only; accessories are apatite and orthite*. The composition averages about 68 per cent SiO_2, 13 per cent Al_2O_3, three to seven per cent K_2O, and 1.5 to 4 per cent Na_2O.

The gray soda-leptite is a microcline, quartz, oligoclase-albite rock. It has varied amounts of biotite and is, in many areas, porphyritic. There is some argument as to whether the porphyritic texture was original or not, but it probably was. Accessories are apatite and magnetite. The silica content is a little higher than that of the red potash-leptite, being about 69 per cent SiO_2; there is between 13 and 14 per cent Al_2O_3, 4.5 to 6 per cent of Na_2O, and only 1 to 2 per cent K_2O. The potash content of 1 to 2 per cent is much less than the 3 to 7 per cent of the red potash-leptite, and the soda content of 4.5 to 6 per cent is much higher than the 1.5 to 4 per cent soda content of the red potash-leptite.

The spatial relations of the leptites to each other are most complex. The lime-potash variant of the potash-leptite is almost everywhere porphyritic. It contains andesine or basic oligoclase. The groundmass is quartz, microperthite, and oligoclase, with some hornblende and biotite. Accessories are orthite, titanite, epidote, magnetite, and apatite. Both the red potash-leptite and the lime-potash-leptite contain garnet-epidote skarn, and in both is found the impregnation hematite. In the gray soda-leptite, on the other hand, an amphibole-pyroxene skarn was developed, and into this rock type was introduced the apatite-magnetite-minor-hematite ore. Thus both skarn and ore associated with each of these two types of volcanic rocks are quite different.

Some of the "leptites" actually are andesite and dacite dikes that are younger than the ore and its host leptites but Magnusson (1938) considers that they belong to the leptite group even though the age difference between the two rock types is considerable.

Before the formation of any younger rocks, the leptite proper was complexly folded. Following the folding, two events occurred, the time sequence and genetic relations of which are uncertain. These events were: (1) the development of three types of skarn, one in close association with the massive ore and the other two developed in considerable fractions of both leptites over much wider areas than those in which ore is found, and (2) the introduction of the two ore types.

Probably much of the confusion as to the relationships of ore and skarn has come about because of the failure of those who have studied the district to make clear that there are three kinds of skarn, plus a quartz-rich alteration of the leptite, which last is directly associated with the hematite impregnation ore. The garnet-epidote skarn in the potash-leptites and the pyroxene-amphibole skarn in the soda-leptite developed first. The areal extent of these skarns is so much greater than that of the ores and position of the ore bodies is so removed from the centers of skarn development that it appears unlikely that these skarns could have been formed as a by-product of ore emplacement. It seems more likely that these skarns were formed by hydrothermal solutions that entered the area prior to the introduction of the ore and that only one variety of hydrothermal fluid was involved. The differences between the two skarn-types were caused by differences in the host rocks rather than by two distinct hydrothermal fluids from separate magma volumes. Obviously, this portion of the skarn problem needs more study.

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* Orthite, also known as allanite, is a calcium rare-earth mineral in which uranium may substitute for the rare earths; if it does, there ordinarily are radioactive halos around it.

The third skarn type is found in close association with some of the magnetite-minor-hematite-apatite ore; its greatest development was in the Lönnfallet ore at the southwest end of the main ore field. The Lönnfallet skarn is predominately amphibole, with minor biotite and apatite, while that in the main Export Field is largely biotite and appears to be much less abundant. This third type of skarn was probably formed concomitantly with ore and was, therefore, closely related in genesis to the ore. The actual mechanism of its development will be considered later in the discussion of ore genesis.

The quartz-rich alteration associated with the impregnation-type hematite ores appears to have formed contemporaneously with the iron ore. Around these ore bodies, the leptite is enriched in quartz, with both quartz enrichment and amount of hematite becoming progressively less away from the ore. The centers of the quartz and hematite mineralization are so far removed from that of the garnet-epidote skarn that the developments of these two modifications of the leptite probably had no direct genetic connection, although both occur in the potash-type leptites. A mechanism for the formation of both quartz alteration and hematite impregnation will be suggested in the discussion of ore genesis.

The two types of ore in the Grängesberg area - magnetite-minor, hematite-apatite and impregnation hematite - probably were introduced after the formation of the pyroxene-amphibole and garnet-epidote skarns, respectively, but at the same time as the amphibole-biotite-apatite skarn and the quartz alteration, also respectively. The relationships postulated, however, are difficult to demonstrate. The apatite-bearing ores, in the molten form, certainly were emplaced in the leptites either as *lit-par-lit* injections of various magnitudes or as the fracture fillings in brecciated soda-leptites. Further, the hydrothermal impregnation hematite ores certainly were not primary constituents of the potash-leptites and must, therefore, be younger than the volcanic rocks.

Where impregnation ore is found in skarn-bearing potash-leptite, the hematite and its associated quartz alteration appear to have replaced both primary leptite and skarn minerals; the skarn minerals were not formed as a phase of the ore-quartz emplacement process. The relationship between the widespread pyroxene-amphibole skarn and the amphibole-biotite-apatite skarn intimately connected with the apatite-bearing magnetite ore is not clear, largely because no attempt has been made, in what geologic maps have been made of the area, to differentiate between the two. Nevertheless, because the garnet-epidote skarn is older than the impregnation ore, it is here considered more likely that the pyroxene-amphibole skarn is older than apatite-bearing magnetite ores than that the reverse is true.

All the other rock types clearly cut the leptites and, except for the granites, cut ore and skarn as well. The first rocks emplaced after the ores were what Looström calls the older amphibolites and what Magnusson calls andesites and dacites (dacite is used here as meaning the extrusive equivalent of quartz diorite and andesite as meaning the extrusive equivalent of diorite). Looström believed that he could distinguish two ages of older amphibolites (1939), the older being strongly schistose and the younger porphyritic. Magnusson (1938) was unable to make this distinction. The dacite and andesite were originally composed of plagioclase (oligoclase to basic andesine), biotite, and hornblende in varied proportions with the quartz content determining which of the two names is given to any specific specimen. The accessory minerals included apatite, magnetite, and titantite, these differing as widely in abundance as the principal constituents do. The chemical composition of a typical dacite is some 64.5 per cent SiO_2, 14 per cent Al_2O_3, 3 per cent Fe_2O_3, 4 per cent FeO, 2 per cent MgO, 4 per cent CaO, 5.3 per cent Na_2O, 0.5 per cent K_2O, and 0.2 per cent P_2O_5. On the other hand, the andesite contains about 57 per cent SiO_2, 14 per cent Al_2O_3, 3.5 to 6.5 per cent Fe_2O_3, 5 to 10 per cent FeO, and 0.5 per cent P_2O_5.

Andesite and dacite were variously associated; one dike might have been all andesite, another all dacite, and in a third andesite may have graded into dacite. It is certain that these dike rocks were not responsible for ore or skarn, but their accessories suggest that they may have come from the same general source.

The next igneous rocks to be introduced were the original or Ur granites of which there were two types, the red salic granite to the west of the ore area and the red intermediate granite to the east. It is not certain if these two granites are cut by the andesite and dacite dikes, but Magnusson (1938) believes the granite to be younger. The two kinds of original granite differ slightly in composition. The red salic Ur granite is a quartz, microcline-perthite, albite-oligoclase rock; there are small amounts of biotite, and the accessory minerals are orthite, zircon, hematite, magnetite, and fluorspar. The composition of the red salic Ur granite is about 77 per cent SiO_2, 12.5 per cent Al_2O_3, 4 per cent K_2O; and the P_2O_5 content is about 0.07 per cent.

The intermediate original granite contains quartz, microcline-perthite, oligoclase, biotite, and hornblende in various proportions. The accessory minerals are titanite, apatite, zircon, and orthite, plus magnetite and pyrite. There is less quartz and microcline than in the red salic granite and more plagioclase and dark minerals. There may be 5 to 10 per cent dark minerals in the intermediate granite, but much less than this in the red salic granite. There may be a considerable myrmekite developed in the feldspars in the intermediate granite but not in the red salic granite. Calcite and epidote are present as alteration products of plagioclase in parts of intermediate granite. In chemical content, the intermediate granite contains 69 per cent silica, a little over 13 per cent Al_2O_3, about 3.5 per cent Na_2O, and 3.5 per cent K_2O; the P_2O_5 content is about 0.16 per cent as opposed to 0.07 per cent in the red salic granite.

In the eastern part of Lönnfallet area, Looström (1939) recognized a third variety of amphibolite in which are included oblong segments of aplite. This amphibolite cuts the older amphibolites (andesites and dacites) and exhibits sharp, clean contacts against them. The younger amphibolite is somewhat more mafic than the average of the andesites and dacites.

All the rocks and ores thus far described show definite and generally marked effects of regional metamorphism although the structural pattern of the leptites is much more complex than that of the younger rocks.

The pegmatites cut all the rocks mentioned so far and show no evidence of regional metamorphism or of appreciable folding; they are, therefore, among the latest rocks introduced into the area. They are considered by Magnusson (1939) to be related to Svecofennian granites known outside the immediate Grängesberg area and to be appreciably younger than the ores and probably younger than the entire Svecofennian orogeny as well.

The pegmatites are composed predominantly of a pale red microcline-microperthite and of oligoclase of varied grain-size and structure; generally, they are completely lacking in dark minerals but contain some accessory apatite and beryl. Where the pegmatite cuts the apatite-bearing magnetite ore, idiomorphic crystals of magnetite in a hematite matrix are often found, and it has been suggested that this hematite resulted from oxidation of magnetite by emanations from the pegmatite. This may well be true, but large volumes of ore showing similar magnetite-hematite relations are known in areas that contain no pegmatite. Here and there small amounts of molybdenite and scheelite appear in the pegmatites, and the wall rocks of the pegmatites are locally enriched in beryllium, molybdenum, tungsten, and tin - a relationship known in the wall rocks of pegmatites throughout the world.

The youngest rock in the area, as in so many other ore districts of the world, is a diabase that cuts across the pegmatites and all older rock types of the Grängesberg ore field. Ore cut by the dikes appears to be unaffected by them.

The first rocks developed in the Grängesberg area were the varieties of leptite; after these volcanic rocks had been completely folded, the three skarn types, the quartz alteration around the impregnation ores, and the impregnation (hematite) and massive (magnetite-minor hematite-apatite) ores were emplaced. Then mafic intrusives (andesite and dacite, now amphibolite and locally at least a younger amphibolite), and two varieties of granite were introduced; all these rocks were regionally metamorphosed. Then followed the intrusion of pegmatite; and, probably after an appreciable interval, diabase dikes were formed. There are no true sediments in the area such as the limestones and dolomites found interbedded with the leptites in other parts of central Sweden.

The macrostructure of the Grängesberg deposits is quite simple. The rocks in general bear a strong northeast-southwest foliation, and the general dip of the formations is 65° southeast. In detail, however, the rock patterns differ widely, each of the early rock types is intricately interlayered with the others, and only the very latest intrusives strike at high angles to the grain of the country. The intermixture of leptites with the later rocks is highly complex, a complexity largely inherited from the intricate folding that affected the leptites before any other rock types had been introduced into the area. No published studies have attempted to explain the intricacies of the original folding, and it may be that the original fold-pattern has been so modified by the successive introductions of ores and rocks that it can never be completely unraveled.

It appears from radioactive age-dating that the leptites were emplaced in the area some 2600 m.y. ago, while the latest rocks that can be approximately dated (the pegmatites) were intruded after the Svecofennian orogeny had essentially run its course. The generally accepted dates for this later orogeny are 1800 m.y. to 1700 m.y. Thus, the leptites were probably formed in the latest Early Precambrian and were folded in the orogeny that was associated with the end of that epoch. Since the ores were introduced after the folding of the leptites and well before the end of the Middle Precambrian (1600 m.y.), the ores can with confidence be dated at Middle Precambrian. Data are lacking that would permit a more accurate placement of the ore-forming period in that portion of geologic time.

The magnetite-minor hematite-apatite ores exist mainly in quite irregularly ellipsoidal masses intricately interlayered and in sharp contact with soda-leptite and the mafic intrusives, both in plan and in section. Detailed plans of these ore lenses are given in the three plates accompanying Looström's paper, and sections are given in Magnusson's Figures 5 and 6 (1938). From these diagrams it can be seen that the ore ellipsoids differ widely in horizontal and vertical dimensions and in detail of outline. Some of the massive ore is bordered in part by brecciated soda-leptite, the matrix of which is magnetite ore (and skarn). The impregnation hematite ores, on the other hand, are more vein-like in character, with the vein centers being largely hematite, with some quartz and residual potash and lime-potash leptite. Outward from the vein-centers the amount of hematite and quartz grows progressively less and unaffected leptite increases.

Where skarn-rich areas are associated with impregnation ores, the garnet-epidote content of the rock volume in question shows no systematic relation to the ore. The magnetite ore usually contains no skarn minerals, although skarn composed of amphibole, biotite, and apatite or any one or two of these minerals may be quite abundant locally. The cement of brecciated leptite is generally high in skarn made up of the same nonmetallic minerals.

The impregnation ore is almost entirely hematite, although in the rare localities where impregnation ore is found in soda-leptite, it consists mainly of magnetite (Magnusson, 1938). Quartz is essentially the only gangue mineral contemporary with hematite in the impregnation ore, although Magnusson (1938) considers the garnet-epidote skarn to have formed at much the same time as the quartz and hematite.

The mineralogy of the magnetite-minor hematite-apatite ore is, as its designation suggests, more complex. Hematite is most common in the footwall: magnetite grows steadily more abundant as the hanging-wall is approached and as greater depths are reached anywhere in the deposit. In the main part of the Grängesberg Field, the amount of skarn (or gangue) minerals in or adjacent to the ore, formed at generally the same time, is minor, except for apatite which, although not abundant, is widespread, with its greatest concentrations being in the hanging-wall portions of the ore masses. The nonmetallics other than apatite are mainly biotite with less amphibole and minor quartz. These nonmetallics appear as individual crystals or clumps of crystals in the ore, whether it is hematite or magnetite. In a few locations in the Lönfallet area, however, the gangue (or skarn) in the ore is more abundant and consists mainly of hornblende, with locally abundant idiomorphic crystals of apatite and here and there magnetite in fine-grained lumps of varied size (Looström, 1939).

On the hanging wall of the Lönnfallet ore, there is an appreciable thickness (20 meters or more as a maximum) of brecciated leptite cemented by a hornblende, apatite, and ore. In narrow fractures the matrix material is mainly fine-grained hornblende, but in wider areas small accumulations of ore occur as fine-grained masses of magnetite and apatite centrally located in the hornblende (Looström, 1939). Throughout the ore there appear to have been essentially no reaction rims developed between ore and gangue (skarn).

In the magnetite-minor hematite-apatite ore of the Grängesberg Field, magnetite and hematite occur in varied relationships to each other. About 80 per cent of this ore is magnetite; it is not only generally the dominant of the two iron minerals but is also usually the earlier. In numerous locations, magnetite octahedra are found in a groundmass of magnetite or magnetite and hematite. Where both iron oxides are present in the groundmass, the more common relationship is for veinlets of hematite to cut the magnetite. Locally, however, veinlets of magnetite cross through hematite, and there are even areas where platy crystals of hematite are found in a groundmass that is usually partly hematite and partly magnetite, with the magnetite the younger of the groundmass minerals.

Apatite is normally found as automorphic crystals in the iron oxide mass. As apatite is most abundant in the hanging wall and as the greater concentrations of hematite are found in the footwall, the apatite crystals usually are found in a magnetite matrix.

The impregnation ore grades from nearly massive hematite with minor secondary quartz and residual leptite through rock that contains less and less hematite, less and less quartz, and more and more unaltered leptite - this strongly supports the idea that this iron ore was emplaced hydrothermally. Although only a minor fraction of the total Grängesberg production has come, and essentially none now comes, from this ore type, its presence definitely indicates that hydrothermal solutions, capable of transporting and depositing iron ore, invaded the ore area. Almost all this ore was deposited in the potash- and lime-potash-leptite, and in this environment it was deposited as hematite.

In a few localities small amounts of disseminated ore were emplaced in soda-leptite, and in this rock the disseminated ore was magnetite. The pattern of the skarn bodies in the potash and potash-lime leptite is so unrelated spatially to that of the impregnation ore and the skarn is so much more widespread that, although probably also hydrothermal in origin, it must have been formed at a somewhat earlier time by solutions from a different source or by solutions from the same source driven off before they had achieved the composition necessary to deposit hematite in quantity.

The manner in which the massive magnetite-minor hematite-apatite ore was emplaced, however, seems to have been entirely different. The large volumes of this ore that contain little or no so-called skarn show knife-sharp contacts with the leptite; this lack of alteration between ore and leptite suggests a molten ore-forming medium in which the content of the water was low. Certainly there is a complete lack in the massive ore of even the quartz alteration associated with the impregnation ores. These massive ores do, nevertheless, commonly contain apatite in varied, but still noticeable, amounts. The automorphic outlines of most, if not all, of the apatite indicate that calcium phosphate was an essential component of the iron-rich melt and that it crystallized before at least most of the iron oxide.

In most of this massive ore, occasional crystals of biotite and, more rarely, amphibole (often altered to chlorite) and quartz make up the remaining minerals of ore. Silicate minerals, aggregated into appreciably more than single crystal masses, have been referred to as skarn, but they might be better described as accessory (and primary) constituents of the ore. There appears to be no doubt, however, that these early silicate crystals were precipitated from the ore-forming medium. This evidence strongly confirms the widely held opinion that the ore-forming medium was an iron-oxygen-rich melt that also contained small amounts of apatite and lesser quantities of silicate material. The presence of these minor constituents, however, appears to have lowered the melting point of the ore-magma, as the unpublished experimental work of Tuttle shows to be possible, so that it probably was introduced at temperatures between 900°C and 600°C.

Locally, particularly in the Lönnfallet area, much larger amounts of nonmetallic minerals (skarn in Grängesberg terminology) are found closely associated with the ore. Looström (1939) reports that the skarn occurs: (1) as distinct, sharp-walled, oblong or lens-like inclusions in massive ore or at the boundaries between massive ore and leptite; these inclusions normally have ramifying tongues that follow the general strike of the area; (2) as fillings in the interstices of brecciated ore - fillings that show no evidence of reaction with the ore; and (3) as matrix material in brecciated leptite on the hanging wall of the ore body; in this latter situation also, the contacts between skarn and the leptite fragments are knife-sharp.

The skarn in all three of the above occurrences consists mainly of hornblende, although, particularly in the larger skarn masses, there may be minor sharp-walled inclusions of magnetite and apatite. Thus, it might appear that iron ore was introduced both before and after the skarn because skarn both fills fractures in the main masses of iron ore and has fractures in itself filled by ore veinlets that apparently do not connect with the massive ore. In most places, however, the skarn masses are so irregularly and intimately intermingled with the ore as to suggest strongly that they were derived from the ore rather than that they were introduced into it in any form at a later time. It appears that these relationships can be best explained by assuming the separation of the nonmetallic materials as an immiscible liquid from the associated and larger volumes of iron-ore melt. This explanation would require that the silicate materials of the skarn masses were initially dissolved in the molten ore and that, as the melt cooled, the solubility of the silicates in it steadily dropped until the amount present exceeded the quantity the melt could hold in solution. At this stage, molten silicate material would have begun to separate from the ore melt and gradually, as the melt continued to cool, would have accumulated to form larger and larger masses that were in essential equilibrium with the ore melt surrounding them.

To produce the skarn found at Grängesberg, these molten masses of nonmetallics would have had to contain not only the components of the dark silicates of the skarn (mainly hornblende) but also appreciable quantities of apatite and iron ore as is indicated by the presence of apatite and magnetite as accessory, or even essential, constituents of the solid product. The presence of such small amounts of apatite and iron ore in the silicate (skarn) masses, as well as the presence of minor silicates in the massive iron ore, is in accord with the origin suggested. The separation of a homogeneous melt into two immiscible melts is never a "clean" one. Each of the immiscible melts will be saturated in the components of the other, though the actual amounts of these components will be small, and the final crystallization product of each melt will include small amounts of the minerals that are the principal products of the crystallization of the other melt.

At Grängesberg, the two molten phases must have coexisted together over an appreciable period of time, but the iron-rich phase appears to have crystallized first. This is indicated by the presence of the nonmetallic phase as the cement of the leptite breccia masses of the hanging wall of the ore bodies. Had both ore and nonmetallic material still been molten at the time of brecciation, both would have been forced into the fractures and both would have been found cementing the leptite fragments. Instead, only the silicate phase provides this cement, so the iron-rich melt must, by that time, have been solid or essentially so.

From these silicate melts (with their minor content of apatite and magnetite, the presence of which probably accounts for their crystallization temperatures being lower than those of the iron-rich melts) that were forced into the brecciated leptite, hornblende solidified first and the magnetite-apatite residue last, with the apatite crystals being idiomorphic in the magnetite. As a result, the silicates fill all the finer fractures and form the outer margins of the larger openings in the leptite, while the magnetite and the apatite are found in the cores of the larger silicate masses. Between these silicate bodies and the larger masses of iron ore that contain or border them, there are no reaction rims, but only sharp contacts; the same relationship obtains between these silicates and the smaller bodies of magnetite-apatite they contain. Thus it would appear that the magnetite-apatite material (whether in the molten or solid state) was in essential equilibrium with the silicates (in either molten or solid form). It must, however, be emphasized that these masses of nonmetallics (called skarn by Looström) are very minor in volume when compared with the total volume of magnetite-minor hematite-apatite ore and should not be equated with the pyroxene-amphibole skarn developed hydrothermally in far larger volumes of leptite. The division of apatite between the iron-rich melt and the molten silicate phase appears to have placed a greater proportion of the apatite into the former as the amount of apatite in the skarn (gangue) is small, and the amount of apatite in the iron ore is not only greater in proportion to that in the skarn but is much greater when the huge total volume of ore is compared with the small quantity of skarn (gangue).

The minor amounts of biotite in the major portions of the magnetite-minor hematite-apatite ore appear to have come out of solution as crystals. This difference in behavior from that exhibited in portions of the Lönnfallet ore that contain amphibole skarn probably can best be explained by the difference in silicate content of the melts in question. Where the iron oxide melt contained considerable silicate material (as it did locally in Lönnfallet), separation of the silicate material occurred as an immiscible liquid; where the melt contained little silicate in solution, separation took place below the crystallization temperature of biotite.

Within the iron ore, magnetite and hematite both exist, as has already been described, in relationships that suggest that in certain areas magnetite formed first and in others that hematite developed before magnetite. These relationships appear to depend directly on the partial pressure of nascent oxygen in the melt and the time in the crystallization cycle when the gas was made available. In considering Kiruna and Gällivare, I have discussed my reasons for believing that this oxygen did not derive from the dissociation of water but came, instead, from the dissociation of nitrogen-oxygen compounds. This suggestion, *mutatis mutandis*, appears equally well to explain the magnetite-hematite relations at Grängesberg.

The bulk of the apatite ore at Grängesberg has magnetite both as phenocryst-like crystals and as finer-grained groundmass. In this ore, hematite is normally present as irregular veinlet-like masses cutting the fine-grained magnetite. On occasion, in the more hematite-rich portions of the ore, the bulk of the ore may be fine-grained hematite crossed by magnetite veinlets, and less commonly hematite plates occur as phenocrysts in a groundmass of hematite containing magnetite veinlets.

From these relationships, it appears that most of the ore was introduced as a magnetite-apatite melt which contained a source of nascent oxygen, such as nitrogen pentoxide, from which oxygen was not normally released until the latter part of the crystallization cycle of the ore magma. As the breakdown of such an oxygen-containing source as nitrogen pentoxide is favored by decreasing pressure (see Equation 1):

$$\begin{aligned}(1)\quad N_2O_5 &= 2NO_2 + O\\ 2NO_2 &= 2NO + O\\ 2NO &= N_2 + O_2\end{aligned}$$

this would suggest that the drop in pressure necessary to initiate the release of oxygen did not occur until the bulk of the iron-rich melt had crystallized as magnetite, some of which may have solidified under sufficient pressure to permit the growth of the large magnetite octrahedra already mentioned. The reaction to convert a portion of the melt to hematite is shown in Equation 2:

$$(2)\quad 2Fe^{+2} + 2Fe_2O_4^{-2} + O = 3Fe_2O_3$$

The Grängesberg ores, as Schneiderhöhn emphasizes (1958), were formed much nearer the surface than those at Kiruna; Grängesberg ore contains about 20 per cent hematite; that at Kiruna, essentially none. This would indicate that the difference in confining pressure between the two deposits might have been an important factor in the degree of oxidation of the iron in them. The lowered pressure at Grängesberg would have favored such dissociation reactions as those given in Equation 1, and it appears reasonable to assume that some such reaction as that put forward in Equation 1 may have provided the oxygen needed to convert an important amount of the magnetite in the Grängesberg ore-magma to hematite. Once the reaction of Equation 1 had been appreciably driven forward by the lowering of confining pressure on the melt, it would be carried essentially to completion by the consumption of oxygen in the reaction of Equation 2. At Kiruna, the degree of dissociation of N_2O_5 was never sufficient to permit any noticeable change in the Fe^{+2}/Fe^{+3} ratio from that initially existing in the ore-magma from which that deposit was formed.

It appears that the confining pressure in portions of the Grängesberg ore magma dropped with sufficient rapidity that nascent oxygen was released before any crystallization had occurred. In such circumstances, the molten material would have been converted to Fe_2O_3 to the limit permitted by the available supply of oxygen. Because, in nearly every instance I observed, the hematite-rich ore contains some late magnetite, it would seem that seldom, if ever, was enough oxygen released to oxidize completely to hematite all the iron in any sizable volume of the iron-rich melt. In those ore volumes in which phenocryst-like crystals of hematite are found, the amount of oxygen-bearing material available should have been larger than in a melt-volume from which early hematite (but no hematite phenocryst plates) were developed. If this is the case, a higher proportion of hematite (of both varieties) would be expected in an ore volume containing hematite phenocrysts than in one from which they were absent. Quantitative checks on this relationship have not been made as yet.

No one today suggests that the iron ore was derived from the leptite by differentiation *in situ* although Johansson (1910) not only thought that this was true but also believed that the other rocks of the Grängesberg complex also were formed by this single process. The detailed work of Magnusson and Looström (1938, 1939) shows that this cannot be true, but they do not demonstrate conclusively what might have been the source of the hydrothermal fluids and of the water-poor melts.

The leptite, the mafic intrusives, and the intermediate Ur granite all contain magnetite and apatite as accessory minerals, suggesting that the source magma(s) of these igneous materials also may have been the source of the iron-apatite ore melt. The ages of these igneous rocks are not, of course, the same as that of the ore melt, but all of the molten material that produced these rocks was formed in the same general portion of the Precambrian and may have come from the same volcanic hearth (to use Magnusson's expression). It is, at present, impossible to say, however, that the ore melt was more closely connected with any one of the magmas that with the others.

In any event, the magma from which the original iron-apatite-minor-silicate melt came must have been rich enough in the constituents of the ore melt that the parent magma became saturated in these materials at temperatures considerably above that of the beginning of crystallization of the silicate minerals from that magma. The temperature ranges over which crystallization took place in the ore melt, on the one hand, and in the silicate melt on the other, must have been sufficiently different that the silicate-melt portion of the parent magma solidified to a considerable extent before crystallization began in the ore melt. Had this not been true, molten silicate material would have been intruded at much the same time as the ore melt, a situation that did not obtain.

Thus, to summarize, the Grängesberg deposits consist of major masses of magnetite-minor hematite-apatite ore intruded as water-poor melts and minor bodies of hematite impregnation ore (this last having been deposited from hydrothermal solutions). Almost all of each ore type is located within one variety of leptite - the magnetite-apatite ore in the soda leptite and the hematite ore in the potash leptite.

The two widespread varieties of skarn (pyroxene-amphibole skarn in the soda leptite and garnet-epidote skarn in the potash leptite) cover so much larger areas than the ores and are so eccentrically located in relation to the centers of ore mineralization that ores and skarns almost certainly cannot have been formed by the same process. It is suggested that the widespread skarns were developed hydrothermally prior to ore emplacement.

The third type of skarn (better designated as gangue) is so intimately associated with the magnetite-apatite ore that it definitely appears to have been formed concomitantly with this type of ore. It seems most probable that the constituents of this third skarn were introduced in solution in the iron oxide-calcium phosphate melt but became immiscible in that melt before it began to crystallize. The very irregular distribution and minor amount of this third skarn suggests that the silicate content of the iron-oxide melt differed widely from one place to another; only from those portions of the melt containing considerable amounts of dissolved silicates did sizeable masses of skarn silicate material separate from it. In the bulk of the magnetite-apatite ore, the skarn (or gangue) is represented only by an occasional crystal (or crystal group) of biotite. This third skarn type forms the matrix of considerable volumes of leptite breccia located mainly in the hanging wall of the magnetite ore and, in many places, contains cores of magnetite and apatite, suggesting that, as is only reasonable, some iron-phosphate material was dissolved in the molten skarn and crystallized mainly after the silicate minerals.

In the magnetite-minor hematite-apatite ore, the hematite appears to have been developed after the molten masses had been intruded, usually late in the crystallization cycle. This suggests that the oxygen necessary for this process was generated in the melt, relatively late in the cycle, although locally early hematite crystals indicate that, at times, the oxygen became available earlier. The concept is put forward here that this oxygen was derived from the dissociation of nitrogen-oxygen compounds such as N_2O_5, a process favored by falling confining pressure.

It follows from the discussion that has been given thus far that the magnetite-minor hematite ore was introduced in the molten state. This material did not have a great effect on the enclosing leptite as the skarn around the ore is minor in amount and consists almost entirely of apatite. This suggests that little silicate material was dissolved in the molten iron-oxide; in some of the brecciated leptite on the hanging wall of the Lönnfallet, however, zones up to 20 m thick may be cemented by hornblende, apatite, and ore. This ore, therefore, almost certainly should be categorized as Magmatic-3b.

The minor amount of impregnation hematite ore found in the Grängesberg area appears to have been introduced by hydrothermal solutions that also deposited appreciable quartz. These solutions probably did not come directly from the molten magnetite but more likely were generated in the late stages of crystallization of the magma chamber from which the molten magnetite-apatite ore came. This type of ore should be classified as Hyperthermal-1.

In opposition to this mode of genesis, Landergren (1943) says that "Grängesberg iron ores certainly should be regarded as magmatic (on account of their geological occurrence) but as of secondary or palingenic formation. The process of differentiation which must be considered to have played a part in the ore-formation, and which have set their geological impression on the iron ores, are secondary or metamorphic, and have worked on a material in which the enrichment of iron has taken place in a cycle within the upper lithosphere, and where this enrichment has mainly occurred in the exogene phase of the cycle".

Landergren then refers to a diagram in the Swedish text of his paper that shows that he believes that the first step in this cycle was the development of a primary body of magma at depth, followed by its intrusion, its exogenic processing, its transport, its chemical iron (and phosphorus) enrichment, the development of an iron-rich product, and finally palingenesis. For details, the reader is referred to Landergren's 1943 paper, p. 59.

Another of the unusual aspects of the Grängesberg ore is the abundance of tin both as cassiterite and in the magnetite fraction. The average tin content of the ore is 250 g/ton in cassiterite and 89 g/ton in the magnetite fraction. Björkstedt (1972) believes that the Sn^{+4} substitutes for Fe^{+2} in the magnetite. If this is so, electrical balance must be achieved by the production of one empty cation space for each Sn^{+4} introduced.

LÅNGBAN, VÄRMLAND

Early Middle Precambrian	Manganese, Iron as Magnetite, Hematite	Hypothermal-2

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Notes

The iron-manganese deposit at Långban is located 18 km NNE of the town of Filipstad (57°45'N, 14°10'E) in an area that contains at least six similar but much smaller and essentially uneconomic mineralized rock volumes. These satellite deposits and Långban are located within a circle having a radius of 30 km. These deposits are abberations in the much larger iron ore district of central Sweden and almost certainly had an appreciably different genesis. Over 1000 iron mines have been worked at one time or another in this Bergslagen district; most of them are now exhausted or dormant.

The total production of manganese ore from Långban appears to have been nearly 3.15 million tons between 1925 and 1959 (Grip, 1978). Grip does not give an average grade for the ore but says that the best-grade ore ran 39 per cent Mn and the second-grade 28 per cent MN; how the total prediction was divided between these two grades he does not say. The iron ore, the tonnage of which Grip lumps with other mines in the Filipstad area, averaged about 60 per cent Fe. Actual mining at Långban ceased in the early 1950s, but manganese ore appears to have been shipped from Långban into the mid 1950s.

At least 135 minerals have been identified from the Långban mine of which at least 35 have been found nowhere else in the world; undoubtedly many more remain to be identified in the large tonnages of Långban specimens that are available for future study.

The rocks in the area that includes and surrounds Långban are mainly supracrustal and are of middle Precambrian age and consist of a wide variety of metamorphosed rocks of both sedimentary and volcanic origin. These are now slates, graywackes, banded hälleflintas, leptites, limestones, dolomites, and spilitic greenstones. In the immediate mine area, the principal rock type is dolomite, of which a considerable portion has been converted to skarn or skarn-rich limestone. The leptites, which constitute a minor fraction of the dolomite-leptite complex, are recrystallized rocks, probably largely derived from volcanic flows and ask tuffs, in which the average grain size is between 0.03 and 0.05 mm. Rocks of similar origin but of finer-grade are known as hälleflintas; whereas the coarser leptites grade upward into leptite gneisses. Some of the leptites initially may have been true sediments, but the material of which they are composed probably was largely of volcanic origin.

The dolomites that have not been converted to skarn are now essentially marbles. They contain widely spaced beds of leptite, but the total mine-area rock must be 75 per cent dolomite or skarn. These dolomitic marbles are the host rocks of the ore, and the mine-containing outcrop of this rock on the surface has a maximum length of about 2.5 km and its greatest width is about 0.95 km. Whereas, in the usual rock section of Bergslagen leptite, the dolomite lenses and beds are a minor fraction of the total, the reverse is true in the Långban mine area. Most of the Långban dolomite is located on a peninsula between Hyttsjön (NW) and Långbansjön (SE). Nor does there appear to be any evidence that the iron and manganese ores originally were deposited as layers or lenses in the dolomite or in leptite. This lack of stratigraphic connection between the ores and the dolomite strongly suggests that the ore material was introduced into the dolomite after it had undergone at least some of the metamorphic effects to which it had been subjected.

The contact, of the ores with the dolomite are gradational, and the entire dolomite body contains at least small amounts of disseminated ore and skarn minerals.

The shape of the dolomite body is that of a nail with the head to the south and the point being narrowest at its northern end. The dolomite is bordered (where its outcrop is not on the shores of the peninsula on which it lies) by potassic leptite on the east and south and on the west by the Hyttsjö granite. Islands of dolomite in the Långsban Lake beyond the south end of the peninsula suggest that the actual length of the dolomite body is appreciably greater than its length on the peninsula would indicate. On at least one of the islands, appreciable showings of ore are recorded by Magnusson on his map on page 6 of his 1930 paper, thus demonstrating that the mineralization of the dolomite continued farther south than the land area of the peninsula.

Of the metamorphosed rocks in the Långban area, the slates, graywackes, hälleflintas, and spilitic greenstones form the upper section of the supracrustal rocks, whereas, the leptites, limestones, and dolomites make up the lower part. These two rock groups are separated by a thick bed of intrusive greenstone believed to have been genetically connected with the spilitic lavas.

The Långban ore deposit contains separate and distinct deposits of manganese and iron oxides that normally immediately abut, the one against another. In only rare instances, however, do the iron ores include more than 2 per cent Mn and the manganese ores normally have less than 2 per cent of iron. The possible reasons for this situation are considered later in this discussion.

Into this suite of metamorphosed rocks, which may or may not have contained the ores at that time, four granites were successively intruded. The first of these igneous rocks was the Horrsjö granite, of middle Precambrian age, and the heat given off by this intrusion is considered by Magnusson (1930) to have been the main cause of the metamorphism of the older sediments. A period of strong folding accompanied this granite, resulting in the rocks of the upper portion of the supracrustals being preserved only in the synclines. After the folding was completed, three younger granites were intruded into the metamorphics and the Horrsjö mass, causing further thermal metamorphism of a more varied character than that due to the Horrsjö and more closely connected with the granite-metamorphosed sediment contacts. Of the three younger granites, the Hyttsjö is the oldest and probably is simply a finer-grained earlier intrusion from the same magma chamber from which later came the younger and coarser Filipstad and Järna granites. In addition to these silicic rocks, the area also contains quartz monzonites, quartz diorites, and gabbros, all of essentially the same age as the Hyttsjö granite and all collectively designated as Hyttsjö diorites. In the period between the Horrsjö intrusion and that of the younger granites, greenstone dikes were emplaced that appear to have been genetically connected with the Hyttsjö diorites. The final igneous activity in the area, as is true of so many mining districts, was the introduction of several diabase dikes.

In the Långban deposits, the ores consist of rounded, somewhat elongated, and distinct bodies that dip westwardly from the surface from 25° to 50°. Two separate systems can be recognized, the upper and flatter-dipping one lying to the west of the lower and steeper one. The two systems are some 60 m apart at the surface, join at the 120 meter level, and diverge from that elevation downward. The bottom of the known ore is about 250 m beneath the surface. The deposits contain both iron and manganese ore, but the two types are clearly separate, the one from the other, though the two types in many places are in contact with each other. The ore bodies are located in dolomite beds; although the dolomite both in and out of the ore bodies is interbedded with many leptite layers, the leptite does not contain ore, even when surrounded by it. The leptites in and near the ore, however, are altered to mica-rich "sköl" rock (Magnusson, 1930).

The Långban iron ores, in general, are quite similar to the skarn-type iron-ore deposits (non-manganese-bearing) in the general Filipstad area and contain magnetite and hematite as the principal ore minerals, the latter making up about four-fifths of the ore bodies. The hematite normally is associated with quartz that is, in many places, highly ferruginous but which ranges from bright red to gray as the iron content decreases. Between the hematite and the surrounding dolomite lie the magnetite concentrations with which are intermixed large quantities of skarn silicates. These silicates usually are a brown garnet [andradite - $Ca_3Fe_2Si_3O_{12}$], light green diopside, actinolite or tremolite, and biotite. The garnet, pyroxene, and the amphiboles always are idiomorphic toward the magnetite. Neither manganese minerals nor appreciable manganese in the skarn minerals occur in the iron deposits, except near the margins where brecciated areas in the hematite may contain skarn minerals rich in manganese (Magnusson, 1930).

The main manganese ore minerals are braunite - $3Mn_2O_3\cdot(Mn,Si)O_3$ - and hausmannite - $Mn^2Mn^3{}_2O_4$, the manganese counterpart of magnetite. Their location in the manganese type of deposit is analogous to that of hematite and magnetite in the iron ore bodies; braunite occupies their central portions, and hausmannite is in the margins of the larger deposits and may make up all of the smaller ones. At the contacts between masses of these two manganese minerals, the hausmannite occurs as crystals in a matrix of braunite, while outward from the massive hausmannite material individual hausmannite crystals are found without any evidence that braunite ever occurred in these areas. Bixbyite - Mn_2O_3 - although found at Långban, is probably much rarer than braunite but occurs under similar circumstances. It apparently is more readily converted to hausmannite than braunite and, because of the difficulty of distinguishing it from braunite, may be more abundant than is at present thought (Mason, 1943). Occasionally, manganosite - MnO - appears in the outer parts of the hausmannite ore and in carbonate rock beyond the ore rich in manganese. In some instances, cores of manganosite are found in hausmannite. Pyrochroite - $Mn(OH)_2$ - is also to be found, but its relation to hausmannite and manganosite is uncertain. Also associated with hausmannite are minor amounts of two other spinels, jakobsite - $MnFe_2O_4$ - and manganese magnetite which is intermediate in composition between jakobsite and magnetite.

The manganese ores are, in the same manner as the iron ores, accompanied by skarn masses; and the relation of the manganese minerals to the skarn silicates is much the same as that of ore minerals to skarn in the iron-rich bodies. In the manganese ores, the skarn minerals probably obtained some of their silica from the silica in the braunite, but the bulk of it must have come from the fluids which altered the braunite to hausmannite with a considerable local assist from the silica of the leptite layers.

Braunite is not even today a well understood mineral. Mason and Roy believe it to be bixbyite - Mn_2O_3 - with MnO (manganosite) and silica in solid solution in it. This is, of course, a very strange association because it amounts to $Mn^2Si^4O_3$ substituting for one Mn_2O_3 in a total of four Mn_2O_3 to give $Mn^2Mn^3{}_6SiO_{12}$. The structure of braunite, however, closely resembles that of bixbyite which, in turn, is closely related to that of CaF_2 but which differs from it through removal of one-quarter of the anions and a slight rearrangement of the remaining ions (Roy, 1958). This structure apparently is sufficiently tolerant of large and small cations alike to admit Mn^{+2} and Si^{+4} as limited substitutions for Mn^{+3}. The bixbyite-braunite structure is not isomorphous with that of alpha hematite or corundum.

The skarn minerals associated with the manganese ore minerals include a yellow garnet (a mixture of andradite and the iron-manganese garnet) and a minor amount of a red garnet (mainly spessartite - $Mn_3Al_2Si_3O_{12}$). The yellow garnets occur in the actual manganese ores and in hematite ores near manganese ores; the red garnets are found near leptite sköls and in calcite lenses in the manganese ores.

Monoclinic pyroxene in the manganese portions of the Långban deposit is represented by two groups, the schefferites and the urbanites. Schefferite includes both salite and ferrosalite and ranges between $Ca_4Mg_3Fe(Si_2O_6)_4$ and $Ca_4MgFe_3(Si_2O_6)_4$ with Mn^{+2} proxying for part of the Fe^{+2}; urbanite is an aegirine-augite without any Al^{+3} and rich in Na^{+1} and Fe^{+3} - $Ca(Mg,Fe,Mn)NaFe(Si_2O_6)_2$. The schefferites are more common in the manganese ores, while the urbanites are more abundant in the hematite ore near manganese bodies than in the manganese ore itself.

Rhodonite - $(Mn^2,Ca,Fe^2)SiO_3$ - is less common than the pyroxenes; and, although most of the rhodonite contains some calcium, it is not enough for more than a small percentage of this silicate to be classed as bustamite - $MnCaSi_2O_6$.

Most of the manganese amphiboles at Långban are of the richterite variety - $Na_2CaMg_5\text{-}(OH)_2Si_8O_{22}$ - that contains Mn^{+2}, apparently proxying for both Ca^{+2} and Mg^{+2}. As the urbanites contain more sodium than the usual pyroxene, so the Långban richterites contain more of that element than the ordinary amphibole, but they also have more potassium than the normal mineral of this group.

The mica in the Långban ores is in greatest amount near the mica-rich sköl rocks (metamorphosed leptite beds or fragments). Nevertheless the mica in the iron ores is biotite and that in the manganese ores is manganophyllite, indicating a definite dependence in mica composition on the ore as well as the leptite. The leptite may have supplied much or most of the aluminum for the sköl mica, but the process of ore formation appears to have been responsible for the distinguishing cation. There may be small amounts of manganese chlorites with the mica.

The most important olivine mineral at Långban is tephroite - Mn_2SiO_4 - which does not contain zinc as does the tephroite at Franklin, New Jersey. All Långban olivines appear to contain some MgO; some are so rich in it as to be called picro-tephroites. Olivines are far less common in the iron ore bodies than are those containing manganese.

In addition to these relatively normal variants of well-known silicates, there are a number of lead silicates found in the deposits. Of these, three are fairly common; kentrolite - $Pb_2Mn^3_2Si_2O_9$ - and melanotekite - $Pb_2Fe^3_2Si_2O_9$ - are ordinarily located with the braunite ores, and ganomalite - $Pb_6Ca_4Si_6O_{21}(OH)_2$ - is associated with hausmannite ore. Kentrolite and melanotektite bear the same relations to braunite as do the manganese silicates, while ganomalite usually appears as scattered grains in hausmannite-bearing dolomite and shows less definite age relations to the hausmannite than the other two do to braunite.

Another unusual silicate is långbanite $(Mn^2,Ca)_4(Mn^3,Fe^3)_9SbSi_2O_{24}$. The mineral is found in the compact skarns of the manganese deposits.

Antimony at Långban is much less abundant than arsenic; but, although the main antimony mineral is a silicate (the others are antimonates), all the arsenic minerals are arsenates. The most common of these is hedyphane - $(Pb,Ca)_5Cl(AsO_4)_3$ - while the other two important arsenic minerals (berzeliite and caryinite) are lead- and manganese-bearing arsenates. These three minerals all belong to the compact skarns of the manganese ores, but caryinite appears to be older than the other two arsenic minerals (Magnusson, 1930).

Later than the minerals of the compact skarns are those which occur in vugs and fissures. There is, however, no great age difference among the three categories because skarn minerals are found among the vug minerals and these latter (though not those also occurring in the skarns) are found in the fissures. The minerals appearing in vugs in the manganese ores often are found in fissures in the hematite ore. The most abundant vug and fissure minerals at Långban are calcite and barite. There are at Långban four beryllium-bearing minerals and two containing boron; these belong to the older part of the vug mineral period. Barylite - $BaBe_2Si_2O_7$ - is probably the most common beryllium mineral; and, although both boron minerals are rare, hyalotekite - $(Pb,Ca,Ba)_4BSi_6O_{17}(F,OH)$ - is probably the more common. Barysilite - $Pb_8Mn(Si_2O_7)_3$ - an uncommon lead silicate is also an early vug mineral as are pectolite - $NaCaMnSi_3O_8OH$ - and another lead silicate, margarosanite - $Ca_2PbSi_3O_9$. Nasonite, the Cl-end member of the ganomalite solid solution series, is also a lead-bearing silicate and an early vug mineral.

Later vug minerals include thaumasite - $Ca_3Si(OH)_6(CO_3)(SO_4)\cdot 12H_2O$ - apophyllite - $KCa_4Si_8O_{20}(F,OH)\cdot 8H_2O$ - fluorite - CaF_2, and scheelite - $CaWO_4$.

There is a huge variety of minerals in the fissure-filling group; these, in addition to calcite and barite, include other carbonates and sulfates, hydrous carbonates and sulfates, hydrated minerals rich in Mn^{+3} and Fe^{+3}, hydrous silicates, hydrated arsenates, fluoride-bearing arsenates, arsenites, chlorine-bearing arsenites, a chlorine-bearing antimonate, and a hydrous vanadate. In short, the peculiar suite of minerals in the Långban fissures is as unusual as those of the compact skarns and the vugs, particularly those of the manganese ores.

Although several sulfide minerals in minor amounts are known at Långban, they are younger than the massive ore-skarn complexes but occur both in and outside the ore bodies among the late minerals in the fissures. Where they are found in the ore, the skarn is often altered and new skarn minerals such as diopside, tremolite, phlogopite, penninite, clinohumite, and spinel are developed. The association of sulfides and these skarn minerals is a common one in the sulfide deposits of central Sweden and suggests a minor and economically unimportant later sulfide mineralization following close to the end of the formation of the iron and manganese oxide ores. Of the sulfides, bornite and chalcocite appear only in or near the manganese ore bodies.

Associated with those late sulfides found in the manganese ores are four native metals - lead, copper, silver, and bismuth - of which lead is the most abundant. The occurrence of considerable quantities of sulfate minerals in these same fissures would suggest the fate of the sulfur which must have been oxidized to produce the native metals. That neither the Cu^{+1}-bearing sulfides nor the native metals are found in the iron ores indicates that the highly negative oxidation-reduction potential of the $Mn^{+2} = Mn^{+3} + e^{-1}$ half reaction is necessary to carry the reduction to such extreme lengths (Magnusson, 1930).

In the area around Långban, the country rock is very nearly theoretical dolomite in composition; but in the dolomite of the ore bodies, particularly in the manganese deposits, magnesium is far less abundant than calcium, the latter being twice the former in the manganese ores and somewhat less in the iron bodies. In the iron ores, the actual amount of unreplaced carbonate is small, while it is much more common in the manganese deposits where the ore is much more disseminated in character. In these manganese ores, a considerable portion of the carbonate is, or approaches, rhodocrosite in composition.

There have been two well-supported explanations put forward for the origin of the Långban ore. Magnusson, who has done the most exacting and extensive geologic work on the deposits, considers that the ores were formed in several stages, the first of which was the metasomatic emplacement of hematite and braunite in a near surface environment by "volcanic after action". This was followed by a thermal metamorphism of the deposit that appreciably changed the ore minerals and also produced the bulk of the wall-rock alteration (skarn) minerals. In what were probably the late stages of the thermal metamorphism, a series of vug and fissure minerals was developed that almost certainly had a hydrothermal origin and that included, in addition to a large number of oxy-salt and sulfide minerals and a few native metals, what Magnusson considers to have been a second generation of the skarn minerals (Magnusson, 1930).

The other widely supported mode of origin suggests that the ores are definitely of the contact metamorphic or pyrometasomatic type. The iron ore bodies are nearly normal specimens of this variety of ore deposit; but, in the manganese bodies, the ore mineral character diverges quite widely from the norm. Of the Långban ore minerals, magnetite and hausmannite are spinels; hematite, an oxide; and braunite, a peculiar material which has a modified fluorite structure. These and similar oxides are common ore constituents of contact-metamorphic deposits. At Långban, the skarn minerals include garnets, olivines, pyroxenes, amphiboles, micas, and a number of other less common types (such as melilites) that are all found in well-attested contact deposits. The differences in actual species of minerals between Långban and the usual deposit of the pyrometasomatic type are thought to have resulted from the unusual history and character of the ore-forming fluid rather than from a variation in mode of origin.

Magnusson suggests that the sharp separation between iron and manganese ores at Långban can be explained only through the assumption of a large supply of oxygen (high partial pressure of O_2).* He believes that oxygen in sufficient quantity could have been available only at the earth's surface; therefore, the ores must have been emplaced in the dolomites while they were near the surface. It is his opinion that the site of deposition was so near the surface that the most probable source of the ore materials was what he calls "volcanic after action" (i.e., gases given off by the same source as that of the porphyries, tuffs, and ash from which, in turn, the leptites were derived). Presumably these gases were able to mingle with atmospheric oxygen prior to depositing metasomatically their load as Fe_2O_3 and SiO_2 and Mn_2O_3, $MnCO_3$, and SiO_2. Just how the debris of replacement, the $CaMg(CO_3)_2$ of the dolomite, was removed in the gaseous state is not explained. There would be

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* The affinity of manganese for oxygen is less than that of iron for that element; thus, in an oxygen-rich environment, the large percentage of iron in solutions containing that element and manganese will be precipitated before the latter begins to come out of the solution. This is true despite the more negative character of the half reaction: $Mn^{+2} = Mn^{+3}+e^{-1}$ than of that of $Fe^{+2} = Fe^{+3}+e^{-1}$.

no problem in disposing of the CO_2 portion of that mineral, but how the CaO was carried off in a gas phase is difficult to imagine.

If the original iron and manganese minerals are emplaced as Magnusson suggests, his genetic sequence involves the conversion of large parts of the hematite in the iron ores to magnetite. Also required is the development of the normal skarn silicates that now are associated with most of the iron ores. These, he thinks, are accomplished by diffusion of material out from the iron ore-quartz masses and in from the surrounding rock. At the same time, in the braunite ores, hausmannite was formed from much of the Mn_2O_3 of that mineral, while its silica and some of its Mn_2O_3 reacted, through diffusion, with the surrounding rocks to form hausmannite and the unusual manganese skarn minerals. In both instances, some of the trivalent manganese was reduced to the divalent state.

This reduction is the main difficulty with Magnusson's explanation of the problem. How, under natural conditions, can Fe^{+3} be reduced to Fe^{+2} with no external aid beyond a rise in temperature? (Mason, 1943) Gamma hematite (probably because of its defective spinel structure) can be converted to magnetite when heated in vacuo at 250°C by driving off oxygen and stimultaneously reducing Fe^{+3} to Fe^{+2}. We are not, however, dealing at Långban with gamma hematite, which is monotropic with the alpha form but with the alpha form itself that cannot be converted to magnetite in air unless heated to 1388°C (Mason, 1943). In other words, the corundum structure of alpha hematite is far more stable than the spinel structure of gamma hematite in the thermometamorphic conditions postulated by Magnusson (heating of the rocks involved to several hundred degrees Centigrade with no addition of water or other material from an outside source). Therefore, it seems impossible for hematite to be reduced to magnetite by the mechanism Magnusson suggests.

Under similar conditions, the breakdown of braunite to only hausmannite (and silica) is not possible either, as equation (I) demonstrates.

$$\text{(I)} \qquad Mn^2Mn^3{}_6Si^4O_{12} + \text{heat} = Mn^2Mn^3{}_2O_4 + SiO_2 + 2Mn^3{}_2O_3$$

The dissociation of the braunite into its potential constituents, without a concomitant reduction of the bixbyite produced, will produce twice as much bixbyite as hausmannite, which certainly did not occur at Långban.

In both types of ore bodies at Långban, the wide variety of skarn minerals is thought by Magnusson to have been produced by thermo-metamorphism alone. Few minerals of the skarn group are believed to have formed at the same time as the hematite and jasper (quartz with abundant inclusions of hematite enclosed in it) and the braunite and rhodocrosite. Even the first skarn minerals - the lead silicates, kentrolite and melanotekite; caryinite, the lead arsenate; and långbanite, the manganese-antimony silicate - are regarded as metamorphic reaction products. The remainder of the skarn minerals also are all products of the thermometamorphism, in Magnusson's opinion, and he also thinks that they developed at far higher temperatures than the earlier braunite and hematite. He believes, however, that the skarn silicates formed as replacements of carbonate minerals remaining after braunite formation and that this is indicated by the rounded inclusions of braunite in the skarn minerals. Thus, the hematite-quartz and the braunite were formed at lower temperatures than the later skarn and magnetite and hausmannite ores.

Because it is almost certain that the braunite and magnetite could not have been formed by thermal metamorphism alone, it is also probable that more than thermal metamorphism was needed to produce the skarn minerals. This is confirmed by the lack of several of the essential elements of the skarn minerals in the early oxides ores and in the dolomite and leptite as well. None of these original sediments appears to have contained the necessary lead, arsenic, antimony, barium, beryllium, boron, fluorine, or chlorine, not to mention the silica (even allowing for that present in the leptites and that introduced with the hematite and braunite) to produce the skarn minerals.

The wide difference in the main skarn minerals of the iron and manganese deposits would indicate that the iron and manganese in those skarn silicates were largely of local derivation, but the presence of the lead silicates, the arsenates and antimonates; the sulfates and the carbonates in both types of ore would suggest that these elements were introduced by the hydrothermal fluid which produced the skarn minerals.

The first main difficulty, under any scheme of origin, is to account for the separation of the primary iron and manganese ores (hematite and braunite) into largely distinct masses, close though the two types are locally to each other in the rocks. Further, as magnetite is more common than hematite in the usual hydrothermal iron deposit, reasons must also be put forward for the primary deposition of the iron entirely, and the manganese largely, in the trivalent states. Perhaps the answer to both problems lies in the same phenomenon, an unusual abundance of nascent oxygen in the ore fluid.

As I have suggested in the discussion on Kiruna and Gällivare in this volume, a possible source of this element might be the dissociation of nitrogen-oxygen complexes, with reduction of confining pressure on the fluid, to give native oxygen. The greater the amount of oxygen made available, the greater would be the conversion of iron to the ferric state and of manganese to the trivalent and even the quadrivalent state, and the more complete would be the physical separation between the two elements. Although the mechanism for the production of free oxygen was originally suggested by me as having been important in an iron-rich melt, there is no valid reason why it should not be equally possible in a hydrothermal fluid and may account not only for the primary Långban ores but also for many hydrothermal hematite deposits that were not developed by the replacement of magnetite.

For the partial conversion at Långban of the first stage hematite and braunite to magnetite and hausmannite respectively, thermal metamorphism alone appears unable to do the job; therefore, some reducing agent, hydrothermally supplied, is necessary to accomplish such conversions. Granted this dependence on hydrothermal fluids, and the need for them to account for the vug and fissure minerals and for the iron oxide and manganese oxide and carbonate minerals in the dolomite at considerable distances from the heavy ore concentrations, it appears reasonable to throw the entire process of later ore minerals and silicate skarn onto hydrothermal agencies. If hydrothermal fluids were present, as even Magnusson admits they must have been in the vug and fissure stages, and were capable of forming the entire suite of ore and skarn minerals, it seems unnecessary to invoke the aid of an inadequate process (thermal metamorphism) for any phase of the ore-forming process.

If hydrothermal solutions are thought to have accomplished both the ore and skarn-mineral development, their competency to do the job must be examined. To Magnusson, the location of hausmannite and magnetite on the borders of the ore masses (with the more oxidized material within, or completely making up the smaller bodies) would seem to indicate that braunite and hematite had been replaced by the two spinels. Although this is the reverse of the usual experience in high temperature deposits, it is possible that hematite may be reduced to magnetite and braunite to hausmannite and silica if a suitable reducing agent that can reasonably be expected to be present under natural conditions can be suggested.

In the copper deposits of Upper Michigan, the ores were deposited from what appear to have been normal sulfur-rich hydrothermal fluids in the presence of huge amounts of ferric ion in the wall rock. The result was not the usual suite of hydrotheramal copper sulfides but, instead, was much bleaching of the wall rock (Fe^{+3} reduced to Fe^{+2}), and the precipitating of native copper, with a little native sulver, a few arsenides, and occasional small amounts of chalcocite - Cu_2S. This suite of minerals certainly is indicative of a large scale reduction of copper and iron. Lacking from these deposits is nearly all the sulfur usually found in hydrothermally formed copper deposits. Inasmuch as the reduction of iron and copper by sulfur as S^{-2} is readily possible, as equation (II) shows, the absence of that latter element seems easily explained

$$\text{(II)}\quad 2Cu^{+2} + 4Fe^{+3} + S^{-2} = 2Cu^{\circ} + 4Fe^{+2} + SO_4^{-2} + 8H^{+1}$$

Manganic ion (Mn^{+3}) can readily be substituted for ferric ion with even greater efficiency in the oxidation-reduction involved. Thus, it seems reasonable that a high-temperature hydrothermal fluid, quite high in sulfur, would have been capable of reducing enough wall-rock hematite to magnetite or braunite to hausmannite or even manganosite to produce the conditions found at Långban.

If such an oxidation of sulfur had been the driving force for the reduction of the ore minerals, some trace of the reactions involved should have been left in the deposit in addition to new iron and manganese minerals. If the hydrothermal fluid were a normal one, it should have contained such elements as lead, zinc, and copper screened by sulfur or SH^{-1} ions. In the normal course of events, these sulfur-metal complexes would gradually have lost their screening anions until the neutral state (i.e. PbS) was reached or until the complex anion existing at the moment was neutralized by an appropriate free cation (i.e. $Fe^{+2} + Cu^1{}_4CuS_4{}^{-2} = FeCu^1{}_4Cu^2S_4$ = bornite) [Ridge, 1968, 1980]. In an environment, however, such as that which existed at Långban when the second generation of hydrothermal fluids entered the rock volumes already containing hematite and braunite, the S^{-2} or SH^{-1} ions removed from the sulfur complexes would be rapidly oxidized to the S^{+6} of $SO_4{}^{-2}$. This would shift the equilibrium in the direction of further removal of screening anions from metal-sulfur complexes. Once lead had been striped of all its screening sulfurs and was present in the ore fluid as Pb^{+2}, it could have been affected by one of two processes: (1) further reduction of Pb^{+2} to native lead with the additional electrons needed for this reduction coming from those released in the various stages of the oxidation of S^{-2} to S^{+6}; the probability of this process having occurred at Långban is demonstrated by the presence of native lead as the most abundant native metal at Långban, or (2) the Pb^{+2} might have been incorporated as a cation in such silicates as kentrolite, melanotecktite, nasonite, ganomalite, margarsanite, and joesmithite. Further use of Pb^{+2} as a cation is shown by its presence in lead tungstates, arsenates, vanadates, and chlorantimonates and chlor-bismuthates. The relative abundance of lead silicates and of the various lead oxysalts would seem to confirm the reasoning just given. It may then be reasonably asked why only lead is present in the Långban silicates; should there not also be silicates containing copper and zinc as well? This is a particularly pertinent question when the abundance of zinc in silicates at Franklin is remembered. The answer at Långban appears to be that the much larger size of the lead ion than that of copper or zinc means that it not only must be screened by more sulfurs than copper or zinc but also has a less firm grip on those anions than do the much smaller and equally charged Cu^{+2} and Zn^{+2}. It is probable, therefore, that when sulfur is being oxidized in a hydrothermal fluid, the lead-sulfur complex will be the most quickly and completely broken down of those normally available in quantity.

The amount of lead to be found in the Långban manganese ore probably is sufficient to account for the amount of braunite that has been reduced to hausmannite. As is shown in equation (III) and (IV), one lead-sulfur ion in dissociating to Pb^{+2} and S^{-2} will provide S^{-2} to reduce 84 braunite molecules to hausmannite.

(III) $PbS_6^{-10} = Pb^{+2} + 6S^{-2}$

(IV) $6Mn^2Mn^3_6SiO_{12} + S^{-2} = 14Mn^2Mn^3_2O_4 + 6SiO_2 + SO_4^{-2}$

On the other hand, the amount of lead in the iron ore bodies is essentially nil; this contrasts sharply with that in the manganese ores. If the explanation for the reduction of Fe^{+3} to Fe^{+2} is to be largely the same as for that of Mn^{+3} to Mn^{+2}; i.e. the removal of sulfur ions from their screening positions around the metal cations of the metal-sulfur complexes (particularly that of lead), the lack of lead in the iron ores must be accounted for. A possible answer is that, when $Fe^{+2} = Fe^{+3} + e^{-1}$ is the other half reaction involved in the oxidation of sulfur, the e.m.f. generated is appreciably less than when $Mn^{+2} = Mn^{+3} + e^{-1}$ is concerned. This difference in e.m.f. should result in less complete reduction of hematite than of braunite (as appears to have been true at Långban, where hematite is four times as abundant as magnetite) and a less complete removal of screening sulfur anions from the lead-sulfur complex. If such were the case, it is probable that almost all the lead-bearing solutions moving through iron ore would retain enough screening sulfur ions to be transported out of the general ore area unless they later encountered manganese masses. Those which escaped such contacts would have eventually deposited their loads as relatively normal sulfides in younger rocks now completely eroded away.

During the vug and fissure stages of this second wave of hydrothermal mineralization (the first wave having deposited the hematite and braunite ores), the reducing effects of the hematite and braunite on the hydrothermal solutions produced somewhat different results than during the skarn stage. By this time, there were no longer such large amounts of silicate ion for the Pb^{+2} of the dissociated metal-sulfur complexes to neutralize. The Pb^{+2}, therefore, in addition to entering in minor amounts into such uncommon silicates as barysilite, united with such available anions as the carbonate, sulfate, arsenite, antimonite, chloride, manganate, and vanadate to give rise to a large fraction of Långban's unusual minerals. Further, some of the free Pb^{+2} ion remained uncombined with any anion for a sufficient time to permit it to be reduced to the native form, while other minor amounts of lead retained screening cations long enough to be precipitated as galena. During this same fissure period, other sulfides also formed in both ore types; but, in the manganese ore, most of the Cu^{+2} of the copper-sulfur complexes was reduced to Cu^{+1} as is reflected in the presence of bornite - $FeCu^1_4Cu^2S_4$ - and chalcocite. Locally reducing conditions were extreme enough for the precipitation of native copper for reasons analogous to those effecting that of native lead. In the iron deposits, where reducing effects on sulfur were less strong, the copper deposited almost entirely in chalcopyrite - $FeCu^2S_2$.

A modern attempt to explain the complex and unusual Långban mineralogy is that of Moore (1971).

The first major problem to explain at Långban is why the iron and manganese ores are normally in immediate proximity to each other but do not overlap each other. The second is to make clear the reasons for the various iron and manganese minerals and for the valences of these two elements in them. Beyond these two problems, the item next in order is to discover why the deposit contains so many minerals unique to it and why so many of the elements are in uncommon valences (or are in the native state).

The dolomites that are the host rocks for the ores are metamorphosed limestones of which the composition is almost exactly $CaMg(CO_3)_2$, though calcium may be slightly in excess. Moore believes that the intrusives of the three younger granites provided heat to add to the energy resulting from the folding and faulting of the older rocks. The initial iron and manganese oxides, the "proto-ores", Moore considers reacted with the silicieous material in the host rocks and with the fluxing and reactive carbonates, at the high temperatures generated during metamorphism, to produce skarn silicates that surround the ore pods and show a banded texture. He also thinks that the definitely late hydrothermal veins, which were introduced after the peak of metamorphism had passed, provided many of the minerals known only at Långban and were formed at quite low temperatures.

Moore suggests that, by implication at least, the iron and manganese oxides were introduced through solutions of volcanic origin that were strongly acidic and contained highly mobile cations such as Fe^{+2} and Mn^{+2} that do not readily complex under such conditions. Carbonates, he points out, can act effectively as neutralizing agents for such volcanic solutions. As the hydrogen-ion concentration is so reduced, Fe^{+2} and Mn^{+2} form relatively insoluble complexes through chemical bonding with water molecules and hydroxyl groups and are, thereby, precipitated. Such complexing agents provide an unshared electron pair and, as he rightly says, are called ligands; other ligands of importance in the Långban ore fluids were such radicals as $(PO_4)^{-3}$, $(AsO_4)^{-3}$, and $(SO_4)^{-2}$. As carbon dioxide is released from such a volcanic solution in the process of being neutralized, the oxyhydrides of the transition elements are precipitated, almost volumetrically replacing the space produced by the dissolution of the dolomite. [This is a rather primitive approach to the mechanics of replacement, but that can be allowed to pass for the present.] He also contends that the formation of insoluble iron compounds, particularly if the ferric state is involved, tends to occur under more acid conditions than are necessary for the formation of Mn^{+2} complexes. In addition, Fe^{+2} is more readily oxidized than Mn^{+2}. Thus, these iron and manganese materials formed in separate areas before the thermal metamorphism occurred. The minerals so produced in the manganese ore bodies were such species as pyrolusite (MnO_2), manganite ($HMnO_2$), and psilomelane [$(Ba,Mn)_3(O,OH)_6Mn_8O_{16}$]. Thermal metamorphism of these minerals would have produced hausmannite and braunite. One example should be sufficient to show the type of equations put forward by Moore to explain how these earlier minerals were converted to later ones [note that Moore uses a formula for psilomelane different from that which was used in the preceding sentence and was taken from Hurlbut's 19th Edition of Dana].

$$3BaMn_{16}O_{32}(H_2O) + 3CaMn(CO_3)_2 =$$
$$16Mn_3O_4 + 3BaMg(CO_3)_2 + Ca(OH)_2 + 14O_2(gas)$$

[Moore appears to have made no effort in this equation equate volume and mass between reactants and products.] The barium in the psilomelane, Moore thinks, must have come from the carbonates associated with the manganese ores. He believes that this is proved by the presence of norsethite [$BaMg(CO_3)_2$] in the carbonate surrounding the braunite ore, although the equation just given shows norsethite as being a product and not a reactant. It is true, however, that the equation given above does contain about equal weight percents of barium on the two sides of the equation, but what this proves certainly is not clear. Further calcium hydroxide has not been reported from the Långban ores, so it must have been eliminated from them by some further reaction not given by Moore; nor does he say what is the role of the $14O_2$ given off by the reaction.

Since psilomelane and hollandite [$Ba(Mn^4Mn^2)_8O_{16}$] both usually contain both barium and some lead, the two elements must have been enriched in the skarn carbonates by such a reaction as that of Moore quoted above because the higher-temperature oxide structures such as hausmannite and braunite cannot accomodate these large cations.

Moore also points out that the three stable valence states of manganese- Mn^2, Mn^3, and Mn^4 - and the coordination numbers of that element that range from four to eight in respect to oxygen give manganese a much greater adaptability to various crystal structures than is true of iron. Moore would get the Sb^5 and As^5 in the Långban minerals bearing these cations from having been precipitated in the proto-ore in subordinate amounts. Later, during the thermal metamorphism and subsequent hydrothermal activity, these cations would end up in the unusual minerals in which they are found in the veins and fissures. Moore also eliminates the sulfides through intense thermal metamorphism [a fixation he shares with Magnusson], and the cations left when the sulfur departed would migrate outward to react with the silicate-rich enclosing masses of skarn to produce the unusual lead silicates, arsenates, arsenosilicates, antimonates, and antimoniosilicates known from the late stages of Långban deposition. He says that the Långban deposits could not have originated from thermal (and probably hydrothermal) metamorphism of a massive sulfide deposit because of the lack of copper and zinc in the Långban minerals.

This brief summary may not be fair to Moore's concept, but his detailed paper should be studied by anyone wishing to know more and more thoroughly about his ideas.

The most recent attempt to explain the genesis of the Långban deposit is that of Baström and his colleagues (1979). These authors, as also is true of Magnusson and Moore, are obsessed with the idea that oxidation of iron and manganese in the Långban deposit in particular can be accomplished only by the action of an abundance of oxygen supplied from sources exogenous to the original (in their opinion) sediments. They forget that oxidation is simply a process of the exchange of electrons between the material being oxidized and that being reduced. For example, to oxidize manganese or iron from the Mn^{+2} state to the Mn^{+3} requires the presence among the reactants of an ion that will give up the needed electron to convert Mn^{+3} to Mn^{+2}. Thus, as equation (V) shows, the oxidation of one S^{-2} will provide for the conversion

$$(V) \quad 12Mn_2O_3 + S^{-2} = 8Mn^2Mn_2{}^3O_4 + SO_4{}^{-2}$$

12 bixbyites to 8 hausmannites without the need for any nascent oxygen at all. I have suggested, however, that the break-up of nitrogen-oxygen compounds in the ore fluid as it rose toward the surface would supply such oxygen (see discussion on Kiruna-Gällivare in this volume).

These possibilities for the reduction of manganic to manganous ion require no help from surface-provided oxygen and remove, thereby, the need for all the elaborate series of reaction proposed by Boström and his colleagues (1979) to achieve the conversion of bixbyite to hausmannite. Remarkably, Boström and his colleagues give an equation on page 1005 of the 1979 paper that is essentially the same as that of equation (V) above, but they do not draw the same rather broad conclusions from it that I do. I can see no reason for not applying this same reaction to the production of the early hausmannite, although I am perfectly willing to consider the introduction nascent oxygen from nitrogen-oxygen compounds as well. Boström and his colleagues do not believe that metasomatism played any part in the production of the Långban ores but think, instead, that the "apparently metasomatic features can be explained by postdepositional metamorphism of a sedimentary deposit, including various redox processes". This complex concept, not thoroughly explained by any means, certainly is far more difficult to fit to the Långban environment than that of equation (V) or the break-up of nitrogen-oxygen compounds. Their use of the abundances of uranium and thorium and of their isotopes in their study shows that the Långban ores are not untransported chemical weathering products, nor do they think, they were formed from manganese nodules. They are convinced, however, that they could have been formed volcanic-exhalative processes because they show U/Th ratios that resemble those of recent exhalative-sedimentary deposits. Because these ratios are not completely conclusive, even to them, they suggest that uranium is much more mobile than thorium. Thus, during the numerous metamorphic processes to which the Långban ores have been subjected in the billions of years since these deposits were laid down, the Långban ratios of uranium to thorium may have changed enough to meet their theories. I cannot but think that this reasoning rests on a rather tenuous base. They list a number of deposits that they conclude must have been formed in the same exhalative-sedimentary manner as they apply to Långban; one of these is Franklin, New Jersey. For an alternative explanation for the formation of the Franklin deposit, I refer them to my 1952 paper on that deposit.

Considering the variety of opinions expressed on the manner in which the Långban deposit was formed, I have considered the matter carefully and still believe that it is a rather unusual, but still definitely a hypothermal deposit in calcareous rocks, that is, Hypothermal-2. There appears to be no question but that the deposits were developed in the early middle Precambrian (i.e. prior to 1800 m.y. ago) perhaps, under some of the suggested mechanisms of formation, the formative process may have extended beyond the upper limits of the early middle Precambrian, but I think that early middle Precambrian is the most satisfactory age to apply to the Långban ores.

FALUN-GARPENBERG, KOPPARBERG

Middle Precambrian	Copper, Zinc, Lead, Silver	Falun - Hypothermal-1, -2 Metamorphic-C Garpenberg - Hypothermal-1, -2, Mesothermal (?)

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Notes

The mine at Falun (60°37'N, 15°40'E) is 45 km northeast of that at Garpenberg; although both lie in belts of leptite, these belts are completely separate. The belts are elongated in much the same directions (Falun - nearly east-west; Garpenberg northeast-southwest). Both belts are part of a zone of leptite that (with other supracrustal rocks) form a broken circle around a synorogenic granite of Svecofennian age. The circle is nearly complete from Svärdsjö (some 40 km northeast of Falun) the vicinity of Orebro (59°17'N, 15°13'E) but becomes much more patchy around the rest of the circle. Several leptite masses, such as that in which Uppsala is located or the one that contains Sala, are completely surrounded by Svecofennian granites.

The Garpenberg leptite mass is nearly completely enclosed by the synorogenic granite, but on the north is bordered by post-orogenic, palingenic, younger Svecofennian granites. Further, on the southwest tip of the Garpenberg belt, the contact is between the leptite and Cambro-Silurian sediments. This belt is about 25 km long.

The Falun leptite has a total WSW-ENE length of about 35 km; its greatest width is some 8 km. The mineralized portion of the belt, including the huge ore bodies of the Falun area, extends from about 5.5 km west of Falun to beyond the Skytt and Näverberg mines and up to 6.5 km east of Falun for a total mineralized distance of about 12 km. The Falun belt is completely enclosed by Svecofennian granites with most of the surrounding granite being synorogenic, but the southwest margin is bounded by the younger palingenic granite.

Examined on a somewhat broader scale, the Falun and Garpenberg deposits are part of a sulfide belt 120 km long and 30 km wide that trends northwest to southeast and extends from Falun to Sala (59°55'N, 16°38'E). If the imagination is allowed even further sway, it is possible to continue the Falun-Sala belt southeast by another 150 km to include the sulfide deposits on Utö Island (58°56'N, 18°20'E), but this seems to be an excessive extrapolation.

In both the Falun and Garpenburg belts, limestone and dolomite are present and are intimately involved in the localization of the ore.

Falun and Garpenberg are discussed together here mainly because they are the two principal sulfide deposits of central Sweden. After each has been discussed in turn, they will be compared as to their similarities and differences.

It has been calculated that, at the time of the discovery of Falun in about 1200 A.D., the deposit contained at least 35 million tons of sulfide ore. The total area of the sulfide-bearing complex (massive) appears to have been about 40,000 m^2; the grade of the 35 million tons in this deposit probably was: 1.× per cent copper, 5.0 per cent zinc, 1.7 per cent lead, and 35 per cent sulfur. In addition, the area contained some 10,000 m^2 of outcrop area with the grade of the ore beneath it having been: 2.5 per cent copper, 0.5 per cent zinc, 0.1 per cent lead, and 10 per cent sulfur. In 1973, over 370,000 tons of ore and wall rock were removed from the deposit, of which nearly 173,000 tons were crude ore. The total amount of copper produced since the mid-13th century has been more than 400,000 tons (Grip, 1978).

The Skytt and Näverberg copper and zinc mines are some 4 km west of Falun in a westerly continuation of the mineralized belt. To dispose quickly of these two deposits, the Näverberg ore is confined to cordierite-rich rocks, whereas the Skytt mine (200 m SE of Näverberg) is in a skarn-limestone complex, the axis of which is oriented almost at right angles to the strike of the enclosing rocks. The Skytt ore is localized by a drag fold that Grip (1980) says was formed during clockwise movements around an axis that dips steeply WSW. The most abundant ore mineral at Näverberg was chalcopyrite, but sphalerite, galena, and pyrite also were quite common. The gouge material now seen on the dumps is quite like that at Falun. At the Skytt mine, the dumps show sphalerite that, to some extent, cuts into limestone and also is enclosed in tremolite or anthophyllite skarn. The Skytt ore also includes some galena, chalcopyrite, pyrite, and pyrrhotite; the principal skarn is of the tremolite variety. The high silver content apparently is contained in the sphalerite. The depth reached by the Skytt mine was 186 m; at Näverberg, it was only 40 m. The total surface area over the mines was 750 m^2; the last mining was done in 1908 (Grip, 1978).

The leptites in the 12 km Falun ore belt has been converted to mica schists and to "ore quartzite" to various degrees. These rocks were arranged parallel to an east-west schistosity zone and are themselves highly schistose. In addition to the impressive drag fold at the Skytt mine, the major structures locating ore at Falun also are strong drag folds; these folds were produced by clockwise earth movements.

In the fairly recent past subsidence has occurred over the areas of most extensive mining and several large depressions have been developed; the largest of these is the "Storstöten"; it is over 95 m deep and covers an area appreciably larger than the area of the original outcrop, that having been estimated to have been about 50,000 m^2 in area.

The terms for the host rocks at Falun as used by Koark (1960) are somewhat different than those employed by Grip. Koark says that the ores are mainly in "camgites" and quartz schists. The camgites are rocks that originally were limestones or dolomites that have had their magnesium content provided or increased by metasomatic reactions and are now largely Mg- marble. The magnesium silicates that they now contain probably were developed without the further addition of magnesium (assuming that the material metamorphosed and metasomatized was originally dolomite). Principal skarn minerals (crystalloblasts) consist of forsterite and humite group minerals (chondrodite and norbergite); further into the alteration process (hydration), these minerals were converted to serpentine and ophicalcite. Such a sequence probably required more SiO_2 than the carbonate rock initially contained. This suggests that this material must have been added from some outside source, although Koark (1960) appears to believe that the degree of skarn formation was controlled by the SiO_2 in the original carbonate rock as well as by the initial supply of Mg^{+2} and Fe^{+2} ions in the carbonates. The later development of Ca-silicates probably took place either after all the original Mg^{+2} and Fe^{+2} had been used up in the formation of skarn silicates or after no more of these two ions was being added from outside sources. These Ca-bearing silicates consist of tremolite-actinolite and diopside-hedenbergite with some grossularite garnet. Phlogopite, chlorite, and spinel also may be present.

The tectonically metamorphosed leptites were converted to mica schists, the main process being the sericitization of feldspar. With appreciable silica additions, quartz schists were developed in which took place the production of additional silicates such as cordierite, andalusite, anthophyllite, cummingtonite, almandite, sillimanite, and most abundantly biotite (usually containing zircon inclusions). In those leptites that were converted no farther than mica schists, only cordierite and sillimanite were produced as post-deformation minerals, in many places occurring as porphyroblasts. Eskola and Geijer favored an external source for the iron and magnesium; others believed that the primary carbonates were sufficiently provided with these two ions and with silica as to need no additions from external sources. Schermerhorn's recent work (1978), however, suggests that these skarn rocks are simply the more metamorphosed equivalents of chlorite and jasper that accompany the pyrite and other sulfides in many Iberian massive sulfide deposits. It is Schermerhorn's opinion that the Falun ore body was autochthonously deposited on the top of a mineralized feeder zone with a considerable content of chlorite and silica. Metamorphism converted the sulfide-bearing material into a chalcopyrite-, anthophyllite-, cordierite-bearing quartzite. He considers that the magnesium autometasomatic process also was responsible for the skarns developed at Orijärvi. The problem does not appear finally to have been solved, but I suspect that Eskola knew more about Orijärvi and Geijer more about Falun than Schermerhorn does about either.

Koark says (1960) that not all quartz schists in the Falun area seem to have been produced by silicification. Some of these schists he believes have a sedimentary structure; in fact, he sees in the debris that has fallen into the "Storstöten" blocks that contain quartz-banded iron ore, in the stratification of which pyrite and sphalerite also are present. This may be argued to show that the original, pre-metamorphism ore was sedimentarily produced, but it also may do no more than indicate that zinc and iron sulfides were introduced into what almost certainly was a sedimentary type of banded iron ore.

The ore minerals are pyrite, pyrrhotite, chalcopyrite, sphalerite, and galena; these range in quantities in the Falun rocks from accessories to almost solid masses of sulfides. The galena and sphalerite most commonly are present in the Ca-Mg skarns and marbles; the chalcopyrite is mostly in the skarn rocks, the quartz schists, and in biotite-, chlorite-rich gouge rocks; pyrite is almost universally developed.

Grip reports that Koark states that there are three types of ore in the Falun mine: (1) complex; (2) disseminated, and (3) sköl or gouge. Actually, Koark gives three designations in Swedish, which are, in the order just given: (1) blötmalm or soft ore; (2) hårdmalm or hard ore, and (3) skölmalm or gouge ore. The complex ore is mainly massive pyrite that includes considerable amounts of chalcopyrite, sphalerite, and galena. The disseminated ore is made up of impregnations and veinlets of chalcopyrite in the so-called ore quartzite; this ore type also contains appreciable gold, being in portions of the eastern part of the mine quite rich in the noble metal. The gouge ore is composed of highly altered leptite (mica schist) that has been converted largely to chlorite and biotite that contains disseminations and veinlets of chalcopyrite or pyrite, pyrrhotite, sphalerite, and galena.

The major part of the complex ore is in a major drag fold that plunges to the south at about 70°. In the southern part of this ore body, the ore contains relict patches of limestone and dolomite, in many places altered to ophicalcite [a rock commonly formed by dedolomitization of siliceous Ca-Mg carbonate-rich rock making a clouded green serpentine-rich rock, this ophicalcite locally is veined by white or pale green (verde antique)], particularly in southern apophyses of the main ore body. The small concentrations of magnetite found in the complex ore are considered by Geijer to have been derived from the reduction of primary sedimentary hematite during the metamorphic process. The main gangue mineral in the complex ore, however, is quartz that, in numerous places, occurs in sizeable segrations. Typical skarn silicates also are abundantly present.

The richest of the complex ore bodies in copper normally are at the margins of the deposits. Surrounding these rich marginal ore volumes in most places are the gouge ore bodies, as 1 to 20 m wide aureoles rich in biotite, chlorite, talc, amphibolite, and cordierite, andalusite, or almandine. These gouge zones are surrounded by quartzite. The gouge also is developed in faults or fault zones that lie outside the confines of the large ore bodies. Grip and Geijer think that these gouge-filled faults provided routes for the entry of ore fluids into the major carbonate masses.

The major complex ore body essentially dies out at the 332 m level, but there is a small body of ore to the southeast that reaches down for about several 10s of meters more until it is cut off by a dike of quartz porphyry that is bordered by amphibolite. Two even smaller ore bodies extend downward to about the 500 m level. Ore shoots that branch off to the west from the main ore body are zinc-rich and were being mined in the late 1970s at least. This ore body is known as the Storgrove.

The zone of disseminated ore (Grip, 1978) is continuous to the east and west from the main group of complex ore bodies. In these ore zones, chalcopyrite is present in thin veinlets and isolated spots; the copper mineral normally is associated with pyrite, pyrrhotite, and sphalerite, plus some magnetite and gahnite. Also in the disseminated ore are veins of white quartz that contain chalcopyrite, galena, and sphalerite. In the eastern portion of the disseminated ore, native gold is quite common and normally is associated with the unusual sulfosalt - weibullite [$Pb_5Bi_8(S,Se)_{17}$].

The ore-bearing area has been intruded by composite quartz-porphyry amphibolite or massive amphibolite dikes, some of which may be as much as 20 m or even more in width. Grip believes that the dikes normally are older than the ores, although some dikes are essentially of the same age as the ores. Both ore and dike materials have been deformed and altered by the same metamorphic processes.

According to Koark (1960) every major concentration of pyrite in the Falun mine has been designated as a pyrite stock. These are oblong lenses or cigar-shaped bodies that this author believes were initially syngenetic sedimentary ores that were tectonically driven into their present positions. Some of these pyrite-rich ores are associated with quartz schist and some with lime-marble and skarn. These ores probably are simply what Geijer and Grip consider to be disseminated ore in either of the two rock types just mentioned.

Koark says that, in the western part of the drag-fold curve of the major mine, three definite horizons can be distinguished (from footwall to hangingwall): (1) sphalerite-galena-rich ore in lime-dolomite marble (camgite); (2) compact stratified pyrite ore in quartz schists rich in anthophyllite; and (3) ordinary pyrite ore in quartz schists. The ore-types probably correspond to Geiger and Grips' types (1), (2), and (3) although the identity of Koark's type (3) with Grip's type (3) is not clear.

Koark names the two minor ore bodies southeast of the Storgruve as Källort and Drottning (K NW of D), but he does not name a third ore body in this area, so that one of the two named actually may be the "small body of ore to the southeast . . . . cut off by a dike of quartz porphyry", but it is more likely that the third mine is the Lovisa mine which is described as being "to the west of the pyrites of the large mine [and containing] pyrite ores".

The so-called hard ores (Grip's disseminated ores) are east and west of the pyrites of the large mine (Koark, 1960), They lie in the volumes most highly silicified and, therefore, most hard. The shapes of the various types of chalcopyrite in these ores are similar to those in the pyrite stocks. The proportion of pyrite in the quartz schists is, in some places, greater than that of chalcopyrite, and such ores, because of the lesser hardness and cohesion of pyrite than quartz, are known as semi-hard. Chalcopyrite is not the only ore component in the semi-hard, and apparently not even in the hard ores; pyrrhotite, sphalerite, and galena also are present.

Although the major gouge (or sköl) ores are most important between ore and country rocks, where they are as much as 20 m wide, they are also to be found within the ores. The developemnt of these gouge ores appears (in Koark's opinion) to have been developed both by differential movements and by metasomatic reactions caused by ore-bearing fluids passing through these zones of weakness, but how much of the minerals developed in these gouge ores was due to one process and how much to the other, Koark does not (or does not believe himself able to) say. He does, however, think that the calcium-magnesium marble skarns are most susceptible to the formation of the gouge minerals (mainly biotite and chlorite but also including talc, serpentine, and muscovite) with the exact products of this deformation depending on the pre-deformation proportions of cordierite, almandine, anthophyllite, tremolite, gahnite and similar silicates in the skarns. These sköls are not, of course, entirely of ore grade, but, where they are, especially in the upper parts, they are known as Knipsköl. This rock type was regarded by Geijer as the channelways by which the ore fluids reached into the rocks that they converted to complex or disseminated ore. Koark reports that the sköl bodies are dominantly aligned in a NNW-SSE direction.

Koark states that quartz porphyry dikes (some of which actually are equigranular) and amphibolites cut through and break up the rocks and ores in the Falun area. These dikes are as much as 20 m thick and are bordered by mafic selvages which, prior to Koark, were thought to be the result of crystallization of intruded magma masses. He, however, believes that he sees characteristics that point to their having been developed metasomatically (i.e. by metamorphic differentiation). Koark claims that these dikes are folded and fault to some extent, but the structures into which they are contorted cannot be related to the tectonics of either the ores or the country rocks. In places, the dikes have been somewhat altered and have been slightly impregnated with sulfides; previous workers thought this showed the ore to be younger than the dikes. The mineralogical character of the dikes and leptites is much the same, which indicates to Koark that the dikes are simply recrystallized portions of the leptites f m which mafic material have been driven out to form the mafic selvages. Grip (above) disagrees with this.

Nor does Koark believe that there is any genetic connection between the synorogenic granites and the ores. He bases this concept on the failure of a supposed dissolution front to have passed through the area (the leptites) so far as he can tell. He also doubts that the gouge bodies were channelways for the upward passage of ore fluids as has been accepted by Geijer because considerable volumes of the gouge rock do not contain ore. He bases this belief on the presence of ore in adjacent Ca-Mg marbles and quartz schists that are much better mineralized than adjacent gouge bodies.

Of course, it is quite possible [I think] that the gouge masses in many instances were no more than channelways and were, simply from the composition of the ore fluids, much less susceptible to replacement by sulfides than the adjacent carbonate and schistose rocks.

Koark is bothered by the difficulties posed by the idea that the materials in the ore bodies were generated in the granites and moved through the leptites as a "dissolution front", but this concept has long since been discarded and no longer is a problem in ore genesis.

The hypothesis that Koark presents is, briefly, that an isoclinal group of folded leptites, including, among other rock types, possible exhalative-sedimentary sulfides, were appreciably metamorphosed both structurally and mineralogically through regional metamorphism to produce the present sequence of rocks and the ore bodies they contain. Toward the end of this phase, considerable faulting took place and the quartz porphyry bodies were introduced. Further, these earth movements largely produced the gouge rocks and some of the available sulfide was incorporated in this gouge to convert it into gouge ore. He considers that this was done by silicate and ore solutions that still were circulating through the volumes of Falun rocks. These solutions deposited mainly chalcopyrite in gouge ore. He considers that such gangue minerals as cordierite, anthophyllite, and tremolite were developed at this stage, indicating a rise in P-T conditions as the deposition of ore and gangue reached its climax. Just where these solutions came from and how they were generated is not clear. Koark does say, however, that a new generation of sulfides may have been added during the regional metamorphism. If he is willing to have some of the ore minerals, at least, produced by hydrothermal solutions foreign in their place of generation, he ought to be willing to consider that all of the sulfides were so introduced. But he believes that only one minor example of sulfides so generated is known to him, the filling of a major fissure with chalcopyrite and galena.

Geiger (1964) argues strongly against Koark's ideas, Koark believes that the close association of sulfides, in some places, with what are agreed by both Geijer and Koark to have been banded iron ores of exhalative origin, means that they must have been formed at the same time. Geijer, on the other hand, is convinced that the sulfides were introduced long after the iron ore had been formed. He believes that the large-scale rock alteration by the introduction of magnesium (mainly) and sulfides, the control that structures had over the localization of the ores, and the major deposition of sulfides in carbonate rocks all indicate their formation at considerable depths and certainly not be sedimentary-exhalative processes operating on the sea floor.

Thus, the Falun ores must either be classified as Sedimentary-A3, Metamorphic-C or Hypothermal-1, -2, Metamorphic-C. I prefer the latter.

The Garpenberg deposits that contain 32 separate ore bodies (Grip, 1978) are located in a leptite belt that extends northeast for 15 km from the Daläven (river) to Garpenberg and then for 10 km beyond that town. The mines are in the township of Hedemora and are 15 km ENE of the Hedamora railroad station. The Garpenberg Odal mine was first mined (for copper) in the 14th century; how long before that the actual discovery was made apparently is not known. The property was originally designated as the Garpenberg Mining Company, and, at some stage, came under the ownership of Zinc Mines Ltd. In 1955, this company was purchased by, and became a subsidiary of, the Boliden Mining Company; this company became the Boliden Company in 1966; since 1958, Garpenberg has been an integral part of the parent company (Pilava, undated, mimeographed).

The leptite belt is a complex of silicic and intermediate volcanic rocks that contains two major horizons of carbonate rocks that are separated by what once was a greenstone but now is largely amphibolite. The strike of the belt is about NE-SW, and the dip is steeply to the south. The leptite is overlain by various arenites, and the whole assemblage of rocks has been folded into a synclinorium that is bounded on all sides by Svecofennian synorogenic granites. All of the rocks of the complex have been isoclinally folded around horizontal axes. A secondary folding is centered around steep axes that are parallel to the plunge of the ore bodies. In the vicinity of Garpenberg, the leptite complex has the form of a synclinal trough.

The Garpenberg mine proper (Garpenberg Odal) has an area of 12,000 m^2 and is in the southwestern part of the synclinal trough; Garpenberg Norra was discovered during the early 1960s and has an ore area of only about 1900 m^2; it was brought into production only in 1967. The Norra deposit is on the same stratigraphic level as the zinc ore in Garpenberg Odal. Norra is 3 km northeast of Garpenberg Odal. In 1973, just over 325,000 tons of ore were produced by both mines; almost 60 per cent of this came from Garpenberg Norra.

Many types of ore are found in the 32 Garpenberg ore bodies with the most important types being disseminated copper ore and complex sulfide ore. In the disseminated ores, chalcopyrite as the major valuable mineral; it is accompanied by smaller amounts of other sulfides. Some quartz-fluorite dikes occur in the area, and these, and the rock volumes immediately around them contain some gold. The leptites around these disseminated ores have been altered to "ore" quartzite and mica schist; this alteration having been caused by a "magnesium metasomatism" similar to that that affected the wall rocks at Falun.

The complex ore includes sphalerite and galena and lesser quantities of chalcopyrite and pyrite; this ore carries appreciable silver. This complex ore is contained in dolomitic limestone, ophiclacite, and skarn, the latter material is composed of tremolite, a mineral of the humite group, talc, serpentine, and some fluorite as well as unreplaced limestone. The leptites that surround the complex ore have been converted to mica schist and mica quartzite; the rocks contain abundant cordierite and cummingtonite and less common andalusite and gahnite. The mica schist is the dominant alteration mineral in the area, the quartzites being present only in minor masses and schlieren in the mica schists (Pilava, undated). The darker, biotite-bearing mica schist is dominant over the lighter, muscovite schists. All degrees of gradation occur between the two types of mica schist and between the schists and the quartzites.

The plagioclase content in these schists has a considerable range but normally is low. The proportion of the anorthite molecule in the feldspar in the mica schists is varied, but it always is low; the anorthite content in these schists, however, is higher than the unaltered leptite away from the alteration zones. In the zones immediately adjacent to the mica schists, however, the reverse is true, the plagioclases being richer in anorthite, apparently having received additions of calcium from CaO driven out of the volumes of mica schists and quartzites.

In both the mica schists and the quartzites, minerals characteristically present are almandite, cordierite, andalusite, and gedrite, the latter in places being replaced by cummingtonite. The dominant minerals in the gouge material are biotite, chlorite, and muscovite in varied proportions. These gouge (sköl) bodies form networks in the mica-schists and quartzites; Pilava considers that these gouge bodies acted as channelways for the sulfide-carrying solutions.

Several limestone bodies are intercalated in the leptite, and these have been altered to dolomite or ophicalcite; the dark portions of the ophicalcite usually are serpentinite, in places with relics of olivine or humite minerals. The principal skarn mineral is tremolite that in many places has been converted to talc. Along the contacts between the altered limestone and the surrounding silicate rocks, broad biotite-sköls are commonly present; these sköls connect to the network that cuts the mica schists and quartzites.

In the portions of the present Garpenberg mines that were known and mined in the 14th century; chalcopyrite in quartz and quartzite was the ore mineral and the mines were known as Overberget, Bergsmansgången, and Biskopen. In the latter mine, chalcocite and bornite also were associated with the Fe-Cu sulfide. Except for these mines in the northwestern and southermost parts of the Garpenberg mines, however, lead-zinc ores were far more important than the minor amounts of copper they contained. The ores are found mainly in the tremolite-dolomite bodies but also in the mica schists and quartzites that surround the altered carbonates and in the sköls that occur in them. The metasomatic emplacement processes, Pilava says, quite clearly belong to the latest phases of the folding. Between the copper ore in the Overberget portion of the mine and the complex-lead-zinc ores in its central portion, some levels contain small bodies of pyrite-pyrrhotite ore that include minor amounts of zinc, lead, and copper sulfides. Iron-sulfide-rich ores, however, are most rare in the Garpenberg mine.

Although the Garpenberg ore bodies in general follow the bedding, Grip considers that they also are controlled by the zones of gouge; these gouges were developed during the formation of the large drag fold in the eastern (or hanging-wall) side of the ore field. The calcareous rocks (those that contain the skarn ores) were, during the drag-fold deformation, folded around axes that dip 70°E. On the contrary, the brittle character of the leptites caused them to fracture and to be moved as blocks.

Grip recognizes a well-developed zoning within the Garpenberg deposits with zinc and lead becoming more abundant from foot- to hanging-wall, whereas copper does the reverse. Also, toward the hanging-wall, zinc increases in relation to lead. Still another change in the hanging-wall direction is the increase in deposit thickness in that direction. The mineralization did not occur in one continuous episode, but several, closely connected stages of mineral development occurred at Garpenberg.

In the rock volumes between the ore bodies are networks of mica gouges that locally contain minor accumulations of ore minerals; in places, these take the form of ball ore. These gouge zones increase in abundance toward the north, and Grip believes that they acted as channelways for the ore-forming fluids. Both the ball ore and the gold-bearing quartz veins are the latest ores to develop in the district, except for the secondary soft ore (see below).

At the beginning of the 20th century, the ores mined were no more than southern offshoots from larger ore bodies to the north. In the 1920s and 1930s, several exploratory efforts were made to find such northern continuations of the northeast-southwest-striking, northwest-dipping known ore, but, in every instance, water under intense pressure was encountered; this water was determined to come from the overlying lake (Grusjön). A dam was finally placed across the lake so that ground north of the dam could be explored and exploited. Here the huge ore bodies of the Strandgruven mine were found. Much of this ore was found to have been altered by descending surface water to "soft" ores. In these ores, the sphalerite, galena, and chalcopyrite had been largely broken down. At essentially the same time and by the same process, the dolomite, skarn, and silicate rocks had been converted to a mass of clay. This clay, in the upper levels of the mine, was colored brown due to surface oxidation, but, in the deeper levels of the mine, is gray in color.

Although X-ray and chemical analyses discovered a zinc chloride in the clay, most of the zinc is absorbed on the major mineral of the clay-kaolinite; in addition, much hemimorphite [$Zn_4Si_2O_7(OH)_2 \cdot H_2O$] was discovered along with minor zinc carbonate. This material appears to resemble the soft, zinc-rich material centrally located in the Sterling mine in the Franklin area of New Jersey. This clayey zinc ore averages over 26 per cent Zn, 0.45

per cent lead, 0.03 per cent copper, and 1 gram of silver per ton. The silica content is over 30 per cent.

In addition to the "soft" clayey ore, complex Pb-Zn-Cu-Ag ores and copper ores are important in the Garpenberg mines. The complex ores are partly softened and partly still massive. The ore minerals in the hard complex ore are sphalerite, galena, and minor chalcopyrite. These ores are associated with the series of limestone layers intercalated in the leptites. Pilava holds that these carbonates have been altered in part to skarn and in part to dolomite by the action of the high-temperature hydrothermal fluids that brought in the sulfides. The most common minerals in the skarn are talc and tremolite. The denser and richer ore concentrations are enclosed by rounded masses of gangue minerals or are present in bands or lenses in the skarn.

The weathered ore, to be distinguished definitely from the clayey material in the Strandgruvan mine, is known locally as "mullmalm" or dust ore. It is loose and earthy in its consistency. It probably was formed by the action of surface waters working their way down along crush and fissure zones to attack the primary minerals. Part of the ore-element content of the dust ore probably was added to that originally present by the descending waters. Some, it is difficult to say how much, of the dust ore probably was removed from the area by glacial erosion. In the massive ore, the silver content is lower and the copper content higher than it is in the dust ore; the lead-zinc contents of the dust ore also are higher than in the primary material. Some of the ore minerals in the primary ore have been replaced in the dust ore by carbonates, sulfates, silicates, and hydroxides.

In contrast to the complex ore, all of the copper ore is of the unaltered type; it is located mainly in quartzite and is richest where the quartzite consists of intergrowths of quartz and fluorite. Locally, the gold content of the Garpenberg copper ore is well worth recovering.

The Garpenberg area also contains iron ore in addition to the major sulfide deposits; the iron-ore horizons can be followed over large distances and are composed of skarn-carbonate ores, the southwestern portions of which are high in manganese. These ores are strata-bound, but they are in rocks lower stratigraphically than the sulfides. Nevertheless, minor amounts of iron ores are associated with the sulfides and sulfides with the iron ores. Not enough work appears to have been done on the iron ores to be able to say whether they were formed by exhalative-sedimentary process or hydrothermally.

The close association of the Garpenberg sulfides with high-temperature skarn bodies in the Garpenberg leptites strongly suggests that they were introduced at much the same time as the skarn minerals were developed and probably were deposited under hypothermal conditions. The presence of appreciable amounts of ore in the quartzites (mainly the hard copper ore) indicates that the categories "hypothermal-2" and "hypothermal-1" should be used to classify the Garpenberg ores as was done for those at Falun. The possibility exists that some of the complex ores at Garpenberg were formed in the upper part of the mesothermal range, although enough microscopic work, particularly, needs to be done before this point can be settled. The amount of metamorphism that has affected the Garpenberg ores appears to have been appreciably less than the amount acting on those at Falun, but it seems reasonable to add the category "Metamorphic-C" to the hypothermal classification already applied.

Of course, if Koark's concept of a sedimentary-exhalative manner of the formation of the Falun ores is correct, it probably also conceivably could apply to the Garpenberg ores as well. The Garpenberg ores, however, seemed to me to be less likely to justify the categorization of these ores as Sedimentary-A3 than do those of Falun. I do not, however, think that the Sedimentary-A3 classification is correct in either instance.

It would be well, before closing the discussion on Falun-Garpenberg to consider what Frietsch (1980) has to say about the manner in which the Falun deposits (and others of similar character) were formed. The first thing he does is to quote Vrana (1975), Schermerhorn (1978), and Berge (1978) who say that the magnesium-rich rocks that are associated with the sulfides at Falun [and Orijärvi] were syngenetically formed. Comparing the results of these authors with Sangster's thoughts on more or less similar deposits, particularly in Canada, Frietsch thinks that the magnesium-rich rocks could be the metamorphic equivalents of the original chloritization and silicification products of the volcanics, the process being associated with the formation of the sulfides. Frietsch, therefore, thinks that the magmatic hypothesis of Geijer and Eskola to explain the development of magnesia-rich rocks should be considered obsolete.

Frietsch, however, goes on to say that it is necessary to postulate a complex history to explain the formation of magnesium-rich rocks and their associated sulfides, a history that would require that the various necessary processes to have acted at different times. He first remarks that the connection between magnesium-rich rock and associated sulfides on the one hand, and the early Svecokarelian granites on the other is uncertain. He knows of few geologic data that support the idea that these events took place over much the same period of geologic time. He knows, however, of evidence that the Mg-rich rocks and the sulfides are at least partly younger than the granites and the intraorogenic mafic dikes that cut through the granites. He continues that, even if it cannot be ruled out the the Mg-rich rocks are metamorphosed formations that were formed paracontemporaneously with the volcanics and carbonate rocks in which they occur, he thinks it certain that the Mg-rich assemblages must have been at least partly formed by solutions that attacked the older rocks. The replacements in question definitely moved outward from fissures in these much older rocks in various degrees of intensity. Such conditions certainly do not conform to the idea that all of the sulfide and the necessary iron and magnesium were present in the rocks from the time when the sediments and volcanics were laid down and only assumed their present relations to each other under conditions of considerable metamorphism.

Frietsch points out that it is definite that the formation of Mg-rich silicates is, at least in some places, post deformation. Certain intraorogenic mafic dikes show an Mg-metasomatic alteration that, at least in part, if not entirely, is directly related to faults and joints. Further, the newly formed minerals lack a preferred orientation to existing planes of schistosity and lineation. Frietsch considers that, in several sulfide deposits in central Sweden, the formation of sulfide ore and tectonic activity were simultaneous, and he accepts the concept (as Koark does not) that sköls often act as ore feeders. He names several deposits in which such a relationship certainly is true; these include Svartik where sköls and breccia zones connect different ore bodies that contain sulfides; Sala where sköls that are parallel to the ore bodies continue directly into the early Svecokarelian granite as schistosity zones rich in sericite; Vallberget where the ore is related to a fissure zone cutting the bedding of the volcanic-sedimentary complex; and Hällefors where numerous veins in the western part of the mine consist of galena and sphalerite.

Still further, Frietch mentions the sulfide ores that are part fissure fillings (that are related to tectonic movements) and also contain so-called Ball ores at Garpenberg, Saxberget, Ryllshyttan, Vallberget, and Floberget. These ball ores are massive and are made up of a sulfide matric with numerous, locally spherical inclusions of different rocks and minerals [Geijer and Magnusson (1952) and Geijer (1971)]. These ball ores are believed to have formed during intense tectonic movements along faults and other dislocations. Both Magnusson (1960) and Geijer (1971) think that these ball ores were developed by the same process that gave rise to the bulk of the sulfides occurring in central Sweden and that was the last phase of one essentially continuous process of ore formation.

Frietsch (1980) thinks that the arguments summarized immediately above make it clear that: (1) the sulfides and the associated Mg-rich silicates of central Sweden are at a minimum partly later than the volcanic sedimentary complex as well as later than the early Svecokarelian granites and mafic dikes and (2) the formation of the ores is definitely connected with, in many instances, the tectonic zones.

All this largely negates Frietsch's statement, quoted above, that the Geijer-Magnusson hypothesis to explain the formation of magnesium-rich rocks should be considered obsolete. Rather, he seems to have done much to rehabilitate it into good scientific standing, so that my conclusion that the Falun-Garpenberg ores should be considered reasonably classed as hypothermal-1, -2 probably is correct.

ÅMMEBERG, NÄRKE AND ÖSTERGÖTLAND

Middle Precambrian	Zinc, Lead, Silver (?)	Hypothermal-2 Metamorphic-C

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Notes

The Åmmeberg zinc-lead deposits is located in the commune of Zinkgruvan on the boundary between provinces of Närke (south) and Ostergötland (north). The town of Ammeberg is 9 km northwest of the mine site, the latter being at 58°49'N, 15°06'E. The first knowledge of the district probably was obtained in the 16th century, but mining as a major operation did not commence until 1857 when the property was bought by La Société des Mines et Fonderies de Zinc de la Vieille Montagne [better known simply as Vieille Montagne (Altenberg in German)] a Belgian company. At least until 1972, the Belgian flag flew from the main shaft at Zinkgruvan.

The production from the mine in 1963 was 232,000 tons of ore that contained an average of 10.3 per cent zinc and 2.1 per cent lead. In 1973, the production of ore was 245,000 tons with a grade of 10 per cent zinc and 2 per cent lead. The galena contains between 0.059 and 0.196 per cent of silver or a maximum of about 2 kg per ton of galena. Whether the recovery of this silver is profitable or not, I do not know. The richest ores are reported to contain from 40 to 50 per cent zinc (Grip, 1978) with all grades intermediate between these and barren leptite are known. The total area of the ore surface is about 3600 m^2. Production in 1945 was 21,769 to tons of zinc concentrates and 2356 of lead; at that time the ore averaged 12 per cent zinc and 2 per cent lead. Thus, zinc grade gradually has declined whereas that of lead has remained practically constant.

The bed rocks of the Åmmeberg (Zinkgruvan) area are supracrustal rocks that belong to the Svecofennian cycle of the middle Precambrian. In a general way, three groups of rocks make up the major part of the Zinkgruvan area; these are, from north to south: (1) red leptite group, (2) gray leptite group, and (3) gray gneiss group. As a result of major folding, these rocks dip essentially vertically, but the detailed arrangement of these groups is far more complex than the previous sentence would suggest. In the first place, much gray leptite is intercalated in the red, and appreciable amounts of skarn, skarn-rich limestone, and limestone also are intercalated in both types of rocks.

Although the skarn almost certainly developed in rocks that originally were limestones, the skarn-forming process proceeded to different degrees in the limestones intercalated in the various parts of the two leptites. Henriques divides the lime-rich rocks into three classes: (1) limestone, skarn mineral less than 25 wt. per cent of the rock, (2) skarn-rich limestones, 25 to 75 wt. per cent; and (3) skarn rocks, more than 75 wt. per cent skarn minerals. These rocks grade gradually from one to another without sharp contacts. The principal skarn minerals are actinolite, diopside, garnet, olivine, phlogopite, and spinel; not all of these are found in any one assemblage, and some calcite always is present. The skarn rocks are much coarser than the limestones, and the western limestones in general are coarser-grained than the eastern; there are, however, many exceptions to this rule.

Henriques considers that the skarns were formed without any major introductions of material from outside the immediate vicinity but involved only the impurities in the limestone and the $CaCO_3$ initially present. He believes that the banding that has been preserved from the original rock structures indicates that not much foreign material could have been introduced into the rock system. His failure to find many examples of skarn cutting across the sedimentary structures also convinces him that only the original sedimentary material took part in the dynamometamorphic reactions. Where he does find such cross-cutting structures, he thinks that they were the result of tectonic activity even when a new suite of minerals was developed. Magnusson (1950), however, believes that the late Svecofennian palingenesis produced "sulphide- and skarn-forming solutions" as well as converting the rocks in the southern part of the Zinkgruvan district to veined gray gneisses. Magnusson thinks all this was the result of "migmitization" and of the production of ore- and skarn-forming solutions by that process.

It does appear that limestone layers are associated with most of the lengths of the two major ore layers, the Main Ore Layer (MOL) to the south and the Parallel Ore Layer (POL) 100 to 200 m or so north of the MOL. The three major limestone ore layers associated with the POL die out about 300 m east of the west end of the MOL. Farther to the east, in the area (gap), between Knallagruvan and Nygruvan, these limestone layers are cut off against an NNE-trending fault or disappear under the waters of Lake Trysjön. East of the lake and of the fault, the layers reappear but are much less regular in number or relationships to each other; they extend, however, in one form or another, almost to the east edge of the Zinkgruvan mineralized rectangle (see below).

The limestone (and skarn) associated with the MOL is in a single and much narrower layer of mainly limestone. This MOL and its accompanying limestone cut off abruptly at the west edge of the "gap" but resumes its association with the MOL where that ore begins in the Nygruvan portion of the district. This association of ore and limestone in the MOL continues to the southeast for about half the length of the MOL. Beyond that end point, no limestone is shown on Henriques' map; probably limestone continues with the ore, but is not shown because of its narrow width.

Beyond the east edge of Henriques' map (the county boundary), the gray leptite in which the ore layers are contained bends to the north, and the ore layers and the limestone follow this pattern. The details of what beds in the north-bearing stretch coincide with which beds in the east-trending leptite is not readily clear. Only minor lengths of ore are included in this north-trending leptite.

The strata-bound mineralization in the Åmmeberg (Zinkgruvan) district is essentially confined to a rectangle about 3.46 km in an east-west direction and 1.88 km from north to south. The actual portion of this area that contains ore-grade mineralization is, of course, only a tiny fraction of the total area of the rectangle. Most of the ore is confined to various sections of the two main mineralized beds, the MOL and the POL. The western one-third of the Åmmeberg ore mineralization is known as Knallagruvan area and the eastern two-thirds as the Nygruvan. Between these two areas is the "gap" (previously referred to). In this gap are several short sections of ore, irregularly oriented to each other; the three largest of these segments of ore in the gap may belong to the MOL, but this is not a firm conclusion.

West of the gap, the Knallagruvan MOL is essentially continuous for 1200 m along an arc that is convex to the SSE. The POL is much shorter in the Knallagruvan area than is the MOL; the POL runs from the same longitude as the west end of the MOL for some 280 m where it stops. At the western end of the two mineralized beds, the two are about 140 apart; the maximum distance they diverge from each other is about 200 m in the center of the 280 m of POL ore.

East of the gap, in the Nygruvan area, the MOL is continuously mineralized for about 1400 m. Followed from west to east, the Nygruvan MOL first forms an irregular arc that is convex to the northeast; for the remaining distance of the MOL, it is convex to the southwest. In this area, the POL is about 1150 m long with only one major gap, about 160 m in length. At the end of the 1150 m, the POL makes a right-angled bend to the north and continues in this direction for about 160 m. Between the MOL and the POL in the Nygruvan area, there are several short stretches of ore that are oriented at random, probably due to the effects of folding (and perhaps faulting); what their relationship to the MOL and POL may be (if any) is not known. In the northeast corner of the rectantle containing the Ammeberg ore (and even somewhat to the northeast of the rectangle boundary) are short lengths of mineralized beds that may be part of strongly folded MOL and POL, but this is far from certain; the longest of these mineralized stretches is about 325 m in a NNW direction.

Essentially all of the rocks in the Åmmeberg district were introduced during the Svecofennian cycle of the middle Precambrian. The earliest rocks are those of the early Svecofennian that resulted from intense volcanism during which the volcanic material was extruded from which the red leptite (and hälleflinta) was formed. These rocks were derived from potassium-rich lavas and tuffs that were rhyolitic in composition. These volcanics were followed by the middle Svecofennian lower Mälar series that included, from bottom to top, minor quartzite that was metamorphosed to gray leptite, limestone (that was later converted in various degrees to skarn-rich limestone or skarn), and then by a series of tuffs and lavas that ranged in composition from mafic material through dacite, rhyodacite, and rhyolite, all of which were strongly metamorphosed and are now gray leptites or gray skarn-banded leptites except for the metabasites that are now classed as gabbros, diorites, and metabasites. These volcanics were followed by epiclastic sediments that are now leptites that may or may not be skarn-banded. These rocks end the lower Mälar series. The rocks of the upper Mälar series that came after those of the lower were derived from tuffs and lavas produced by weak volcanism or from epiclastic sediments; the upper Mälar series is now all gray leptite. Following the upper Mälar series a stage developed during which even greater amounts of skarn-rich limestones and skarn were formed by the tectonic activity that, by this time, had decreased appreciably in intensity. This metamorphism of carbonate rocks was followed or accompanied by granitization of the epiclastic sediments, tuffs, and lavas of the Mälar series. Igneous activity then produced a wide variety of rocks ranging from peridotites (now largely altered to amphibolite) through gabbro and diorite to granodiorite and granite. This brought the middle of the Svecofennian cycle to an end. According to Henriques, the late Svecofennian cycle began with wide-spread migmitization that increased the proportion of limestone converted to skarn-rich limestone and to skarn. Migmatization affected the intermediate mafic rocks but did not change them enough to prevent them from being recognized as originally diorites and gabbros. The basites, however, were considerably altered toward amphibolites. Some of the migmatized rhyolites and epiclastic sediments of the Mälar series were changed to red gneiss and veined red gneiss, whereas a considerable (probably the larger) fraction of them was changed to gray veined gneiss.

During this late Svecofennian cycle, increased igneous activity produced medium- to fine-grained red granite and medium-grained gray granite as well as elated granite aplite. In the early late Precambrian (the Gothian cycle), some porphyritic granite was introduced into the area. The last of the Precambrian rocks to enter the district were to so-called Algonkian (late, late Precambrian) diabases.

The interpretation of the structural pattern of the bed rock in the Åmmeberg district is complicated by much of the rock mapped as gray veined gneiss includes metamorphosed gray leptites of both the lower and upper Mälar series. On the other hand, those areas marked on the map as gray leptites include no younger sediments because these younger rocks seldom, if ever, are in the non-gneissic state. All limestone intercalations of any importance are in the lower Mälar series. Thus, any migmatized rock that is host to any limestone (or metamorphosed limestone) inclusions almost certainly was derived from rocks of the lower Mälar series.

Further, much of the red leptite on Henriques' map probably was developed from epiclastic (red) sediments that belong to the Mälar series (both lower and upper) and, therefore, are appreciably younger than the early Svecofennian red leptites. A still further complication is provided by the gray leptites of both the lower and upper Mälar series that were converted into veined gneisses that are macroscopically similar to each other.

The known rocks of the two Mälar series that have been preserved are those that have been protected from erosion by down-folding. The Åmmeberg district is part of the largest of these down-folded units, and it has, to the north and east, a markedly undulating boundary that is characterized by bulges and indentations. The most westerly of these bulges contains the ore zone of the Zinkgruvan field. The lower Mälar series rocks are exposed in a narrow zone that runs from Lake Vättern (west) along the contact with the Gothian granites (to the northwest) and through the Zinkgruvan ore field. This zone of lower Mälar rocks widens out and bulges north towards Skepjön (lake) and finally SEE to Bresjön. This brief description does not complete the outline of the down-warped Mälar beds, but it is enough for a study of the Zinkgruvan district. The rocks of the lower Mälar series (the leptite-hällefinite group of red leptites) are exposed not only along the outer contact zones of this downfold but reach for considerable distances into the mass of Mälar rocks. All of the rocks dip at angles between 65° and 90° into the series of flat synclinoria of which the Åmmeberg zone is composed. The Zinkgruvan area constitutes a part of a steeply inclined synclinorium that is open toward the west. In the southwest parts of the synclinorium (except for minor and local deviations), the axes of the primary folds are oriented ENE-SWS. Henriques found it impossible to determine in detail the dips of the axes of these folds. This is partly due to the later development of variously oriented cross folds. In the central portions of folds, he was able to determine that the mean dip is practically horizontal, that is, the flat parts of the synclinorium. The outcrops of the leptite-hällinflinta series in the Zinkgruvan area can be explained by the alternations of depressions and culminations in these axes. It appears to Henriques that this synclinorium (as have the other synclinoria in the Åmmeberg area) has only a shallow depth. Further to the east, Henriques finds it possible to follow the synclinorium toward the northwest; the primary axes, however, are, with minor exceptions for the small culminations and depressions, run northwest-southeast.

In the Zinkgruvan area, the primary structures are determined by the shape of the underlying rocks and by the different thicknesses of the various original sediments, and the secondary ones are controlled by such processes as folding and such types of faulting as have affected the area. Up to the completion of Henriques work, it has been impossible to make certain the degree to which these various factors have controlled the present shape of the Zinkgruvan synclinorium (and of the others in the larger area).

The lobe-shaped bulges out from the synclinorium, such as that in which the Zinkgruvan ores are contained, may have resulted from interference between folding in two directions, the main folding (ENE-WSW) and the open cross-folds that are lined up generally along a north-south direction. The folding in two directions has caused some strata to have been greatly compressed and the schistosity of the gray gneiss, where it developed, to be oriented parallel to the original stratification of the parent. In these compressed strata, appreciable breccia, as well as schistosity, has been developed; these breccias are composed of metabasite and leptite that are connected to well-defined beds in the gneiss.

Henriques is uncertain in deciding the extent to which deformations of the ore zone in the Zinkgruvan area date from its original formation or occurred during later tectonic activity. He believes, however, that the main ore layer was deformed through slumping, that is, before the initial ore-bearing sediments had lithified. Nevertheless, he argues that the tectonic evolution (except for rare late Precambrian faulting) was completed before the formation of the late Svecofennian veined gneisses and granite. In fact, in the Knallagruvan portion of the ore zone, the section in the western part of that zone, the gneiss intersects the ore beds. In the same way, although not apparently affecting actual ore, the granite northeast of Trysjön(lake) and north of Gustafgruvan and the granite-aplites have intruded leptite beds that were already folded, but probably not mineralized.

The layered structure of the Nygruvan ore zone has been appreciably folded, particularly in the upper parts of the deposit in large numbers of open folds that usually have steeply inclined axes. Most of these folds are small-scale and give the ore layers no more than a gently undulating shape. In some places, however, the folding has been strong enough that material in the layers has been concentrated in the crests and troughs of the folds and the flanks have been attenuated. In other localities, the width of the ore layers has been considerably increased by the combining of folds that have differently oriented axes. Further, the ore layers are twisted by a change in direction of the fold axes, resulting in the shape of the ore bodies being considerably different from one level to another. Still further, in some places, the folding has affected only one of the ore layers (i.e. the Main or the Parallel ore layers). Only a few faults have had an effect on the Nygruvan ore layers; except in one instance (the fault that separated ores G and g) the displacements in both horizontal and vertical directions by the faults cutting the Nygruvan ore zone have been small.

In the Knallagruvan ores, located largely southwest of the Nygruvan ore trend, the deformations that affected the ore zone were much more impressive. Henriques believes that such effects were partly due to interference among folds along differently oriented axes (examples are ore A and 160-m East). In addition, considerable thickening in the ore layer has been achieved by isoclinal folding that, in one instance, has increased the width of the ore layer by six times. Such thickenings have taken place where the angles of different fold axes have converged at a common point. For example, the bowl-shaped Ore H that is open toward the east resulted from successive twisting of the ore layer plus its conjunction with the flat-lying Ores 165 m and 65 m West. These effects can be visualized only from large-scale three-dimensional models that obviously cannot be produced here nor are they shown in Henriques paper in which they are described in words; such models can (or could have been) seen in the geological office at the Åmmeberg mine.

Faulting does not commonly affect the Knallagruvan ore layers, although one large fault has cut through Ore 35 m West (actually the French words for east and west are used in the names of those ore bodies that have these directions incorporated in them). Henriques believes that deformation by folding, except where such was caused by slumping, seems to have been produced principally by concentric cleavage folding (a term he derives from de Sitter in his 1956 "Structural Geology"). Small late Precambrian movements in the bedding that were necessary for the concentric cleavage folding are in part indicated by faults in Svecofennian granite and pegmatite dikes that are displaced for short distances, where they cut the beds more susceptible to such movement, along the bedding planes.

In the ore zones, the hanging wall is on their northerly side, and the part of the ore layer that faces in that direction had been affected to various degrees by internal folding; this folding is more easily followed in the non-mineralized leptite bands. As the footwall is approached, the folding becomes less intense with the leptite intercalations near the footwall being conformable with the ore boundary without any evidence of major deformation. From this, Henriques says that it follows that the folding principally affected the upper beds of the Main Ore Layer, whereas, the lower parts of that layer and of the Parallel Layer have remained almost undisturbed.

Henriques says that there are several possible explanations for this type of deformation: (1) the upper beds of the MLO may have been folded independently of those beneath them, that is by disharmonic folding. If this were true, then the lateral shortening of the lower layers must have been of about the same order of magnitude as that of the upper ones. Henriques has seen no evidence of such shortening in these lower layers, so the folded nature of the MLO cannot be due to disharmonic folding, (2) the observed structures may have been formed by slumping of not completely stabilized sediments. That the upper beds slumped, whereas the lower did not, suggests that the lower were, at the time slumping occurred, sufficiently stabilized that they did not slump, (3) Henriques thinks that similar slump-like structures also might have been caused, even after they had been covered by younger sediments, such structures having been produced by internal flowage or readjustment resulting from the load effect of these overlying sediments.

Henriques sees no absolute proof of the manner in which these so-called slump-structures were formed, but it seems probable to him that they were caused by slumping or other movements that were confined within the layers affected. Later folding may further have emphasized these structures. Henriques assumes that these sediments, at the time that the slump-like structures were formed, contained all the ores that ever were to be included in them. This aspect needs, and will receive, further consideration later in these notes.

As has been mentioned, the economically workable ore bodies in the Zinkgruvan area occur in two neighboring beds, the Main Ore Layer (MOL) and the Parallel Ore Layer (POL). The ore zone is divided in an eastern part (Nygruvan) and a western portion (Knallagruvan). The ores are disseminated in various layers of the leptite, mainly in or adjacent to skarn or skarn-rich layers; the ores run in grade from poor to very rich with the maximum reported being over 50 per cent zinc and lead combined. In almost all of the ore bodies, the sphalerite is greater in quantity than the galena, but a few bodies in the Knallagruvan section on the POL lead in places exceeds zinc in amount. Other sulfides and oxides are present only in small amounts. The galena is silver-bearing, the silver ranging from about 0.06 to 0.20 per cent in that mineral.

As has been pointed out, the ores have a banded structure with the amount of mineralization being different from one band to another. The width of these bands ranges from a few millimeters to about 1 meter. When examined microscopically, the bands are seen to contain numerous sub-bands that differ in composition in both metal content and type of host rock - leptite, skarn-rich leptite, or skarn.

It should be made clear that the sedimentary bands, now from slightly to extremely folded, have been rotated by that process. This means that, for the ore layers to be standing nearly vertically (as they do now), they must have been rotated about 90° from the flat angle at which they were deposited. Since the red leptite lies to the north of the ore layers and gray veined gneiss roughly to the south, this indicates that the stratigraphic sequence will be oldest rocks on the north, intermediate aged rocks in the center, and the youngest rocks to the south. Thus, the stratigraphic hanging-wall is generally to the south of the steeply dipping ore layers and the stratigraphic footwall is to the north. It appears that, although the stratigraphic footwall normally is the younger beds; locally the situation may be reversed. It must be remembered that the ore-containing layers that form the floors and backs of the drifts are at angles of nearly 90° to what they were when they originally were laid down and that the rotation of the originally flat lying ore-containing layers has been up to, or even greater than, 90° perhaps the understanding of the ore-layer structure will be easier to visualize. Had the folding of the beds not involved this drastic rotation, the structure would have been appreciably easy to decipher.

The individual bands in the layers of the leptite have considerable continuity in the direction of their strike (allowance being made for the complex isoclinal folding that they have undergone) and can be followed for long distances. Thus, it is possible that one layer can be traced from one ore body to another or even through several. Further, the various mineralized bands show great vertical continuity. The orientation of the axes of the isoclinal folds that bring a given layer back along itself generally have the same orientation as those of the plunge of the ore body (or part of an ore body).

It must be emphasized that a given ore layer will contain some constituent minor layers that will have no ore content at all and no two layers in any ore layer are likely to have two minor layers that contain exactly the same amount of ore. In addition to minor layers of leptite, skarn-banded leptite, and skarn, they also will contain limestone and skarn-rich limestone layers. Further, the number of mineralized layers in a given ore body will differ from one place to another in the various ore layers that lie along the strike of the major layer. Obviously, the more ore-containing minor layers in an ore layer, the larger the body will be, and the richer the mineralization in the mineralized layers it contains, the higher-grade the major layer will have. Some major ore layers, in addition to having unmineralized leptite and skarn bands in places will include fragments of these same rocks, the size of which will range from a few cm^3 to several dm^3. These fragments normally have sharp contacts with the ore in the bands that contain the fragments. In some places, however, where the tectonic effects have been unusually strong, these fragments have been fractured and the fractures filled with sphalerite and galena. Henriques thinks that these fissure fillings were the result of remobilization of primary (sedimentary) sphalerite and galena. In most instances, this secondary mineralization (so-called) results in with the bulk of this mineralization being concentrated in the outer portions of the fragments. The matter will be discussed further later.

From all this it follows that what is seen in the walls of the drifts are horizontal plans of the ore-containing layers (and of the entire leptite formation for that matter) and that the floors and backs are what originally would have been vertical strike sections through the ore-containing layers. Allowance must be made, of course, for the departure of these plans and sections from a true 90° to the position they original occupied when the leptite layers were tectonically undisturbed. Thus, the lengths of the ore-containing layers are the strike lengths of those bodies as they were deposited; the vertical heights are the original widths, and the present widths are the initial horizontal thicknesses; that is, the original ore-containing layers were rotated around their strikes as axes of rotation. The greater the degree to which the ore-containing layers have been folded and refolded, the greater the vertical thickness of the layer in question and very probably of the present ore layer width as well.

The ore are separated from the wall rocks by sharp contacts that lack transition zones (so far as the mineralization is concerned) and are parallel to the stratification of the wall rocks. If the outlines of the ore bodies are irregular (as they are in many places), these irregularities are matched by similar irregularities in the wall rocks. Locally, the ore contain pegmatitic material that has vague contacts with the rocks (leptites) that surround them and probably (Henriques thinks), the result of partial remobilization and slight movement of molten material derived from the leptite. These pegmatites have cut the ore only to a very limited extent. These pegmatites are most common in the veined-gneiss area south of the MOL and close to the boundary of that layer with the gneiss.

In the Nygruvan ore layers, the content of lead and zinc diminishes to the east; the higher-grade ores (identified by letters from A to G are located to the west of a northeast-striking fault that is displaced to the southwest on the southwest side. The Knallagruvan ore, however, are much the same in metal content, although slightly richer ores are located in the central part of that ore layer. On the lower levels of the various ore bodies, the grade of the ore drops off due to a greater content of leptite and pegmatitic material. Between 1956 and 1963, the Nygruvan ore averaged

11.0 per cent zinc (the 1963 grade was 9.6 per cent) and 1.1 per cent lead, (the 1963 grade was 1.3 per cent). The Knallagruvan ore averaged 11.75 per cent zinc (the 1963 grade was 13.7 per cent) over these eight years and the lead grade was 4.0 per cent (the 1963 grade was 5.6 per cent). For the entire operation the 1963 grades were 10.3 per cent zinc and 2.1 per cent lead (all these per cents were weight per cents). The ratio of zinc to lead in Nygruvan is about 10 and in Knallagruvan about 3, confirming that the lead increases in the ore as it is followed to the west. In the Nygruvan area, the Zn:Pb ratio is greater than one in the MOL and less than one in the POL.

The mineral composition of the ores ranges widely because the ores consist of varied amounts of mineralized and non-mineralized bands and different proportions of leptite, skarn, and limestone. If the ore minerals are left off the account, the ore layers contain the following percentages: quartz - 32,8; plagioclase - 1.0; microcline - 27.3; biotite-chlorite - 4.3; muscovite - 1.6; amphiboles - 11.7; diopside - 9.9; garnet - 42; epidote - 0.4; zoisite - 3.1; calcite - 2.5; plus minor to trace amounts of several other minerals.

In addition to the sphalerite and galena, the ore minerals are made up of: minor amounts of pyrrhotite (0.2%); chalcopyrite (0.1 to 0.2%); plus traces of native silver, graphite, tethrahedrite, proustite (?), pyrite, arsenopyrite, magnetite, and secondary marcasite and limonite.

Considerable work on the minor-element contents of the Zinkgruvan sphalerite has convinced Henriques that these depend on the availability of the minor elements and not on any equilbrium constant. Too little of these minor elements was available to saturate the sphalerite with any one of them, so they tell little about temperatures or other conditions of sphalerite formation. Henriques spent a great deal of time and effort on trying to use Kullerud's diagrams to determine the temperature at which the sphalerite were initially formed or, since he believes that they were deposited originally as primary sediments, the temperatures that may have affected them subsequent to their deposition. The discrediting of this method of determining temperatures of sphalerite formation or recrystallization makes it unnecessary to consider this topic further. A matter of considerably more use, however, is the presence of exsolution bodies of chalcopyrite in the sphalerite. No agreement has been reached as to the temperatures that such exsolution chalcopyrite may indicate, but it almost certainly is high enough that some time during the life history of the Zinkgruvan sphalerite it was at high enough temperatures to dissolve some chalcopyrite and later dropped to low-enough temperatures that the chalcopyrite came out of solution. It is difficult to imagine sedimentary sphalerite (granted such was the origin of that mineral in the Zinkgruvan deposits) at a high enough temperature to initially have taken chalcopyrite into solid solution. It is conceivable, of course, that, during metamorphism long after deposition, the temperature of the sphalerite rose sufficiently to take chalcopyrite in solution in it, granted that some reasonable method for moving the chalcopyrite into the lattices of the sphalerite can be suggested. Perhaps the pegmatites intruded into the neighborhood of the ore-containing layers may have provided a mechanism for introducing chalcopyrite into sphalerite, the temperatures developed well may have been high enough, but it seems unlikely this rather round-about method of adding copper and iron to the sphalerite is the most satisfactory one that can be devised. In any event, the presence of these exsolution bodies of chalcopyrite in the sphalerite indicates that, at some stage in its career, sphalerite was being deposited or metamorphosed at temperatures in excess of 400°C.

Going from south to north in the Zinkgruvan ore field and just beyond the veined gneiss, an elongate sulfide zone that consists mainly of gray leptite with various amounts of pyrrhotite and pyrite appears. This zone is about 100 m from the MOL and POL. Still further north, but mainly only in the east part of the field, are additional ore layers, though these are short and generally narrow. All of the MOL and POL ore-containing layers, plus those to the north of them contain zinc-lead ores, whereas those south of the MOL form a pyrrhotite-pyrite zone. Nevertheless, in pyrrhotite-pyrite layer contains some minor amounts of sphalerite and galena; the reverse is true in the MOL, POL, and the small ore-containing zones to the north of them. It is, however, definite that the zinc-lead ores have been developed in carbonate-rich rocks and the iron sulfides have been concentrated in normal leptites. Although Henriques points out that the variations in ore type from one minor layer in an ore-containing major layer to another may be due to the compositions of these minor layers being more favorable in some instances, to the deposition of sphalerite and to that of galena in others, he recognizes that the lack of veins or veinlets connecting one minor ore layer with another is an argument against epigenetic mineralization, but he admits that this type of mineralization does not require such "jumping" from one minor layer to another. He thinks, however, that the argument that thin ore laminae in sedimentary beds is not evidence for a syngenetic manner for formation for ore bodies because "deposits of demonstrable epigenetic origin have such a structure - the Sullivan mine in British Columbia for example." About the same time as this paper of Henriques was written, the concept of a syngenetic origin for the Sullivan ores began to become popular, a theory that commands nearly universal acceptance today.

Henriques also points out that the essential lack of wallrock alteration in the leptites immediately adjacent to the skarn-rich rocks in which the MOL occurs may indicate that the ore must have been syngenetically introduced. What alteration there is in rocks adjacent to ore-containing layers, that is, moderate reduction in the amounts of microcline, plagioclase and epidote-zoisite and the moderate increase in quartz and muscovite could be used to justify the postulate that minor wall-rock alteration had taken place during ore emplacement, minor because of the relative stability of the gray leptite in relation to the ore-forming fluids that deposited the ores, whether they be hydrothermal fluids in the narrow sense or hydrothermal fluids debouching from beneath the floor and mingling with the sediments already deposited on the sea floor.

Henriques raises the point that the alteration of the Zinkgruvan rocks is minor in comparison with that of Falun where he is convinced that the ore-forming fluids were hydrothermal in the narrow sense; this, he believes, argues for the hydrothermal fluids have come from the rocks immediately below the sea floor.

No doubt is possible as to the replacement characteristics shown by the sphalerite and galena ores at Zinkgruvan. All types and stages of mineral replacement have been observed by Henriques such as pseudomorphs, replacement outward from cleavage planes and fractures, and rim and vein replacements. The rocks of the ore layers have been subjected to mechanical movement that has produced a multitude of cleavages outward from which much replacement has taken place. Those minerals most readily replaced are carbonate and amphibole and pyroxene minerals; these are much more readily affected than skarn garnet and original quartz and feldspar in the leptites. As Henriques believes that the ores of the Zinkgruvan field were formed at the same time as the sediments of the Mälar series, they later must have undergone extreme metamorphism; this means he thinks that ore and host rocks possessing replacement features well may be of sedimentary origin. Such metamorphism is known to have taken place in both middle and late Svecofennian orogenic activity.

As for paragenesis (in the narrow American sense) the order of crystallization, except for the abundant sphalerite and galena, is difficult to establish. This much is certain: normally sphalerite crystallized before galena; pyrite, pyrrhotite, and chalcopyrite crystallized after sphalerite and galena; and the position of the rare arsenopyrite could not be established. This order of mineral deposition agrees with the usual sequence proposed for hypothermal minerals in the [modified] Lindgren classification, allowance being made for not unreasonable reversals in the exact order due to the tremendous preponderance of sphalerite and galena over the other sulfides.

Henriques suggests that the sphalerite-galena mineralization may have been syngenetic and the pyrite, pyrrhotite, chalcopyrite later and probably epigenetic. Such a scenario requires considerably more coincidence than either the straight syngenetic plus metamorphism sequence or the straight peigenetic one. Henriques thinks that it is easier to envision to two-period sequence as being correct.

Henriques points out, as is well known [though much more so than when he was writing], that syngenetic sulfide ores may have deposited: (1) from hydrothermal ore fluids [in the narrow sense] that reached the sea floor; (2) from volcanic exhalations that also reached the sea floor, a concept now thoroughly discredited because of the inability of such gases [unless under very high confining pressures, in which instance they would behave as liquids] to transport the ore elements and the necessary sulfur whether it was derived from depth or picked up from sea water, (3) by chemical sedimentation of material brought in by rivers from the surrounding land surface this concept is largely neglected these days simply because of the difficulty of moving sufficient metal at any given moment of time to produce concentrations in the sea-floor basin great enough to force precipitation of the sulfides required, and (4) by precipitation from solutions generated by first downward movement of leaching groundwater and later deposition from these solutions (now upward moving) that had picked up the necessary metal ions on their travels beneath the surface and had later deposited sulfide ores when they encountered [with a rare sense of direction] waters containing the necessary amounts of sulfide ion to produce the usually wide variety of sulfide present in the average sulfide ore deposit. Henriques favors, most reasonably everything considered, as the far most possible of the four possibilities cited above, that of hydrothermal fluids from depth, equipped with the required sulfide ion, and able to precipitated sulfides either in lithified rock or on the sea floor.

If the ores were deposited originally on the sea floor, they cannot have achieved their present characteristics without the aid of metamorphism on an appreciable scale; to this concept, Henriques also subscribes. If the hydrothermal fluids that deposited the ores on the sea floor derived from the middle Svecofennian granites, as it is conceivable that they may have done, then their metamorphism almost certainly is related with those orogenic effects that definitely produced late Svecofennian gneisses so important in the Zinkgruvan area. In addition, to the metamorphism having converted sedimentarily produced ores into what is now found in the Zinkgruvan area, Henriques goes to considerable length to show that the zinc (and of course the lead as well) could have been driven off from the rock material converted to have formed the ores palingenetically outside the area of gneiss development. This is essentially Knight's "source-bed" concept proposed in 1957 and gathered its ancestors shortly thereafter. Henriques prefaces his final summary on the genesis of the Zinkgruvan ores by saying that "the genetic interpretation of sulphide ores of pre-Cambrian age is as a rule beset with considerable difficulties - often insurmountable with the methods available at present". There can be no quarrel with this statement, but Henriques goes on to remark that "this investigation has shown no infallible interpretation of the genesis of the ores was possible; the implication of most factors could be linked equally well to epigenetic or syngenetic ore formation. Nor has a postulation and discussion of different modes of ore formation provided unambiguous proof as to the genesis". Henriques' subjective opinion, however, is that "in all probability, the ores were syngenetically formed from ore material originating from submarine emanating hydrothermal solutions".

If this is to be the final judgement on the genesis of the Åmmeberg ores by Henriques, I might reasonably be allowed to give my own subjective opinion as to the matter. My disagreement with Henriques arises from: (1) the recent indications that sulfides poured out on the sea floor are quickly oxidized and are not incorporated in bedded sulfide deposit either ones similar to Åmmeberg or Sullivan or any other sulfide deposit that follows the bedding in a host rock. Rather, to preserve their sulfide character, these ores must be emplaced at some depth, perhaps only a few meters or 10s of meters below the sea floor but certainly not on it.

In the case of the Åmmeberg ores, however, the temperature at which they may have been deposited lies in the hypothermal range, that is 400°C or higher, so that the distance to the sea floor probably would have been as much as 100 m below the sea floor with the rocks between the depositing ores and the sea floor being quite well lithified by the time sulfide deposition took place.

Henriques is convinced that none of the factors that can be connected with the formation of the Åmmeberg ores oppose the syngenetic interpretation; he believes that actually most of his observations that bear on ore formation do more than not oppose a syngenetic origin for the Zinkgruvan ores but that suggest that they were developed syngenetically. He is certain that the structures that are characteristic of the ores agree more consistently with a syngenetic than with an epigenetic mode of ore emplacement. He selects for this purpose: (1) the consistently identical stratigraphic position of the MOL even though the ore zone (he says) contains a wide zone of apparently equally favorable host rocks (skarnbanded leptite) that are joined directly to the MOL, (2) the lack of contribution by pre-ore tectonism to the localization of the ores, (3) the long distance over which the fine layering of the ores can be followed even through more than one ore body in the MOL, and (4) the absence of any veins or veinlets connecting one ore-bearing fine layer with another. (5) Henriques is particularly impressed by the slumped structures in which the ores are incorporated; he is convinced that, when this slumping occurred, the ore already was present and took part in the actual slumping process, (6) evidence for wall-rock alteration is minimal (as has been detailed above), and (7) the low temperature of formation of the ores as indicated by the low iron content of the sphalerites (the indications for this being based on Kullerud's now-discredited method of sphalerite geothermometry). Henriques thinks that the low content of epiclastic minerals in the ore does point toward a syngenetic hydrothermal mode of ore formation but says that this fact does not eliminate chemical sedimentation as the way in which the Zinkgruvan ores were formed.

Henriques quotes Lindgren saying in the last (4th) edition of his "Mineral Deposits" that "one of the most mysterious of ore deposits is that of Åmmeberg in Sweden, where the zinc is disseminated in banded and contorted gray gneissoid "leptite." Lindgren concludes that "the deposit certainly seems to be of syngenetic origin and the mineral association shows that it has been subjected to high temperatures.

It is difficult to argue against the man who has studied the Åmmeberg ores longer and more thoroughly than anyone else and against the Dean of American ore geologists in the first one-third of the twentieth century, but some argument seems reasonable and even to be rather soundly based. The first point, mentioned earlier in these notes, is that recent work indicates that sulfides extruded in a gray mist on the sea floor are so quickly oxidized that the possiblility of their having been the parents of massive or banded sulfides deposits is unlikely. The second point is that the chances of a subterranean source furnishing alternately deposited and microscopically narrow bands of igneous minerals followed by ones of similar size that are to a large degree composed of ore minerals is a mechanism not easily envisioned.

All of the arguments about lack of wallrock alteration, lack of veins or veinlets connected one ore-containing bed with another, the slumped structures in which much ore is concentrated, the consistent positioning of the ore at a single stratigraphic sequence(s), the failure of equally, the lack of pre-ore tectonism than can be said to have provided pre-ore structures that localized the ores, and the supposed low temperature at which the ores were formed. All of these items, individually or as a group, can, as Henriques admits, be compatible with epigenetic ore formation. Certainly, his concept of a low temperature of ore formation, based on sphalerite geothermometry has been discredited, and the exsolution textures of chalcopyrite in the sphalerite are good evidence that the ores were either deposited initially at temperatures over 400°C or were affected by metamorphism that not only raised the temperature of the ore originally deposited at low temperatures to 400°C but also provided a mechanism by which chalcopyrite could be not only moved from somewhere but also introduced into a temperature high enough that the chalcopyrite would be taken into solution and then, with lowered temperature, exsolved. Finally, the presence of ore minerals in what must have been slump structures produced by slumping of unconsolidated sediments, but this origin of the slumping is no guarantee that the sulfides were among the sedimentary materials that slumped.

Certainly, the final verdict on the formation of the Åmmeberg ores is not in, and it may never be. To classify this deposit, however, is simply to choose between two categories in the modified Lindgren classification. The first is: by interaction of solutions in quiet waters (inorganic reactions), plus metamorphism-C or hypothermal in calcareous rocks (the ores are always associated with skarn-rich leptites or perhaps with skarns) and not with plain siliceous leptite. Thus, the choice must be made between Sedimentary-Ala, plus Metamorphic-C or Hypothermal-2 plus Metamorphic-C(?). My choice is for the latter with the understanding that I may well be wrong.

SKELLEFTE DISTRICT (GENERAL)

Middle Precambrian	Copper, Silver, Gold, Zinc, Arsenic, Pyrites	Hypothermal-1 to Mesothermal minor Leptothermal

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Notes

For the purposes of this volume, the Skellefte district is divided into three geographically separated areas. These are: (1) Boliden-Rakkajaur; (2) Kristineberg-Rävlidmyran and (3) Adak-Rudjebäcken; these are much broader areas than those into which Grip (1978), the authority on Boliden, divides the district; his division is as follows: (1) Boliden; (a) Boliden; (b) Lånsele; (c) Långdal; (d) Akulla; (e) Kankberg; (2) Renström (alone); (3) Petikträsk; (a) Holmtjärn; (b) Aliden; (4) Mensträsk-Elvaberget anticlinorium; (a) Mensträsk; (b) Elvaberget; (5) Malånäset anticlorium; (a) Högkulla; (b) Maurliden; (c) Norrliden; (d) Bjurfors; (6) Näsliden dome; (a) Rakkejaur; (b) Näsliden; (7) Vindelgransele and Kristineberg anticlinoria: (a) Kristineberg; (b) Kimheden; (c) Rävlidmyran; (d) Rävliden; (8) Adak dome: (a) Adak; (b) Lindsköld; (c) Brännmyran; (d) Rudtjebäcken; (9) Kedträsk anticlinorium: (a) Svansele; (b) Udden; (c) Kedträsk; (d) Åsen. The district also contains two nickel deposits one of which, (Lainijaur) is discussed in this Volume and one pegmatite (Varuträsk) that is discussed in the Volume also.

The Main Ore Belt, as chosen here, extends from Boliden at the extreme southeast of the district to the Rakkejaur mine on the northwest, a distance of nearly 65 km. The Adak-Rudtjekäcken district is located some 33 km further northwest from Rakkejaur. The Kristineberg-Rävlidmyran district lies about 33 km west-southwest of the Rakkejaur mine. Far from all of these mines have been described in the literature; in Grip's 1960 reference under the Skellefte (General) heading, some information is provided as to the Renström mine. In Kautsky's 1960 reference in the same bibliography some data are given as to the Rakkejaur ore body, the Adak-Lindsköld mines. The Kristineberg and Rävlidmyran mines also are discussed in Kautsky's 1960 reference [Skellefte (General)]. In addition to the considerable number of papers published on the Boliden mine, individual papers in print are concerned with the following mines: Mensträsk, Kankberg, Adak, Lindsköld, Brännmyran, Rudtjebäcken, Kristineberg, and Rävliden. As can be seen from the four lists of references concerned with deposits in the Skellefte area, several of the major mines are not described in the published literature, although they must be treated at length in reports of the Boliden Company.

Intensive prospecting in the Skellefte region did not begin until after the end of World War I in 1918. At that time very little was known about the geology of this supracrustal belt that extends westward from the coast (the Gulf of Bothnia) for some 200 km, except that there was wide-spread dissemination of sulfides in black phyllites and a presence of a minor number of veins that carried gold, quartz, and arsenopyrite. Mining had been tried on a small scale as early as the 18th century, but no success was reported then or for many years. As is true of so many districts in Scandinavia, the most important stimulation to prospecting was the discovery of heavily mineralized boulders in the glacial debris, by specially trained boulder hunters. These discoveries, correlated with the direction of movement of the glacial ice, made it possible to indicate the approximate location of the ore body from which the boulder came. This work was followed by detailed geologic mapping and tectonic analysis and then by geophysical investigations without which last few ore bodies were found. Of the geophysical techniques employed, electrical methods were the most helpful. The first discovery by that method, that of Kristineberg, was made in 1918 only a short time after full-scale prospecting had begun. The types of geophysical exploration were broadened to include airborne studies and borehole logging. The electrical methods were supported by magnetic and gravimetric work, and their sensitivity greatly improved through studies carried out in the Skellefte district.

The first ore body to be mined was that at Holmtjärn (a body of small size but rich in gold) in 1924 through 1925. The first discovery of major economic importance was that of Boliden itself; the boulders that lead to finding it were encountered in 1921, and its location by electrical measurements and diamond drilling was accomplished in 1924. Boliden was initially mined in 1926 and was continuously operated until 1967 when the ore was exhausted. The average grade of the Boliden ore was 1.4 per cent copper and 15.5 ppm gold (Grip, 1978).

The total area of outcrop and suboutcrop of all the 100 Skellefte ore bodies is approximately 150,000 m^2, and the average grade of all of them is 1.5 ppm gold; 39 ppm silver; 0.8 per cent Cu; 2.3 per cent Zn; 0.2 per cent lead; 0.8 per cent arsenic, and 30 per cent sulfur and 31 per cent iron. Of the 100 ore bodies found, 25 have been mined, and the production from them has totaled 48 million tons through 1973. In 1973, 10 of the 25 mines still were operating. The 18 most productive mines (through 1974) are in total tons mined:

Kristineberg	9,941,000	Rävlidmyran	3,878,000	Långdal	723,000
Boliden	8,342,000	Näsliden	1,102,000	Kankberg	253,000
Adak & Lindsköld	5,923,000	Åkula	980,000	Bjurfors	200,000
Långsele	4,801,000	Udden	958,000	Kimheden	131,000
Rudtjebäcken	4,625,000	Rävliden	888,000	O. Högkulla	114,000
Renström	3,879,000	Rakkejaur	726,000	Kedträsk	61,000

In productivity, these mines divide into three groups: (1) seven that range upward from nearly 4 million tons to nearly 1,000,000,000; (2) six that range between over 700,000 upward to 1.1 million; and (3) five that lie in the range between 250,000 and 60,000. Obviously, the relative ranking of these 18 mines will have changed in the nine years since Grip's figures were compiled, but the ranking groups listed above probably contain the same mines though with different totals (except for Boliden) and different rankings.

The total production from the 18 mines listed comes to about 47.525 million tons. This is only 475,000 tons less than the total given by Grip (1978) for all 25 mines over the history of mining. So the remaining 7 mines have averaged only about 68,000 tons each. If this is true, then some one (or even more) of these mines must have had production greater, at least, than that of Kedträsk. Or, of course, Grip's figure of 48 million tons may have been rounded upward from less than that amount. Or, finally, enough tonnage may have been recovered and processed from the 75 ore bodies not officially mined to drop the average of the last seven of the 25 to below that of Kedträsk. The answer to this problem almost certainly never will be known.

The ore bodies of the Skellefte district as a whole are located in supracrustal rocks (i.e. rocks above the Archean basement), and this belt of ore bodies lies in a belt 200 km long and the width of which is, at a maximum, about 60 km. The district follows the SSW-trending course, of the river - Skellefte älv. - or, better, the river follows the belt of SSW-trending supracrustal rocks which certainly antidate the formation of the river valley. Granites and gneisses surround the valley on all sides except the north where the adjacent district in that direction - the Arvidsjaur - is mainly composed of volcanic rocks.

The port of Skellefteå on the Gulf of Bothnia is located at 64°45'N, 21°00'E and is at the southeast end of the Skellefte belt; Boliden is situated at 64°50'N, 20°20'E. The area is the northern edge of the Norrland geosyncline that reaches southward for about 400 km to end in central Sweden.

The oldest rocks in the district are the Volcanic group that is made up of quartz porphyries, quartz keratophyres and keratophyres; this sequence includes pyroclastic equivalents of these rocks, plus some interlayered sediments. These silicic volcanics are principally ignimbrites; because of their high content of water, they were highly fluid when erupted and, therefore, covered large areas. In many places, these volcanics included schlieren or fragments of the composition of the groundmass surrounding them. Few true lavas have been identified in the volcanic group.

At the same time that the Volcanic group was folded about horizontal axes by presure directed from the NNE-SSW, the synkinematic, soda-rich early Svecokarelian Jorn granite (actually mainly a granodiorite). Its age is estimated by Welin and his colleagues (1971) as about 1900 m.y. (not based on radioactive determinations but on general geologic relationships). Some of the folding appears to have been related to buckling of the supracrustal rocks due to the development of granite diapirs. The dome structures so produced are directly related to the localization of many of the Skellefte ore bodies. Long, thin granite bodies in the southern portion of the district are closely integraged with the fold tectonics (Grip, 1978); also so related is the major massif of the Jörn granite. This granite has almost straight-line contacts on its northeast and southwest sides; such a linear arrangement is repeated in several later stages of intrusion.

After a period of uplift and erosion in the district, later rocks were deposited unconformably over the whole district; this discordant relationship is best seen in the northern part of the district where the Jörn granite, originally intruded at great depth, was raised to the surface and deeply eroded. To end this erosional stage, the seas transgressed the area from south to north; coarse-grained debris was brought in from the north to cover what is now the Skellefte älv valley, one mass of which is now known as the Vargfors conglomerate. At much the same time, volcanism recommenced in the district on a large scale with the principal types of rocks being of intermediate or mafic composition that were extruded from both fissures and cones (Rickard and Zweifel, 1975). These authors conclude that this volcanism was accompanied by the formation of volcano-exhalative sulfide ore bodies. This volcanism also was joined by the introduction of the mid-Svecokarelian

Gallejaure granite. Some limestone was deposited in the basin of sedimentation where deeper water was the rule.

During the period of transition between the Volcanic group and the overlying phyllites, the area was tectonically unstable, but the time of deposition of the Phyllite group was much less so, it being accompanied by only minor volcanic activity, mainly of a mafic type, extruded under submarine conditions. Within the phyllite group are a series of intraformational conglomerates that show that sediments that lay above water level was eroded and redeposited as part of the phyllites.

In contrast to the northern area where intraformational conglomerates were common, the southern sea was appreciably deeper and sediments were so thick that recrystallization and granitization took place there. A second development of diapir domes was produced by the upward movement of this palingenic granite. What Grip calls a period of northward movement of the migmatitic front occurred at this time, and a second period of folding took place. Initially, at least the direction of pressure was about the same as that of the early folding (i.e. NNE-SSW) but gradually this direction was rotated in a counterclockwise direction (i.e. further toward the north). This folding of the second phase was more intense with the deformation producing drag folds and steeply dipping schistosity; local block faulting took place. Although Grip (1978) is certain that this Revsund granite was palingenic (with an age of about 1685 ± 40 m.y.), he points out that the granitic material, particularly in the north of the district, was intrusive into the older rocks, forming tongues trending north or northeast or developing diapirs, both within the enclosing sediments.

Grip holds that the closing phase of granitization (palingenesis) caused mobilization of the earlier, strata-bound sulfides and the development of the ore bodies now known in the present Skellefte district. The ore fluids produced during palingenesis moved to their sites of deposition through channels provided by the ancient land surfaces encountered, through fractures and zones of weakness resulting from the folding of beds of different lithologies and competence. Deposition occurred in surfaces of schistosity, drag folds, small faults, thrust and gash veins where these were covered by suitable cap rocks. As has been mentioned above, the numerous tectonically produced domes were major localizers of ore deposition.

Rickard and Zweifel (1975) show that the ores were emplaced in a wide variety of rocks. For example, Kristineberg and Rävliden occurred in regionally sericitized phyllites, low in the formation of Kristineberg and near to, or at the top of Rävliden. Boliden was deposited in a stratigraphically equivalent level to Rävliden, although mostly in keratophyre rather than phyllite. Näsliden was developed at the contact between keratophyre and quartz porphyry (below) and phyllites with local limestone (above) and, therefore, slightly above Rävliden. Elvaberg was introduced into calcareous tuff at about the stratigraphic level of Rävliden, and Udden was at the same level but in skarn-banded tuff. Renström was emplaced about the level of Kristineberg, as was Kankberg, the former in mixed rocks and the latter in tuffitic keratophyre. Långsele was slightly above the level of Boliden, probably in the transition zone between quartz keratophyre (below) and phyllite (above). Långdal is slightly below the level of Boliden in quartz porphyries and quartzitic schist. Svansele is located in the midst of quartz porphyry at a stratigraphic level that is not matched by any other deposit diagrammed by Rickard and Zweifel.

To summarize, all of the deposits diagrammed are in the Volcanic series and not in the Phyllite series except for two (Rävliden, in part, and Näsliden) that are at the contact between the two series. It appears, therefore, that the rocks of the Volcanic series, under the tectonic activity to which they were subjected in late Svecokarelian time, were much better hosts to deposition from the ore-forming fluids that invaded the rock volumes of the Skellefte district than were those of the Phyllite series. It is possible, of course, that every ore fluid that entered the rock volumes that now contain the Skellefte ores deposited essentially all of its sulfide load by the time it had passed the contact between the Volcanic and the Phyllite series. This seems less likely than that, once the ore fluids had passed that contact, they did not find enough structurally favorable sites to enable them to deposit sufficient sulfides to make economically exploitable ore deposits.

Although there are younger rocks than the Revsund granite in the general Skellefte area, including the 1625 ± 45 m.y. Sorsele granite, none of these is directly or indirectly connected with the ore-forming process. If the ore fluids that deposited the Skellefte ore bodies did not come from the Revsund granite, no matter whether it was magmatic or palingenetic or miraculous, it is difficult to imagine from whence they did come.

Rickard and Zweifel (1975) limit their discussion of the Skellefte district to those deposits listed in the first page of this discussion to those listed under headings (1) - Boliden-Rakkejaur and (2) Kristineberg-Rävlidmyran they omit the Adak dome. The total surface area of these two sets of deposits is given by Rickard and Zweifel as 150,000 m^2. Further, they include only deposits that are pyrite-rich and stratiform, which, they say, make up most by far of the mineable ore in the district; neither the two nickel deposits nor the quartz-arsenic veins are discussed by them.

They consider the ore as occurring in lens- or slab-shaped beds more or less elongated in the direction of strike and particularly in that of the local linear structures. They prefer the term strata-bound rather than what they consider the more controversial stratiform designation. They do this because they hold that stratiform implies certain basic parameters that govern age of ore and of regional metamorphism.

The Skellefte ore bodies normally are made up of several distinct bodies or lenses. Of the literally hundreds of such bodies and lenses, the Rakkejaur is the largest; it is about 550 m long and 65 m broad and had been followed down to 320 m below the surface - its area is as large at this depth as it was at the surface. It contains an estimated 17 million tons, of which only about three-quarters of a million had been mined through 1974 (Grip, 1978).

Rickard and Zweifel distinguish three major ore types: (1) sphalerite as the main ore mineral; (2) disseminated ore that is mainly pyrite and chalcopyrite; the chalcopyrite is the main ore mineral in disseminated zones outside the massive ore bodies around which the disseminated ore usually is located and in the massive lenses of pyritic material as well; (3) banded ore with the banding occurring mainly in pyrite-sphalerite ores and being due principally to alternate layers rich in one of the two sulfides, but sulfide layers also may be separated from each other by layers of quartzite. The thickness of these bands ranges between some millimeters and some centimeters. Essentially all ores have this banding, however, although it is less well emphasized in the massive ores. Some of the best examples of banded ore are found in the mines of Renström, Rakkejaur, and Udden and in the sphalerite-rich ore of Rävliden and Rävlidmyran.

The most consistently common mineral is pyrite with the remaining sulfide minerals being present in highly varied proportions from one ore body to another or even within one ore body. The main non-ferrous sulfide minerals are sphalerite, chalcopyrite, and galena with arsenopyrite being common and, in places, abundant. The pyrite content ranges widely with sulfur at Långsele being 36 per cent but at Långdal only 12 per cent.

The two most productive horizons in the Skellefte district are: (1) near the contact of the Volcanic and Phyllite series and (2) the lower portions of the Volcanic series. Of the most important ore bodies, many are situated in silicic and intermediate pyroclastics near the black, graphitic schists that make up the lowest part of the Phyllite series; some of the ores are associated with limestone or dolomite horizons, and a few lenses are present in the lowest part of the black schists. The ores in the lower part of the Volcanic series generally are in pyroclastic rocks, again near either calcareous zones or lenses of black schists. Some of the ore bodies are near the contact between quartz-rich rocks or keratophyres. Locally, some ore lenses are surrounded by metasomatically produced quartz-sericite quartzites in which relict volcanic textures can be identified.

Rickard and Zweifel consider that the association of the ores with the rock types in which they are found indicates that they were formed in a submarine volcanic environment. This does not mean that the ores were formed at essentially the same time because the ores are emplaced in solid rocks, many of which have been broken by later tectonic activity. This association of rock types and ore does not indicate a genetic connection but, instead, that the contact zone between volcanics and phyllites was structurally the most favorable for the emplacement of ore sulfides being brough in in ore fluids very probably derived from the same general magma chamber as the earlier volcanic materials.

The order of base-metal abundance (Rickard and Zweifel, 1975) is iron, zinc, copper, and lead with iron always overwhelming in its dominance. The relative abundance of the remaining three differs among ore bodies and even within ore lenses in any one deposit. Pyrrhotite is abundant in a few places and is consistently erratic in quantity. Over 60 per cent of the deposits have sulfur content between 20 and 35 per cent; only 25 per cent of the deposits have arsenic present as more than 0.5 per cent; of these Boliden has the highest arsenic - 6.8 per cent; South Nasliden and Rakkejaur were next with 1.6 per cent and 1.3 per cent, respectively. Much of the arsenic occurs in arsenopyrite-quartz veins associated with the Revsund granite and normally independent of the principal sulfide deposits; gold, in part, is related to the concentrations of arsenic. Thus, the quartz-arsenic-gold association may be largely, if not completely, unrelated to the sulfide mineralization.

Cobalt is strongly concentrated in pyrite and nickel in pyrrhotite; where chalcopyrite coexists with pyrrhotite and pyrite and contains detectable amounts of cobalt and/or nickel, the chalcopyrite is highly enriched in nickel. Work reported by Rickard and Zweifel suggests that, at equilibrium, cobalt should be concentrated in pyrite and chalcopyrite and nickel in pyrrhotite. They believe that Co:Ni ratios should be more indicative of the degree of metamorphism of the ores than the source of the ore consitituents. Except for Kristineberg, the Skellefte ores are remarkably homogeneous in lead-isotope ratios. They consider that these ratios indicate that the Skellefte ore-leads fall close to Slawson and Russell's best single-stage growth curve for primary leads and are nearly identical with those of Stanton's stratiform ores of marine-volcanic association. These ratios may mean no more than that the leads in such marine-volcanic associations came from the same magma chamber as the volcanic rocks and not that the leads were introduced into the rock sequence by volcano-exhalative processes.

The model ages for the normal leads of the Skellefte district are around 2×10^9 years or much the same age as that determined by Welin for the volcanic rocks.

The delta-^{34}S values for the Skellefte sulfurs range between 1.0 and -3.5, suggesting that they came from a magmatic source in which isotope fractionation of sulfur was small between its inclusion in the rocks from which were derived the magmas that produced the volcanic rocks and the Skellete ore fluids and the deposition of the ore-fluid loads as sulfides in the district. Much thought has been expended on the Åsen deposit in which about

8 per cent barite is present. As would be expected, the barites are enriched in ^{34}S and the associated pyrites in ^{32}S. This probably means no more than that appreciable reduction of Fe^{+3} took place in the Åsen ore fluids, a reaction not of appreciable importance in the other ore fluids of the district.

Chalcopyrite and sphalerite tend to be separated in the ore deposits, but the interpretation of the problem is made more difficult by complex folding, the presence of arsenic-rich ore, and the occurrence of quartz-tourmaline lamprophyre dikes and quartz veins in some deposits of which Boliden is the prime example. Zinc, however, does appear, from observation of the veins themselves, to be more likely to be concentrated near the hanging-wall of a deposit and chalcopyrite in the footwall lenses. Rickard and Zweifel consider such arrangements as similar to what Stanton says are the conditions in stratiform sulfide deposits of his marine-volcanic association. This hardly establishes that the Skellefte ores are of this latter type. No regional zoning of ores has been established. Some zonation has been suggested as being spatially, at least, related to intrusive tongues of the Resund granite. Since the Revsund is younger than the primary ore, such zonation, if it was not primary, must have been related to the metamorphic effect of the granite; the primary zonation explanation seems more reasonable to me.

Zones of alteration are associated with the ore bodies of the Skellefte district. These alteration zones normally are confined mainly to the silicic rocks of the Volcanic series, these volcanics originally have been, in most instances, transformed into sericite or chlorite schists. Locally, almandine, biotite, cordierite, andalusite, staurolite, and cummingtonite were formed in the siliceous rocks; in more calcareous rocks, actinolite-hornblende, andradite, diopside, and epidote may have been produced.

In the usual case, the ore bodies are not centrally located in the alteration zones, but all ore bodies are in close association with such zones. Of the known alteration zones, about 60 per cent are unmineralized or contain only uneconomic sulfide mineralization. Some of these barren zones may once have contained ore bodies, now eroded away, or they may hold ore bodies that are covered at the surface by unmineralized alteration rock volumes.

Gavelin (1939, Boliden-Rakkajour bibliography) considered that the alteration minerals may have been formed by metasomatism and metamorphism, particularly as concerns sericite and chlorite and biotite, although he suggests that some of these minerals may have been original constituents of the now altered rocks. He thought that the rock alteration was later than the metamorphism (another argument for believing that the ore were introduced appreciably after the rocks had been solidified). Of course, it is probably true, as Rickard and Zweifel point out, that the alteration minerals may have been produced in a variety of ways - as primary minerals in the volcanics, alteration by volcanic solutions, low-grade metamorphism while the Volcanic series was being accumulated, local thermal and regional metamorphism due to the intrusion of the Jörn and Revsund granites, and most metamorphic minerals produced by action of the ore fluids. The intrusion of the Revsund granite was accompanied by that of quartz veins and many pegmatites, but it appears that the major amount of alteration was due to the fluids that deposited the ores.

The metamorphic grade of the Skellefte district rocks is quite low except where the host rocks were near granite contacts or zone of migmatization. Gavelin (1939, 1955) recognized a southern chlorite-biotite facies and a northern facies and he stated that the isograd for the reaction series - sericite plus biotite parallels the line of the gneiss front over the entire district. The metamorphic grade increases from the isograd toward the gneiss. The most convincing evidence for metamorphic zoning comes from

Du Rietz work (1953, Kristinberg-Rävlidmyran bibliography) where he saw four zones spatially related to the granite contact: (1) quartz-sericite schists with andalusite (nearest); (2) cordierite-biotite-sericite-quartz schists; (3) biotite-sericite-quartz schist; and (4) chlorite-sericite-quartz schists (farthest away). All that seems certain about the metamorphism is that the grade generally increases toward the gneiss.

Rickard and Zweifel say that the ores basically are stratiform, though, of the four diagrams that they give in their Figure 8, only one of them seems to come near to fitting that description. Nevertheless, it appears that this statement is repeated enough in various papers on the district that there must be more truth to it than the illustrations given would suggest. My own notes, made many years ago, say that the banding at Kristineberg follows the bedding which well may indicate that the Kristeneberg ore is stratiform, but this gives no guarantee that the ore was deposited at the same time as the enclosing sediments. The ore bodies in many places are related, according to Grip, to the series of cupolas (domes) that he identifies in the district. Rickard and Zweifel, however, think that these cupolas are poorly defined in most instances, but where they are recognized by these authors, they are thought to have been formed by the refolding of WNW-striking flexures. Some of the ores are, at least partially, in obvious tectonic structures such as a drag fold at Boliden and an anticline at Rävliden-Rävlidmyran. In general, however, they are convinced that the ores follow steeply dipping "linear structures". These they believe to be composed of drag folds, schistosity, steep tight folds (with their flanks paralleling the schistosity) or a combination of all these tectonic structures.

A general picture of the Skellefte folding is lacking. They argue that many of the deposits are in structures that could have resulted from the transposition of bedding during regional metamorphism. In this process, alternate beds were flexure-folded as a result of the intrusion of the Jörn granite. These folds were ruptured by additional compression caused by the formation of the Revsund granite that gave rise to folding around steeply dipping axes. Locally, the presence of flat-lying fold hinges produced structures in which lenses of ore appear to be parallel to both the primary bedding and to the schistosity. These authors, not unreasonably, except that shear zones would have formed during the transposition of various beds.

Rickard and Zweifel consider the macroscopic evidence for ore metamorphism in the Skellefte district normally to be ambiguous. One example of this is that, at Boliden, the crushed and brecciated arsenopyrite that is cemented by pyrite and chalcopyrite has been interpretated by Odman (1941) as being due to two stages of ore formation, yet Rickard and Zweifel appear to believe that the cementing by pyrite and chalcopyrite may have resulted from the sulfides having been forced into fractures in the arsenopyrite, both breaking and cementing having been caused by metamorphism alone. This explanation might be valid if only chalcopyrite were involved, but pyrite is at least as brittle as arsenopyrite. This problem will be mentioned further in the discussions of specific districts.

These authors also think that the general shape of the ore bodies may have been due to metamorphism. They hold that the ore bodies may have been further drawn out lengthwise by metamorphism even though they were stratiform and lens-shaped at the time of their deposition. On this basis, they think that the relationship between ores and linear tectonic structures may be explained by the metamorphism they have undergone.

Their observations on schistosity imply that the ore and bed rock both have been folded and that different directions of schistosity have been developed in different parts of the folded structures. Although the ores appear (to me) to have been introduced after the Revsund granite, they certainly were, in many places, much broken, if not further folded, by tectonic forces active after the Revsund granite had been introduced. Certainly, Grip's concept is that the emplacement of the ores occurred after the cupolas or domes had been developed in the Revsund granite and that these structures were definitely highly favorable sites for ore deposition. If this is true, either a drawn-out period of tectonic activity must have continued long after the Revsund granite had been emplaced or another and later stage of tectonism must have intervened to cause the brecciation and distortion of the ore bodies. It would seem that Rickard and Zweifel have established that the ores, at least in many places, were folded after they had been introduced. This probably means that folding was developed by the same tectonic activity that accompanied the introduction of the Revsund granite and caused the development of part of the Revsund granite material as actual magma. After the molten part of the Revsund magma had been emplaced and the granite cupolas (domes) produced thereby, structures would be available into which the Revsund-granite-generated ore fluids could enter and deposit the various Skellefte ore bodies. Yet this cannot have been the end of the tectonic activity of which the formation of the magma portion of the Revsund granite was a part. Further folding and faulting must have occurred to produce the structures in the development of which the ore bodies also were involved.

In such processes, some movement of the sulfides would have taken place, with the relatively ductile chalcopyrite leading the way. Rickard and Zweifel's report describes an example of this in which banding in a layer of cupriferous pyrite is equally folded with a well-bedded amphibolite. Chemical analyses show that chalcopyrite is concentrated over pyrite in the hinges of the folds, strongly suggesting, if not proving, that the ore was present before the folding in which ore and amphibolite were equally involved.

Another example is of angular pyrite fragments in massive pyrite ore, and this relationship indicates that tectonic activity not only followed the deposition of Skellefte ore but also took place while ore depositon was going on or continued during a pause in ore formation.

Rickard and Zweifel report that some of the ore was metamorphosed on a microscopic scale. They examined etched pyrite masses, and these almost in all cases were found to be pyrite micromosaics in a manner that showed the pyrite has been annealed. In one case, at Näsliden, they discovered a weak schistosity in the pyrite that was developed by merging the grain boundaries of pairs of similar-sized pyrite crystals normal to the strike of the schistosity. In the Näsliden pyrite, the triple-junction angle approaches the 120° equilibrium quite closely. In other examples, the angles deviate considerably from the 120° value. The annealing and then deformation of the pyrite caused the re-growth of the pyrite to be irregular and much fracturing of pyrite to occur.

Another example shows the pyrite containing large poikilitic porphyroblasts and lacking micromosaic structure. These porphyroblasts are heterogeneous and include a variety of textures of which small groups of pyrite crystals are directly similar to the micromosaic textures of other ore bodies.

From this variety of textures mentioned here (and several other for which space for description is not available here) seem to establish that metamorphism affected the minerals of the Skellefte ores after they had been deposited, but does not definitely establish that the deposition of the ores and the effects of the development of the truly molten poritons of the Revsund granite were not both parts of a single long cycle of ore deposition and metamorphism.

. There seems to be no doubt but that the Skellefte ores were deposited in volcanic rocks produced by eugeosynclinal volcanic activity. It seems doubtful, however, that this necessarily demonstrates that the ores, introduced at a variety of stratigraphic horizons, were deposited on the sea floor and were incorporated in the growing volcanic pile with the passage of time. Rather, the relationship of the ores to structures developed during an appreciably later period of metamorphism indicates that they were introduced by solutions given off from the Revsund granite. The development of these solutions took place long enough after the volcanic activity that produced the Volcanic series for the Phyllite series to have been formed between the end of volcanic activity and the introduction of the ores into tectonically developed structures in the Volcanic series and, to a much less extent in the Phyllite series.

The suggestion that the wall-rock alteration was produced appreciably before the end of the tectonic activity that affected the ores is reasonable enough. It is much less likely that that alteration was caused by exhalative fluids that affected the rocks as they were laid down on the sea floor. It is necessary to agree with Rickard and Zweifel that much further study is needed on the matter of both ore depositon and wall-rock alteration in relation to sea-floor exhalation and to tectonic metamorphism. Arguing that the ores and their alteration is similar to that in the Kuroko deposits of Japan does not prove anything because there is some doubt as to whether the Kuroko deposits were emplaced on the sea floor or in the young volcanic pile after it had been appreciably lithified. This also applies to comparisons of the Skellefte deposits with those of the Bathurst-Newcastle area of New Brunswick.

The concept that the lead-isotope studies show that the ores were formed contemporaneously with the rocks of the Volcanic series but earlier than the Revsund granite also needs further study as much of the processes for determining ore ages from lead-isotope results also needs appreciably more study than it has received. That the Volcanic series, the Phyllite series, and the Revsund granites are part of an almost continuous series cannot be denied, but I cannot believe that the exact position of formation in that series has yet been settled. Grip's evidence that the structures in which the ore has been emplaced were developed during the intrusion of the granite seems to rule out the possibilty that the Skellefte ores were deposited on the sea floor at as many times as the ores are found at different levels in the Skellefte volcanic stratigraphy.

The original character of the ores, though it has been obscured to some extent by tectonic activity after deposition, points to the ore minerals having been deposited under hypothermal conditions in non-calcareous rocks and further deposition took place in the mesothermal range, and, to a minor extent, in the leptothermal. The Skellefte ores, therefore, are here classified as Hypothermal-1 to Mesothermal, plus probably minor leptothermal. The age of the ores, of course is late middle Precambrian.

BOLIDEN-RAKKAJAUR, VÄSTERBOTTEN

Middle Precambrian	Copper, Silver, Gold, Andalusite, Arsenic, Pyrites	Hypothermal-1 to Mesothermal (Leptothermal-minor)

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Notes

Although 20 mines are listed (Grip, 1978) for the area that is included under the heading - Boliden-Rakkejaur - in this Volume, the only mines that are described individually in the literatures are Boliden itself and those of the Målanäs anticlinorium. Unfortunately, the only one (Östra Högkulla) of the Målanäs anticlinorium mines that is included is in the list of the 18 largest producers in the Skellefte district (Grip, 1978, Table 8), and it is in 17th position with a production (through 1974) of only 114,000 tons. Other of the Målanäs mines that are even smaller than Östra Högkulla are Norra Norrliden, Bjurliden, and Bjurträsk.

Of the other major mines in the Boliden area - Långsele (4), Renström (6), Näsliden (8), Akulla Västra (9), and Udden (10) - none is discussed individually in the literature, although all are summarized by Grip (1978) from the extensive source material that must be available in the files of the Boliden Company.

Some information on some of the deposits in the Boliden-Rakkejaur area is provided as follows: Grip (1960) the Renström mine; Kautsky (1960) the Rakkejaur mine. Obviously, this material is old, but it is better than nothing.

The deposit at Boliden was first indicated by the discovery of a huge glacial boulder that weighed about 1 ton. Geoelectrical measurements located anomalies, but these turned out to be caused by pyrrhotite disseminated in graphite schist. Following these, the contact with a phase of the Jörn granite, a synorogenic Svecofennian granite that is younger than the volcanics and older than the overlying phyllites of the sedimentary sequence of the Boliden area, was located. Further anomalies were encountered as the contact was followed to the east, the most impressive of which were 2.5 km north of the name "Boliden" on the official topographic map of the area. [The name "Boliden" was a misprint for "Bjurliden" and was adopted for the name of the location and for that of the mine and of the mining company.] The first drill hole cut through sericite schist heavily impregnated with pyrite; the next penetrated similar material. On 10 December 1924, this second drill hole entered ore, rich in copper and arsenic about 10 m wide; further analyses found this ore to be rich in gold as well. Detailed drilling set the outer limits on this ore body, and by the end of 1925, the upper part of the Boliden ore body was clearly outlined. Despite the thickness of the glacial cover (locally as much as 20 m), electrical prospecting was highly successful. In that same year, enough underground development was completed to permit mining to begin, though on a small scale. Some years later, after much close in drilling, drill holes were commenced well over in the hanging-wall so they would cut through the ore between the 250 and 300 m levels. This made possible detailed study of the hanging-wall phyllite. To permit larger-scale mining, a central shaft was put down in 1929 as far as the 250-m level. During the same period, the overburden was removed from the surface, and an open pit was begun. Both above and below the 250-m level, the ore body was explored at 40-m vertical intervals. Later, additional prospecting was carried out below the ore and outward from it. It was considered that the ore body bottomed above the 330-m level, but the exploration encouraged the continuation of the blind shaft down to the 570-m level with exploration drifts being pushed forward on the 410-, 490-, and 570-m levels. Drifts were driven out from these levels at regular intervals and horizontal drilling also was carried out outward from these drifts. This work found gold-bearing quartz-tourmaline veins and a few masses of sericite-andalusite rock; as a by product of this drifting and drilling, the company had a much better understanding of the complete alteration halo around the deposit down to the 570-m level and beyond.

Further, the company had decided that the contact between the volcanics and the phyllite stratigraphically overlying it deserved attention and further drifting and drilling were done to obtain information about this contact.

The mineralization in the Boliden district, while not actually on the boundary between the two formations (volcanics and phyllites), is, in part (two ore bodies), controlled by a drag fold involving the two formations with the folded volcanics penetrating into the phyllites. These ore bodies are surrounded by the envelope of largely sericitized volcanics so that the ore appears never to be in direct contact with unaltered volcanics. In the center of the drag fold, is a shear zone that stands essentially vertically and appears to form the structure that controls the location of the ore. The zone widens downward, and Grip and Wirstam (1970) are convinced that it provided the channelways for the solutions that first altered the volcanic rocks and second deposited the ore. The process of alteration and ore-deposition may all have been part of one continuous event that produced a steadily advancing zone of rock alteration that was followed, less rapidly, by the metasomatic replacement deposition of the ores.

Grip and Wirstam recognize three stages of ore formation: (1) arsenopyrite mineralization, contained in several ore bodies; (2) quartz-tourmaline and lamprophyre veins in which long zones of gold mineralization that are arranged subparallel to the principal schistosity that strikes about east-west and has a generally vertical dip; (3) pyrite-chalcopyrite ore bodies (two) that enclose the earlier two types; this py-cp ore was emplaced in part by replacing wall rock and in part by filling voids in brecciated older ore. These two ore bodies initially had an en echelon relationship to each other but later faulting (that strikes about NE-SW moved the right-hand side northeast relative to the left-hand one so the result actually was four ore bodies, two large and two quite small since the fault cut through the original two ore bodies near the eastern tip of the western one and the western tip of the eastern one. Prior to faulting, the eastern tip of the western ore body lay north of the western tip of the eastern ore body.

At the contact, south of the ore bodies, the sedimentary (phyllite) rocks dip about 45°S, whereas the volcanic beds dip about 60° to 70°S. Except in the vicinity of the drag fold, the phyllite (or graphitic schist) contact is quite smooth, but in the area of the drag fold, the contact is irregular and exhibits evidence of thrusting and shearing. In a contrary interpretation, this lobe-shaped structure has been reported by Nilsson (1968) to have been caused by a major fault in which the northern side has been dropped down relative to the southern, thereby producing the vertical schistosity of the alteration zone that cuts across the bedding. Nilsson goes on to say, however, that, whether the structure is a drag fold or was caused by a major fault, the fracturing of the relatively competent volcanic rocks created the permeable zone through which the altering and ore-depositing solutions passed. The ore bodies in this zone, and the zone itself, plunge about 60°E and the dip of both is essentially vertical.

The Boliden deposit consists (as has been mentioned) of two ore bodies, the Eastern and the Western; these two have a total length, from west to east, of approximately 600 m and an outcrop area of 10,580 m^2; the maximum width is some 40 m. The area of the compact ore, at the 90-m level dropped to about 8000 m^2, at the 130-m level to about 7200 m^2; at the 210-m level 2200 m^2; and at the lowest level (the 250-m) only 150 m^2. The steep plunge is parallel to that of the drag fold and of the alteration zone.

Ödman, who favored the drag fold concept, considered that the ore-forming process took place in six stages overall: (1) solutions passing through the drag-folded alteration zone produced sericitization of the already somewhat altered volcanic rocks; (2) further sericitization of the original quartz keratophyre; (3) the formation of sericite-andalusite rock during which process, Odman believed that CaO, MgO, and FeO were liberated and passed to the margin of the altered zone to form plagioclase, biotite, and hornblende as a marginal aureole; (4) the introduction of the arsenopyrite ore (under which term also are included rutile rock and pyrite-apatite ore) - these materials were emplaced largely, in Ödman's opinion, by fracture filling; (5) the intrusion of the lamprophyres into fissures formed by continued stress and into which, at a slightly later time, quartz-tourmaline-bearing solutions were introduced that carried in solution, among other elements, boron and fluorine to produce chloritization of the lamprophyres and still further sericitization of the wall rock; in the andalusite rock, these solutions developed corundum, diaspore, and kaolinite; still further effects of these solutions were the production of sulfosalts in which Bi, Te, and Se were prominent; (6) the formation of large masses of pyrite by both fracture filling and replacement; pyrite was the first mineral to crystallize in this ore type and was followed by pyrrhotite in abundance, and chalcopyrite, quartz, sphalerite, and calcite as common minerals. Present-day thinking does not consider the ores to have been introduced in concentrated and viscous solutions. Rather, it appears more probable that the ore-forming fluids were much more likely to have been water-rich solutions in which the ore-forming constituents were no more concentrated than in any other hydrothermal solution at moderately elevated temperatures, low-intensity hydrothermal to mesothermal.

Grip and Wirstam list over 75 minerals as having been recognized in the Boliden ores. Of these, the numbers that are abundant or even common in the various stages is small. The arsenopyrite ore contained only arsenopyrite, pyrrhotite, chalcopyrite, quartz, rutile, calcite, and fluor-apatite in abundance and only sphalerite, cobaltite, gudmundite, tetrahedrite, and falkmanite ($Pb_3Sb_2S_6$) were common. In the quartz-tourmaline stage, the only abundant minerals were tourmaline, pyrrhotite, chalcopyrite, quartz, rutile. The only common minerals in this stage were bismuth (native), tetradymite, tellurobismuthite (Bi_2Te_2S), cobaltite, galena, "selenocosalite" [$Pb_2Bi_2(S,Se)_5$], "selenokobellite" [$Pb_2(Sb,Bi)_2(S,Se)_5$], scheelite, calcite, fluor-apatite, sericite, and mariposite. In the pyrite stage, the abundant and common minerals are listed in the paragraph above.

The altered rocks have been named by Ödman, who was followed by Nilsson, to give nine different varieties to which the volcanic wall rock of the ores has been converted. These are: (1) sericite-quartz schist (normal), (2) sericite-quartz schist (gray), (3) chloritic sericite-quartz schist, (4) pyrite-banded sericite-quartz schist, (5) andalusite-bearing sericite-quartz schist, (6) silicic sericite-quartz schist, (7) sericite-quartz schist, (8) sericite rock, and (9) andalusite rock.

Of the two major ore bodies at Boliden, the western one is the larger at the surface, where the total length, including the small-faulted-off portion at the east end, is 365 m; the maximum width is 16 m. The strike averages about east-west, but, at the western end of this body, the strike bends 20° toward the south, as also is true for the complex of rocks in which it is enclosed (Ödman, 1941). The dip of the ore at its western end is 70° to 80°S; the central portion is almost vertical, although at depth it is 80°N. The ore-body contours are irregular, and it seems not to have any definite pitch, except locally where it may pitch steeply to the east. The fault that cuts off the east end dips rather flatly down to the 90-m level and has moved the eastern block obliquely upward for about 30 m along the fault plane.

The wall rock of this ore body is normally sericite schist, although some of it is rich enough in pyrite to be designated pyritic-sericite schist, particularly in the upper parts of the ore.

The eastern ore body contains more ore than does the western, most of which is of the pyrite type; its length on the surface is 357 m, including the part faulted off to the south. Its maximum width is 37 m. The strike of the ore is about E10°N on the upper levels but is almost east-west on the lower ones. The dip is quite regular, being between 70° and 85°S down to the 130-m level. Below that, down to the 170-m level, it dips nearly vertically and below that 170-m level, the dip is 70° to 85°N. On both eastern and western ore bodies, the dips of the surrounding schists parallel those of the ore bodies.

The keel of the eastern ore body pitches 55°E; farther up, the pitch is steeper and may be as much as 70°E. The wall rock is sericite schist that, in places, contains chlorite or pyrite.

Both ore bodies contain arsenopyrite, pyrite, and pyrrhotite ore, with the west ore body containing much more of the latter, centrally located in the ore body; the eastern ore includes lamprophyre dikes and quartz-tourmaline ore to a greater extent than does the western ore. For all practical purposes, it appears that the massive ore died out by the 250-m level; even so, several lower levels were opened up with the 570-m level apparently being the lowest, but no ore seems to have been found for all this work. Drifting and cross-cutting and underground drilling also were continued far beyond the ore bodies in a mainly east-west direction, also without locating anything more interesting than a vile-smelling gas that contained enough helium to cause the management to consider, but never carry out, its recovery.

At the 250-m level, the eastern ore body is nearly gone, only three small patches remain on that level; all three are completely enclosed in sericite rock. Below the 250-m level, although no massive ore has been found, the alteration continues to be as strong and varied as it was higher up. At least, this demonstrates that the altering solutions had a much wider effect than those (same?) solutions when they later were depositing the various types of ores.

From the detailed and colored maps in Ödman's study on Boliden (1941), it is possible to see clearly the relationships of the various ore types to each other. In particular, it appears obvious that the isolated bodies of arsenopyrite ore in the major pyrite ore masses must have been introduced first. The introduction of the ore solutions that formed the pyrite ore certainly were able to work their way through the arsenopyrite ore, replacing it by the pyrite type. Of course, it is difficult to say, even with the examination of huge numbers of polished specimens, how much of the pyrite ore replaced the arsenopyrite variety and how much replaced sericite schist. The presence of large volumes of disseminated pyrite in sericite schist (or rock) bordering the pyrite ore suggests that a considerable fraction of the pyrite ore was emplaced by the replacement of sericite rock. Thus, it can only be stated that pyrite ore was emplaced in two ways: (1) by the replacement of arsenopyrite ore and (2) by the replacement of sericite rock. It must be remembered that included with arsenopyrite ore are rutile rock and pyrite-apatite ore.

The quartz-tourmaline ores are the most peculiar of the Boliden ore family. The lamprophyre dikes in many places include lenses of quartz (ladder veins); these normally strike NE-SW or ENE-WSW and pitch about 50° to 60°E (the general direction of the pitch of the ore bodies). The quartz, again only in places, contains tourmaline, plus pyrrhotite, pyrite, chalcopyrite, and sulfosalts. Locally, the ladder veins include some gold. This gold is commonly associated with the sulfosalts, particularly with selenokobelite. The gold-bearing quartz veins are contained in sericite- and sericite-chlorite schists; these include small schlieren of much deformed sericite rock. Gold also can be as much as several decimeters outside the actual quartz-tourmaline veins, being emplaced as along the planes of schistosity as thin gold leaves.

Some lamprophyre dikes exist without tourmaline veins and a few tourmaline veins are not directly associated with lamprophyres, nor do all quartz or tourmaline veins contain gold.

One area of high concentration of gold-bearing tourmaline-quartz veins was slightly north of the west end of the Eastern ore body, and their strike was controlled by that of the schistosity of the surrounding altered volcanics. On the 250-m level itself, the strike of the veins was determined by the old land surface and was around N60°E. As the tourmaline veins were followed upward, the strike gradually shifted to about east-west. This area was, of course, in the immediate neighborhood of the major fault that offset the tips of the two main ore bodies, and one result of this was the presence of several small faults and fissures. The portions of the tourmaline that contained mineable gold were between the 270- and 170-m levels. Other areas of such high-gold content were elsewhere in the Eastern ore body. In more than 50 per cent of the scheelite occurrences, they were in or near quartz veins; the remainder of the scheelite localities were in unaltered rocks and even in massive ore. In such altered rocks as contained scheelite, that mineral was associated with sulfides. The largest concentration of scheelite was in the westernmost portion of the Eastern ore. In such sulfide-rich areas, the scheelite was in the pyrrhotite-rich matrix of the brecciated arsenopyrite ore; it also could be in pyrrhotite ore that was introduced as part of the lamprophyre-tourmaline ore stage. On the contrary, scheelite was not a constituent of solid arsenopyrite ore nor in the late massive pyrite ore, thus demonstrating that the pyrrhotite ore was part of the lamprophyre ore stage. The scheelite in pyrrhotite ore, therefore, also is part of the second ore stage, and, in this ore, quartz-tourmaline ore locally grades into pyrrhotite ore with that includes varied amounts of chalcopyrite; the sulfides brecciate the tourmaline ore. Such pyrrhotite-chalcopyrite ore is known in parts of the mine where it is unconnected with tourmaline-type ore, but Ödman believed that their paragenetic relations indicated that they were part of the second stage of ore introduction and not temporally connected to the third (pyrite) stage ore.

At one location (at the west end of the Eastern ore), 230-m to 232-m level the scheelite ore was so rich that mining of it was considered; this ore also was exceedingly rich in gold, and the scheelite contained a small percentage of molybdenun (0.05% Mo). No other molybdenum-bearing minerals were found in the Boliden ore bodies.

Some of the quartz-tourmaline bodies are not veins but have a lens shape. Such lenses normally are nearly all tourmaline but may contain quartz, sericite, and several ore minerals. Some of the lenses have sufficient vertical extent that they are more reasonably considered to be pipes. The contacts of these lenses with the wall rocks are sharp, the ore following the foliation and normally lacking any cross-cutting effects.

The pyrite-chalcopyrite mineralization that ended the period of ore formation at Boliden is far less unusual than are those of the first and second stages of ore mineralization. An explanation for the materials that formed these three stages has not been readily forthcoming. Ödman's theory that the materials of the arsenopyrite stage were introduced as a "speiss" (in metallurgy a mixture of impure metal arsenides formed as a melt beneath the slag in certain smelting processes) now seems to have been discarded with the ores of stage one having been introduced in pneumotectic solutions (gaseous products the late-stages of a crystallizing magma). The possibility of solutions or mixtures of gases developing as products of magmatic differentiation is, because of the high pressure obtaining in any magma, no matter what its manner of origin (complete melting, palingenesis, anatexis), essentially impossible. Although the water-rich material separated from the magma would technically probably be a gas (as it would be above its critical temperature), it would be so dense that it would behave as would a true liquid. Because of the essentially unique character of the arsenopyrite ores at Boliden (and at a few of the other mines in the Boliden-Rakkejaure area), the differentiation process that produced the ore fluid for the

Boliden stage-one ores must have been unusual, either in the composition of the differentiating magma or in that of the ore fluid so produced. Certainly, not enough has been done in any way to explain where or how the magma from which the stage-one ore fluids came was produced or how it differentiated or of what the actual ore-forming fluid was composed except to say that it was water-rich and contained the constitutents (in some form or other) of which the stage-one minerals are made up is not enough. Nevertheless, from the minerals that stage-one ores contain, it is almost certain that the hydrothermal ore fluid from which they were deposited was operating in the hypothermal range, in the large part, although the sulfosalts found in this ore in occasional schlieren probably were formed in the mesothermal to leptothermal range. Even galena is present only in association with these sulfosalts and appears to have precipitated after them.

The quartz-tourmaline [QTL ("L" stands for lamprophyre)] ores also seem to have been formed largely in the hypothermal range. The dominant diagnostic mineral, tourmaline, certainly formed under high-temperature. The later pyrrhotite, which brecciates and replaces the tourmaline, also appears to indicate deposition in the hydrothermal range. In contrast to the first-stage ores, the QTL ores contain considerable quantities of galena; it is one of the latest minerals. The chalcopyrite, locally of major importance, is later than quartz, pyrite, pyrrhotite, and tetrahedrite (and of course later than tourmaline). On the basis of these relationships, the chalcopyrite probably was deposited, to a considerable degree, in the mesothermal range. All of the sulfosalts, except tetrahedrite, and plus galena and gold, are later than the first generation of chalcopyrite, so some minor fraction of the QTL ores were precipitated in the leptothermal range. A minor second generation of chalcopyrite may indicate a minor return to the mesothermal range.

The third-stage pyrite ore is, of course, dominated by pyrite, in small grains that are normally without crystal form, although some euhedral cubes have been noted. Pyrite is replaced by almost all the other minerals of this ore type. Chalcopyrite is the more common economically valuable mineral of this ore-type. Although most of this ore is associated with pyrite, some pyrrhotite-chalcopyrite ore is present in marginal bodies.

The pyrite almost certainly was emplaced by the replacement of previously deposited minerals of the first and second ore stages or of the altered volcanic rocks in which the earlier ore stages were emplaced.

Pyrrhotite and sphalerite were about contemporaneous. Ödman considers that the arsenopyrite was the result of remobilization of mineral from the first-stage ores. It seems equally likely that the replacement of arsenopyrite during the deposition of pyrite would have raised the arsenic content of the ore fluids to such a degree that arsenopyrite could be precipitated as a late mineral in the third-stage mineralization. Some arsenopyrite is present as small euhedral crystals enclosed in pyrite; even Ödman concedes that this is a primary constituent of the third-stage of ore formation. Stannite is quite ubiquitous in the Boliden ores; there it is confined in its associations to sphalerite, pyrrhotite, and chalcopyrite. Chalcopyrite replaces stannite; sphalerite is replaced by stannite; the relations of this tin mineral to pyrrhotite are uncertain. The stannite is no more than a rare mineral, but it indicates that, at the time it was being deposited, the ore-forming fluid was operating in (probably) the upper mesothermal range. The pyrrhotite and sphalerite (the latter common in the third-stage ores) suggest that they were emplaced under low-intensity hypothermal or high-intensity mesothermal conditions with the intensity level dropping off slightly to upper mesothermal during the deposition of the later chalcopyrite and stannite.

Only occasionally do the third-stage ores contain schlieren of sulfosalts; these probably are common enough to justify the addition of leptothermal to the classification of these ores.

This summary of the geology of the Boliden ores cannot, by any stretch of mind, be called complete. The complexities of the various ore stages and the difficulties of relating what went on during one stage to the conditions obtaining in the other two are major. Boliden now appears here to be a closed book by Swedish ore geologists with Grip and Wirstam (1970) having written the obituary of this remarkable deposit. Surely, however, there is enough material in various collections in Sweden that more work could be justified.

Because of the paucity of published literature on the other deposits in the Skellefte region in general and the Boliden-Rakkejaur area in particular, the ore geology of only five of the other mines in the area is to be summarized here: these are: Långsele, Renström, Näsliden, Rakkejaur, and Mensträsk. The first three of these are among the eight largest in production through 1973; Rakkejaur is the 12th largest in production and is the largest in mineralized volume; and Mensträsk is particularly interesting as to its geology, although Grip does not even mention it among those for which he gives production figures. All five of them contain some early arsenopyrite, but none of them is as rich in that mineral as Boliden. Boliden contained 6.8 per cent of arsenic; the next highest of the remaining four was 1.6 per cent at Rakkejaur. Of the five ore bodies, other than Boliden, only one, Långsele, may questionably contain any quartz-tourmaline mineralization associated with lamprophyre dikes. Separate studies are available on the Mensträsk mine but not on the others.

The Långsele deposit is only 5 km southwest of Boliden, so close to that mine that the Långsele ore was hauled underground to the Boliden concentrator for processing. The deposit was discovered in 1926 after boulders had been traced to the mine site and electrical measurements had been made. Underground study did not begin until 1947 and continued until 1955 with the actual mining beginning in 1956. The mine had been worked up to about 1978 (Grip) and probably continues in production to the present time. In 1973, it had the fourth highest total production in the Boliden area.

The total outcrop area of the ore at Långsele is about 3600m^2 and consists of five ore lenses; the deposits extends over a distance of 360 m and, in 1978, had been followed down for as much as 400 m, and the bottom had not then been found. The ore lenses strike northeast and dip 50°SE; they plunge 50°S. The average grade of the ore (probably at the end of 1973) was 0.8 ppm Au, 25 ppm Ag, 0.4 per cent Cu, 4.3 per cent Zn, 0.3 per cent Pb, 0.4 per cent As, and 35 per cent S. The tonnage mined in 1973 was just over 350,000.

Whereas the Boliden ore is slightly below the top of the Volcanic series, the ore at Långsele is in the Transition zone between the Volcanics (beneath) and the Phyllite series (above). The ore bodies are located in a synclinal fold axis that plunges 50°S. This fold was developed where the contact zone intersects a zone of north to NNE schistosity with the five ore lenses being in part controlled by the bedding planes and in part by the schist structure. As Grip puts it: "most of the lenses appear to hang vertically from the overlying schists".

Both the volcanics in the footwall of the ore and those that enclose the ores are quartz porphyrites or felsites; these are overlain by: (1) a thin layer of black phyllite on the bottom and (2) several 10s of meters of dacites or sediments of dacitic composition. The probability is good that there is a discordance between the rocks of the Transition zone and the phyllites that overlie them. The Volcanic beds are cut by many greenstone dikes, containing abundant amphibole, and with these dikes are many galena-bearing quartz veins and a great many sulfosalt minerals. These dikes and the metalliferous minerals they contain are older than the dominant pyrite ores. These dikes and the minerals associated with them probably (Grip, 1978) are the first stage (although they do not constitute actual ore) of several stages of ore formation, and probably are the equivalent of the lamprophyres in the Boliden mine.

The ore itself is a zinc-bearing pyrite that contains small amount of copper and lead. In addition to the pyrite, the ore minerals are pyrrhotite and sphalerite with even lesser quantities of chalcopyrite, galena, arsenopyrite, and locally sulfosalts. The sphalerite contains appreciable mercury (the content of 250 ppm Hg is the greatest in the Boliden district), that element being in the lattices of the sphalerites. The district underwent several stages of ore formation with the greenstone dikes, as was mentioned above, being the first. The second stage was the complex ores just described, and the third stage was made up of galena,sulfosalts, and quartz in dikes and schlieren.

In the footwall of the ore is an alteration zone of sericite schist that extends from the contact between the ore lenses and the phyllites for about 1 km following the zone of schistosity to the NNE. This footwall both contains appreciable disseminated pyrite and some schlieren of that mineral that are several meters in width. This alteration zone also includes some chlorite schist in which andalusite and cordierite are present. The contact of the ore with wall rock is quite sharp, although the massive ore in places cements wall-rock breccia.

The Renström area is located west of a north-trending tongue of Revsund granite and north of a rather small mass of Jörn granite that is disconnected (at the surface at least) from the main Jörn body farther north and lies in a "V" of rocks of the Volcanic and of the Phyllite series enclosed between these two granites. The Renström area is about 15 km almost due west of Boliden. Grip reports (1978) that the structure of the area is very complex. The first structural event was a gentle folding of the rock around horizontal axes that have a WNW direction. The second event was a much stronger folding around vertical axes that resulted from the intrusion of the Revsund granite. The whole mineralized area was strongly altered to a mass of sericite-chlorite schist; this alteration is most intense nearest the Revsund granite and becomes less impressive as it is followed to the northwest. The quartz-porphyry tuffs and agglomerates in which the ores are contained have, near the Revsund granite, been migmatized (Grip, 1978) and have thus acquired considerable cordierite, biotite, hornblende, chlorite, and sericite. Only a short distance to the northwest, the high-temperature minerals of the alteration aureole die out, leaving only sericite and chlorite, particularly in zones of high schist development.

The actual Renström deposit is made up of three groups of ore bodies, imaginatively named the West, the North, and the East, that are located in the vicinity of older quartz-porphyry dikes. The West ore bodies are economically the dominant group; there are two of them that have a normally north-south strike and plunge nearly vertically. The more important of these (the A lens) has been traced (1978) to a depth of 850 m or more. The trend of this ore body follows the bedding in the polymict clastic series that contains intercalations of schist and carbonate rocks. The zone of schistosity in these rocks apparently provided the channelway through which the ore-forming solutions moved upward and passed outward into the country rocks. The B ore body lies to the south of the A and is extremely irregular, in fact only a small narrow pipe reaches the surface. Below that outcropping neck, the ore body is confined to a zone of schistosity and brecciation that strikes NNE; the lower part, however, is parallel to the A ore body.

In the northern group, the Kyrkväg ore body apparently is the only one worth mentioning; it is located to the NNE of the A ore body but is on the opposite side of a quartz-porphyry dike.

The Renström East ore body is some 600 m east of the B ore body; since it strikes, and its enclosing schistosity is, east-west, it has a quite different orientation from the A ore body; it plunges at 70°E. The rock in the zone of schistosity containing the ore body is sericite-chlorite quartzite.

The total area of all the Renström ore bodies is about 1000 m^2. All the ores in the Western ore body are much the same in composition and character; the Kyrkväg ore is lower in copper than the West ore, and the Renström East is appreciably lower grade.

The country rocks of the Western ore bodies initially were silicic sediments with intercalated carbonates. This rock sequence has been converted to a sericite-biotite quartzite that retains the calcareous bands; within the ore zone itself, a major magnesian alteration has produced dolomitic quartzite, calcareous quartzite, and chloritic quartzite. Locally, these altered rocks may be high in serpentine, confirming the concept of an impressive magnesian alteration. The stratigraphic position of the Renström ore is not certain. Actually it is in highly altered volcanic rocks near the contact with phyllites, but these phyllites (or black schists) are present only in lenses; the Phyllite series as such probably is absent from the Renström area.

The two ore bodies of the Western zone - A and B - consist of a massive, fine-grained ore; the grains in only a few places are larger than 0.05 mm. The ore is strongly banded; these bands are something like 1 cm in width and include large contents of pyrite, chalcopyrite, and sphalerite. The ore may contain fragments of wall rock of varied sizes and high in quartz and dolomite, suggesting an appreciable brecciation of the country rock before ore formation; the contacts of the ore with unbroken wall rock are, in most places, sharp. The banding of the ore suggests that certain of the original bands in the host rocks were more favorable to replacement than others. Locally, moderate-sized apophyses that contain mainly chalcopyrite extend out from the ore into the wall rock. The main ore minerals are (in order of abundance): pyrite, sphalerite, chalcopyrite, galena, pyrrhotite, arsenopyrite, tetrahedrite-tennantite, pyrargyrite, gold, and even less common accessories. The mineral character of the Kyrkväg (North) ore is much the same as that of the Western ore bodies; the grade of the Eastern ores is appreciably less than that of the other two ore-bearing areas.

The Eastern ore bodies do not have the sharp ore-wall-rock contacts of those in the other two areas, and this Eastern ore grades out into the wall rock as disseminations of pyrite and minor chalcopyrite.

The Renström Western ore has (up through 1973) an average grade of: 3.9 ppm Au, 184 ppm Ag, 0.9 per cent Cu, 11.7 per cent Zn, 2.0 per cent Pb, 28 per cent S, and 0.25 per cent As. The mine began operation in 1952 and had produced nearly 4 million tons of ore through 1974. 1973 production was about 200,000 tons.

The Näsliden ores are contained in the Näsliden dome that is composed of volcanic rocks - felsite, keratophyre, quartz-keratophyre, and dacite, plus pyroclastic rocks of similar character. In the southern part of the dome, the volcanics are overlain by the sediments of the Transition zone; these are more calcareous than is usual and are of quite varied grain size. These sediments are rather different to the north, being thin or, at best, of uneven thickness. In the Rakkejaur area (to the northeast), however, the Transition sediments reappear, having been deposited on an eroded land surface. These Transition beds are overlain by those of the Phyllite group.

The entire Näsliden dome area has been folded about horizontal axes to produce an anticlinorium; this structure changes both to the southwest and northeast into synclinoria in which the major rock type is phyllite. Locally, in the crests of the synclinal folds, the older Volcanic-group rocks are exposed. When the later igneous rocks - mafics and granites - were intruded and the ore-forming fluids introduced, a secondary folding, related to nearly vertical axes, also produced shearing and lineations.

Two of the deposits discussed here are located on the Näsliden dome, Näsliden proper on the southwest flank and Rakkejaur on the northeast.

The Rakkejaur ore deposit is the largest in the Skellefte district, although (through 1973) it was only 12th in total production, but it does have an outcrop area of 19,290 m^2, is 550 m long, and is as much as 65 m wide. Drilling, through the early 1970s indicate that it extends downward for at least 320 m and is stock-like in form. The average assay results through 1973 are: 1.0 ppm Au, 45 ppm Ag, 0.3 per cent Cu, 2.3 per cent Zn, 27 per cent S, and 1.6 per cent As.

Rakkejaur is another Skellefte ore body that lies in the Transition zone between the Volcanics (below) and the Phyllite (above); the ore body follows the Transition zone-Phyllite contact. The footwall on the west side of the more-or-less north-striking ore body dips vertically and is cut off to the north against a drag fold oriented about an axis that plunges 80°NNW. This footwall is made up of a quartz porphyry that is now a sericite schist. Stratigraphically above the quartz porphyry is a weathering breccia composed of calcareous materials, plus a polymict conglomerate that locally contains granite boulders. Above these beds are banded phyllites that are, in places, directly in contact with quartz porphyry and sericite schist.

The ore of the deposit is low-grade, mainly pyrite that includes various amounts of zinc-, copper-, and arsenic-bearing minerals. The arsenopyrite is present in several lenses and the chalcopyrite is a few. Except for disseminations passing into the footwall in the southeast of the deposit, the ore-wall rock contact is quite sharp. The arsenopyrite always is fine-grained or even dense. The lenses of this mineral are quite well-developed in the upper part of the ore body, but, at depth, they become much less common where they occur only in a few schlieren. They correspond to the arsenopyrite ore at Boliden as they are the first stage of ore formation. Their failure to be found at greater depths suggests that the Raffejaur ore body has been more cut down by erosion than the one at Boliden; certainly the Rakkejaur content of arsenic is only about one-fourth that at Boliden.

The pyrite ore body is massive only rarely and usually contains a great deal of (probably unreplaced?) country rock, much of which is as breccia fragments. Replacement affects the coarser-grained rocks more effectively, particularly the weathering breccia and the conglomerate; some schists and phyllites also were replaced, and appreciable silicification occurred during that process. Such sphalerite, arsenopyrite, and galena as are present in the ore are there in schlieren. Within the pyrite are local, fine-grained disseminations of chalcopyrite, plus even more restricted amounts of pyrrhotite.

In addition to the copper ore in the pyrite, the third stage of ore formation produced copper ore as irregular bodies elongated NW-SE; these are of rather consistent thickness in vertical section. The age of this copper ore is definitely younger than that of the pyrite ore because both the pyrite ore and inclusions of country rock were brecciated prior to the introduction of the copper as it cements the fragments of this breccia type. The copper ore is almost entirely chalcopyrite and pyrite with some minor pyrrhotite and even less sphalerite, arsenopyrite, and tetrahedrite. Grip believes that position of the copper ore at an acute angle to the strike of the enclosing beds suggests that this ore was developed during continued counter-clockwise rotation that produced gash veins that provided space into which most of the copper ore was precipitated.

A fourth stage of mineral emplacement resulted in the emplacement of many carbonate dikes in which some chalcopyrite and such sulfosalts as boulangerite and tetrahedrite are present.

The country rock, particularly on the west (footwall) side of the ore body, normally is sericitized with the porphyries and conglomerates being most altered; locally some chlorite is developed but no biotite or amphibole.

The Näsliden deposit, on the southwest side of the Näsliden dome, is somewhat unusual in that its ore horizons are contained in both the rocks of the Transition zone (below) and the Phyllite group (above). The underlying felsites of the Volcanic group are several hundred meters thick and are overlain by the Transition zone of interbedded phyllites, sandstones, and conglomerates, all of which are highly calcareous; within this zone are several ore horizons. The overlying Phyllite group also is calcareous and contains ore horizons near its base.

The rocks enclosing the Näsliden deposits have been metamorphosed only to a low level, being of the sericite-chlorite facies. The ore has favored replacement of the graphitic phyllites, and many partially replaced fragments of this rock can be found in the ore.

The major structure in the Näsliden ore field is an anticlinorium, on the southwest flank of which cross folding produced secondary drag folds about axes that plunge 70° with a trend of S60°W. These secondary folds caused the development of shear zones that are oriented N20°W and N7°W; both dip west at 70°. These shear zones are localized in the Transition zone between the volcanics (below) and the Phyllites (above). During a later counter-clockwise rotation, ore was emplaced along the N20°W shear planes with repeated movements taking place along the N7°W shearing. Most of the ore was deposited in the calcareous zone; Grip thinks that this was due to its greater deformation that made it a most effective channelway for ore-fluid movement. He does not mention the possibly greater ease with which the carbonate material could be replaced as opposed to the silicate materials in the Transition zones.

The only mine in the Näsliden area (at least through 1973) was Näsliden Södra, the ore of which is contained in the highest portion of the calcareous zone that lies next below the Phyllites. In the upper part of the mine, there are three ore bodies arranged in an en echelon pattern; these bodies are as much as 160 m long and some 20 m wide; the total area of the three is 3500 m^2. The (unexpressed) suggestion is that the three ruler-shaped bodies join with depth, but Grip does not actually say so. The ore is a massive sulfide material in which pyrite, sphalerite, pyrrhotite, arsenopyrite, galena, sulfosalts, of which tetrahedrite is the most important of the constituents. The massive ore is surrounded by a halo of disseminated ore that contains the same sulfide minerals as the massive material. The Näsliden Södra mine began operation in 1970, and, by the end of 1973, over one million tons of ore had been mined, and the ore had been developed down to the 350-m level.

Näsliden Norra appears to have been developed but not mined (at least prior to 1973) but consists of five lenses that are contained in as many separate horizons that are immediately above the felsites. Of these lenses, the biggest is 300 m in length and the smallest 100 m. The total area of the lenses is 2100 m^2. The ore here assays: 0.6 per cent Cu, 1.7 per cent Zn, 0.3 per cent Pb, 29 per cent S, and 1.3 per cent As. The structural control of these ore lenses is much the same as at Näsliden Södra, and the minerals present are essentially identical except for the presence of appreciable magnetite in the Norra deposit.

The Mensträsk (träsk = swamp in Swedish) ores are located at 64°04'N, 19°22', although they are included in an area 3 km long. All but the two ore bodies farthest east are under Lake Mensträsk, and no one of them outcrops; the 3-km zone strikes NW-SE. The six ore bodies are, from northwest to southeast: Outer Lake Ore, Inner Lake Ore, Rundklumpen, Lomviken, Rågången, and Långviken. The Inner Lake Ore (Inre Sjömalmerna) appears to be the only one that has been mined, although Grip (1951, 1978) does not actually say that it has been, and he gives no production figures for it.

The Mensträsk deposit (the Inre Sjömalmerna is also known as the Mensträsk mine) is under Lake Mensträsket just west of a peninsula pointing southwest from the east shore. The Mensträsk deposit is located on the Mensträsk-Elvaberget anticlinorium that is in the center of the Skellefte district, is some 40 km long (WNW-ESE), and 10 to 20 km wide. Mensträsk is at the west end of this structure; the next ore deposit to the WNW is Rakkejaur on the Näsliden dome, and the Elvaberget deposit is 7 to 8 km almost due south of Mensträsk. Underground development was completed at Mensträsk in 1948; then (as Grip says, 1978) "no mining (was contemplated) in the near future and it (was then) uncertain when the mine (would be) accessible again". Therefore, Grip published the results of his study of it in 1951. As the grade for the Inre Sjö deposit averaged 0.5 per cent Cu, 2.2 per cent Zn, and 31 per cent S, it probably would not be surprising if the Boliden Company had done nothing more with the deposit.

The Mensträsk-Elvaberget anticlinorium extends slightly south of east from Mensträsk area through the Målanäs area to the Kedträsk area at the east end of the structure. In several of the synclines that lie between the anticlines of the anticlinorium, the rocks of the Phyllite group are exposed. To the north, the rocks of the Volcanic group dip beneath the sediments and andesites in a shallow syncline. In the south, these same rocks pass under a boundary zone where the sediments, locally, are isoclinally folded. In the center of the major structure, a tongue of Revsund granite cuts into the southern phyllite syncline (or synform) and causes a depression in the fold axes of the various structures. In the southeast of the major structure, the boundary is not easily determined because of a considerable volume of the rocks involved have been migmatized. Later cross-folding has broken up the major structure into several domes that Grip thinks were major factors in the location and emplacement of the various ores deposits connected with the anticlinorium (the Mensträsk, Malånäs, and Kedträsk areas, from west to east).

The isoclinal folding in the northeastern, ore-bearing portion of the Mensträsk area has produced steep dips; in the area south of Lake Mensträsket the dips are flat, but, to the south of the Malån River, the dips again become steep. A late secondary folding has developed drag folds as well as steeply plunging lineation; further, conglomerate pebbles and boulders have been elongated in the direction of lineation. The ores of the Mensträsk area are made up of more or less compact sulfides, and all the ore bodies are contained in the zone of coarse sediments and weathered breccia that is located between the felsites (below, stratigraphically) and the phyllites (Grip prefers to call them slates). Some minor mineralization also is present in the felsites in those places where they have been brecciated. All five of the ore bodies occur in the steeply dipping beds on both sides of the anticlinorium. Some sulfide mineralization, however, if present in flat-lying breccia south of Lake Mensträstet, but it is the dissemenated type found in the brecciated felsites on the north side of and under the lake.

The ore in the so-called Outer Lake Ore is oriented along the steep middle fold axis of the drag fold, this structure being favorable for ore emplacement. The Inre Sjö ore bodies pitch 45°NNW and are contained in a conglomerate that dies out to the northwest against what Grip considers to be an erosion hill on the old land surface. This pitch of the ores corresponds to the direction in which the conglomerate pebbles were elongated. The remainder of the five ore bodies in the area were too little drilled to provide more than a scanty idea of their character.

The Inre Sjö deposits is made up of six lenses of massive ore that extend over an area 300 m long and 70 m wide; at a depth of 30 m, the total area of ore is 4190 m^2. The strike of these lenses is N50°W, the dip is 70°NE, and the plunge is 45°NNW.

Grip considers that the location of these ore lenses resulted from several tectonic factors. In the first place, the conglomerate that contains the ore is located between two beds of markedly different competence. The older bed is brittle quartz porphyry and the younger is the rocks of the Phyllite group. When these rocks were subject to stress, the difference in their responses to it caused the conglomerate between them to be brecciated; this breaking was unusually intense because the old land surface below was very uneven. This surface rose to the northwest; therefore, the conglomerate wedged out in that direction, making it even more subject to the effects of differential movement. Cross fractures thus developed in the conglomerate served, Grip is certain, as feeders for the passage of the ore fluids that produced the alteration around the ore and the ore itself. The ore bodies are grouped along the margins of the wedges of conglomerate breccia; these plunge 45°NNW, and the ore bodies not unreasonably plunge in the same direction.

The wall rock has been strongly altered by the development of sericite, chlorite, and calcite. The presence of fragments of sericite and chlorite schist in the massive ore indicate that at least a considerable amount of alteration had taken place before the ore was emplaced.

The altering solutions apparently changed with time and achieved a composition that resulted in ore deposition. The first ore sulfide to deposit was arsenopyrite, and it occurs in schlieren in the ore body; these are typical of the entire district. The arsenopyrite, however, locally has heavily impregnated the host rock. After the arsenopyrite, pyrite began to deposit and was followed by far less abundant sphalerite, pyrrhotite, and chalcopyrite, some of these sulfides appear to have replaced the pyrite. A modest group of sulfosalts followed the ore sulfides but as a later phase of the crystallization. The ore minerals are contained in schlieren that lie parallel to the schistosity of the host rock. The ore minerals in the schlieren are banded with largely pure pyrite alternating with bands that are heavily impregnated with sphalerite, pyrrhotite, and, locally, sulfosalts.

Much of the Mensträsk mineralization is like that at Rakkejaur, thus: (1) the lenses of arsenopyrite die out downward; (2) pyrite ore with various contents of the ore sulfides brecciates and replaces the host rocks, especially the coarse clastic rocks but also the quartzites and phyllites. Structures in the ore show that this ore deposition took place during the formation of the drag fold; the slate around the breccia zone aided ore formation by impounding or greatly slowing down the speed of ore fluid movement; (3) the chalcopyrite ore is in irregular patches, though these are somewhat elongated in a NW-SE direction and are much more continuous vertically than they are horizontally. The chalcopyrite cements brecciated pyrite and is confined entirely to the massive bodies of that mineral; (4) masses of calcite, chalcopyrite, and sulfosalts make up the fourth stage of ore formation - these are what Grip calls ore pegmatites though why is not quite clear; they may resemble some pegmatites in texture, but they certainly were not formed from molten material; (5) the stratigraphical position of the Mensträsk ores is the same as that of several of the deposits in the Skellefte region, such as Boliden and Långsele that have been discussed here.

Nothing that has been said here about the five other ore deposits of the Boliden-Rakkejaur area justified giving them a different classification or date from that assigned to the Boliden ores.

KRISTINEBERG-RÄVLIDMYRAN, VÄSTERBOTTEN

Middle Precambrian | Copper, Pyrites, Gold, Zinc | Hypothermal-1(?) to Mesothermal

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Gavelin, S., 1943, On the distribution of metals of Rävliden, northern Sweden, and in some other copper-zinc ores: Sveriges Geol. Undersök., ser. C, no. 454, Årsbok 37, 34 p.

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Notes

The center of the Kristineberg-Rävlidmyran district is located S75°W from the Rakkejaur mine at a distance of about 32 km. The mines of this district are related spacially and structurally to the dome-shaped Kristineberg anticlinorium. Slightly north of west from this structure is the Vindelgransele anticlinorium that much resembles that of Kristineberg except that it has only two minor sulfide occurrences although disseminated sulfide showings are common there. In 1973, the total production from the Kristineberg mine was almost 10 million tons, over 1.5 million more than were recovered from Boliden over its entire life span; this dominance of Kristineberg over Boliden (in tonnage, though not in grade) has increased since 1974, because Kristineberg still is being mined and Boliden, of course, is not. Grip reports the average grade of the ore from Kristineberg as being: 0.09 ppm Au, 23 ppm Ag, 1.0% Cu, 2.4% Zn, 0.3% Pb, 27% S, whereas Boliden was: 15.5 ppm Au, 50 ppm Ag, 1-4% Cu, 0.9% Zn, 0.3% Pb, 25% S, 6.8% As, so the difference between the two is marked but not overwhelming.

Kristineberg was discovered in 1918 by electrical methods about six years before the first actual discovery of the Boliden ore body, also through electrical geophysics, plus diamond drilling. The electrical work done in the Kristineberg area resulted in the first ore found by this means in Sweden. Underground investigation, however, did not begin until 1934 and mining was not commenced until 1941. By about 1950, development had reached as far down as the 490-m level. Over the 10 years from 1964 through 1973, Kristineberg mine produced about 9.94 million tons in contrast to the 4.80 million that came from Långsele in that time. In the time since, if these comparative rates have been maintained, total production from Kristineberg would still exceed that from Långsele by a considerable tonnage.

The center of the Kristineberg dome is composed of the volcanic-rock series that is the oldest supracrustal rock in the entire Skellefte area; these volcanics have been intruded by synorogenic Svecofennian Jörn granite. The usual Transition group of rocks encircles the dome and underlies the Phyllite group that is outside the dome area proper. The Transition zone contains large bodies of limestone and dolomite; the sedimentary and volcanic rocks have been folded into linear bodies parallel to the fold axes. Alteration of the volcanic rocks in the area to sericite-chlorite is extensive, particularly so in the vicinity of the sulfide ores.

The Jörn granite borders the volcanic rocks of the Kristineberg area on the east, the Kristineberg deposit lying at the WSW tip of this Jörn granite mass. It is du Rietz' opinion (1953) that the Jörn granite either was folded with the volcanics or was intruded during the folding. At that time, the Jörn granite was considered to be of Archean age, a concept now discredited by radiometric dating; since the Jörn is now thought to be about 2000 m.y. old and syntectonic, its intrusion during the folding must be favored. Grip's sketch map (1978) shows the rudely triangular mass of Jörn granite swimming in a sea of volcanics; these, in turn, are bordered on the northeast and southwest by Revsund (palingenic) granite and along the arc that curves convex to the north by the Phyllite group. The schistosity of the Jörn granite almost corresponds to that of the extrusive rocks. Although, in most places near the contact, the Jörn granite may be porphyritic, locally it is aplitic in character; in the neighborhoud of Kristineberg, the granite is a porphyry that has a well-defined granophyric texture. Further into the rock, its composition is that of a tonalite and is darker in color. Projections of the granite into the volcanics have been altered in part to much the same material as the adjacent altered volcanics, that is, largely chloritized as well as partly sericitized. The altered Jörn granite next to the original extrusive albite porphyry du Rietz calls a sericite-bearing quartz-chlorite schist with relatively more chlorite than in the altered volcanics. The biotite of the granite largely has been converted to chlorite.

The complex of volcanic rocks adjacent to Kristineberg initially was made up of sodic quartz porphyry (albite porphyry) but now is completely converted to quartzitic sericite schists. The rocks of this complex have been interfolded with the overlying phyllites to the west and the Jörn granite complex to the east. The folding is isoclinal and the axes in general pitch to the west. To the south the volcanics are intruded by the palingenic Revsund granite, a few hundred million years younger than the Jörn.

Locally, to the north and west of the Kristineberg mine, isolated patches of essentially unaltered porphyry can be found, at least 2 or 3 km from the ore body. The most common rock now to be seen north of the mine is quartz-sericite schist, but less strongly altered porphyries are more or less sericitized and chloritized; these rocks may contain some residual biotite, but normally, it has been completely converted to chlorite. These rocks contain some disseminated pyrite with the much altered rocks containing more of the pyrite than those less affected by the alteration process.

In the southern portion of the Kristineberg area, alteration has gone farther than the development of sericite; biotite, cordierite, and/or andalusite have been formed. Northward, towards the mine, both cordierite and biotite disappear, cordierite doing so before biotite.

A few small areas of dacite and hornblende-bearing dacite rocks remain essentially unaltered; they may be chloritized, but they are less sericitized than the rocks that surround them. These effects are due more probably to their original more mafic character than to any change in the altering solutions.

The primary volcanics consisted of both lavas and tuffs, and the remnants of these rocks that are least altered were silicic flows with well-developed albite crystals.

The Jörn granite to the east of the mine area apparently dips west toward the volcanics, and these, therefore, almost certainly lie on a basement of Jörn granite. The normal dip of the volcanics overlying this granite is between 20° and 45°W. At Kristineberg, the average dip appears to be about 30°, whereas at Rävliden it is only 20° with this dip being the normal one for the volcanic complex. Between Kristineberg and Rävliden (the latter being about 4 km due west of the former), the thickness of the volcanic beds probably is not more than 2 km and normally, the thickness is appreciably less due to the isoclinal folding of the complex.

In the far northern part of the area, the schistosity dips about vertically, but, in the immediate neighborhood of the mine, the dip of the schistosity is 60°S. Because of the strong alteration of the volcanic rocks and their uniform composition, none of the original stratigraphic subdivisions of the volcanics can now be recognized.

Within the Kristineberg area, several greenstone (formerly lamprophyre) dikes cut both the volcanics and the Jörn granite; quartz-tourmaline veins also are present and they cut the same rocks as the dikes. The tourmaline has not been altered, as were the volcanics, which indicates to du Rietz that these dikes were introduced after the alteration of the volcanics. He considers the tourmaline and its associated quartz to have been emplaced during the regime of the Revsund granite but were younger than the sulfides that were developed as a result of the Revsund magmatism.

The Kristineberg ore body is located in the altered volcanics just above the wedge of the underlying anticline in the Jörn granite. In the folding, the granite was plastically incorporated into the volcanics and appreciably brecciated.

The deposits of Rävliden, Rävlidmyran, and Mörkliden are at the top of the volcanic complex on the west side of the dome; this location is in contrast to that of Kristineberg is well down in the volcanic series, a position unusual in the Skellefte district.

The Kristineberg ore is a copper- and zinc-bearing pyritic ore; the Rävliden ore is mainly a zinc-copper ore, and Rävlidmyran is zinc-lead ore. The ore body known as Sture in located near to Rävlidmyran and contains a copper-bearing ore.

Although the Jörn granite does not show discordant contacts with the volcanic beds, the Revsund granite does; nevertheless, Gavelin, considers the Revsund to be migmatitic. The Phyllites that usually are immediately younger than the Volcanics (with allowance being made for the Transition zone that normally is present between them), but the Phyllites are absent in the neighborhood of the Kristineberg mine more probably having been removed by erosion than that they never were in the sequence. The Phyllites are, however, part of the rock sequence in which the Rävliden and Rävlidmyran ore bodies are found, both being near the contact between the volcanic and sedimentary rock types. The Kimheden mine, northeast of Kristineberg, definitely is within altered volcanic rocks.

The actual Kristineberg ores are arranged in two parallel zones of sulfide deposits. These two zones strike generally east-west, although the more northerly of the two zones (that contains from west to east the A_4, A_3, A, C, and D) strikes about N65°W for the A_4 part of its length. The more southerly of the two zones includes only the B ore body. The total length (on the 170-m level) of the ore bodies A_4 through D is just over 1000 m; if the overlapping portions of ore bodies A_3 and A were added to the length, it would be about 100 m more.

The B ore body is quite complex with the component parts anastomosing and with some parts in an en echelon arrangement with the others, its total length is about 355 m, and it is opposite the eastern half of A ore body and the western half of C ore body. A fault cuts B ore body about one-third of the distance from its east end with the fault offsetting the east end of the B ore body about 60 m to the north. This fault strikes sinuously about N15°W and cuts between ore bodies A and C where no mineralization is mapped. On the 90-m and 250-m levels, the ore bodies have much the same structure but are appreciably shorter than on the 170-m level.

These ore bodies are on the inner portion of a large anticlinorium (Grip, 1978) that arches over a core of Jörn granite and gneiss. Just north of ore-bearing area is a tongue of this granite that has chilled margins against the altered volcanic rocks. The dip of the ore zone is about 60°S and plunge is 40° in the direction S60°W. These ore bodies are quite narrow and are so long in the direction of plunge that their lower limit had not been reached when Grip's 1978 paper was written.

As is true of so many of the Skellefte ore districts, this mineralization required three periods of tectonic activity to complete. Preceeding the first of these was the introduction of scheelite-bearing quartz-tourmaline dikes. The first tectonic activity produced linear structures in the planes of schistosity that now, after two more stages of tectonism, dip 40° in a S60°W direction. It was these structures that localized the first stage of ore emplacement; this mineralization, however was preceded by a strong magnesium metasomatism that moved outward from the shear zones and converted the wall rocks to sericite-biotite-chlorite schist. This type of wall-rock alteration also accompanied the emplacement of the ores and, to a minor degree, continued after ore formation had ceased.

The sulfides of this first stage of ore mineralization had pyrite as their major representative; this pyrite normally is quite coarse-grained, the more so where the pyrite is disseminated in the altered schists. The ore minerals - sphalerite and chalcopyrite - were developed as matrix material among the pyrite grains and replaced not only pyrite but each other, the relative result depending on what chemical conditions were in the part of the ore body being formed.

The second stage of mineral development saw the enrichment of the ore in copper, the chalcopyrite being particularly added around segregations of quartz in places where structures developed by rotation about horizontal axes compressed the southern block of the ore body against the northern.

The third stage was the introduction of the definitely younger sphalerite into a narrow fissure that opened where the southern ore block moved horizontally toward the west along the hanging wall of the A and A_3 ore bodies (that is on the south sides of the south-dipping veins). Rotated fragments of chlorite schist and minor folds along the contact confirm the developemnt of the ore in such a fissure. When Grip wrote his 1978 paper, the ore had been proved to a depth of 690 m, but even then the ore was known to extend farther downward. This geometry gives the ore bodies a distance in the plunge direction of 1100 m. The area of the body is 10600 m^2, and its production tonnage [see table in Skellefte (General)] almost certainly will remain (as it is now) the largest in the district.

The composition of the ore, so far as was known when Grip wrote, is given at the beginning of this discussion of the Kristineberg district.

Gavelin (1953) considered that the Kristineberg ores were of four types: (1) quite compact pyrite ore; (2) chalcopyrite-bearing pyrite ore (wet ore); (3) zinc ore (ZnS-bearing-pyritic ore with some chalcopyrite and galena); and (4) quartzitic ore (chalcopyrite-bearing ore with more or less pyrite and considerable unreplaced but highly altered wall rock, i.e. dry ore).

Prior to the introduction of any of the four Kristineberg ore types, the ore area was cut by what are now greenstone dikes and originally probably were lamprophyres. In Boliden, these lamprophyre dikes are associated with sulfide-bearing quartz tourmaline ore; in Kristineberg, the quartz-tourmaline veins are younger than the lamprophyres (greenstones) and have no sulfide mineralization accompanying them. In both mines, some scheelite is found with, or near, the quartz-tourmaline veins, but, in neither mine, is there enough scheelite in any one place to make it worth recovering. In Kristineberg, the quartz-tourmaline veins are cut by the pyritic ore; in Boliden, the situation is essentially the same. In Boliden, the first of the main types is the arsenopyrite ore; at Kristineberg, the amount of arsenic in the deposit is not worth mentioning in the analysis given by Grip.

Thus, except for the lack of arsenopyrite mineralization at Kristineberg, the mineralization process at the two mines was not markedly different, despite the 90 km that separate the two deposits. Zinc is appreciably more abundant at Kristineberg and copper at Boliden.

The quartz-tourmaline veins (or simple quartz veins without tourmaline) are abundant at Kristineberg with their greatest concentration being in the footwall of the A ore; around the B vein, or south of it, these veins are quite rare. Although quartz-tourmaline veins may strike in any direction, their most common strike directions are NW-SE (actually N30°W) or NE-SW (actually N60°E). The dips of these vein-dikes ranges quite widely, but they average about 70°S. The NW-SE veins dip toward the SW or NE with the SW direction being the more common; the NE-SW veins also dip in both possible directions, but the NW veins are appreciably more abundant than those that dip SE. The veins normally are quite narrow, only a few cm wide, so they seldom merit the term dikes. The veins in many places are made up of alternating zones of tourmaline and quartz; in the narrower veins, the tourmaline is present as aggregates of needles. Although most of the dikes have sharp walls with the enclosing host rocks, some of them show appreciable evidence of having replaced the sericite-quartz schist.

The quartz-tourmaline veins are most common on the footwall side of the A vein, indicating that a zone of brittle weakness existed in that area prior to the introduction of the quartz veins or the later ores.

If all the ore in the complex ore present on the north side of the deposit (as far east as the break in the vicinity of the fault that cuts the B ore) is considered as A ore, then the major part, as far west as the A_3 should, for the sake of clarity, be considered as A_1. These A ores are dominantly pyritic with the notable pyrite-sphalerite ore being located on the hanging wall (south) side of the pyritic ores. The A_3 ore consists almost entirely (at least on the 170-m level) of pyrite-sphalerite ore with some normal pyritic ore entwined with it. The A_4 ore to the west is all pyritic ore.

In most instances, the width of the footwall pyritic ore is much greater than that of zinc ore, especially on the upper levels. On the 170-m level, the pyritic ore is about 350 m long and is up to 15 m wide; the narrowest is about 2.5 m. Not all of the pyritic ore is copper-bearing. The copper-bearing A_1 ore can be separated into three distinct varieties: (1) footwall ore - lean ore with appreciable unreplaced chlorite; (2) rather solid pyrite (in the center); and (3) quite copper-rich pyritic ore on the hanging-wall, next to the zinc-rich ore. Type (3) is usually thinnest where it is richest in chalcopyrite; locally this ore changes to copper-rich zinc ore. Further, pyritic ores locally may be associated with the zinc ore.

The common gangue minerals in pyritic ore are chlorite, quartz, and locally sericite and amphibolite, though the last two apparently do not occur together, the former being on the footwall and the latter on the hanging. Sporadically present are talc, tourmaline, epidote, and one carbonate or another. How much of the gangue is truly such and how much remains after the sulfides had replaced gangue is uncertain. Some of the gangue minerals corrode the pyrite, suggesting that these at least may have formed from the ore fluids rather than being relics of earlier alteration minerals. The chalcopyrite in the pyritic ore has an irregular distribution, generally occurring as brecciating veins in chloritic pyrite ore and thus, probably is later than the iron sulfide. Although sphalerite is no more than an accessory in the footwall ore, it may be quite abundant in the copper-rich hanging-wall ore, thus providing a gradual transition from copper to zinc ore.

Magnetite is a rare accessory; it is idiomorphic but is more corroded than pyrite.

Where pyrite grains have been broken, they may be replaced by chalcopyrite along the cracks so produced. Locally, chalcocite and covellite replace chalcopyrite; it hardly seems possible that these two copper sulfides are of secondary origin, but Gavelin does not say what he thinks about this problem.

The A_1 zinc ore is concentrated on the hanging-wall side of the complex vein structure with pyrite almost always being the more abundant of the two; the sphalerite is fine-grained and forms the ground mass around the pyrite grains. The sphalerite is so light-colored that it is difficult to distinguish from some of the gangue minerals. The zinc ore, as a unit, however, is easily recognized because of its special gray tone. The ground mass, in addition to the sphalerite, contains much talc and ball-like inclusions of dark chlorite. The ore is commonly banded, showing transitions from granular pyritic ore to light-colored zinc ore and from chloritic sulfide ore to grayish zinc ore. The rock on the hanging-wall may be impregnated by pyrite and sphalerite. The zinc ore may be divided into distinct sub-veins by small balls, slabs, or schlieren of chloritic rock. Zinc ore along the hanging wall may contain appreciable galena and tetrahedrite.

The A_3 ore begins at the western limit of the A_1 ore where the latter fingers out into thin streaks. This low-grade A_1 ore may extend out to the west or northwest in an en echelon pattern with the A_3 ore. The easternmost part of the A_3 ore is a zinc-bearing pyritic ore with a varied, but low, content of pyrite. Farther west the A_3 ore is at its richest, having about 4 per cent Zn and 1 per cent Cu. In the far west, the A_3 ore is mainly of the pyritic type but includes a thin band of zinc ore along the hanging-wall; this ore dies out to the west as pyritic ore low in copper and zinc. The wall rock of the A_3 ore is less broken than that of the A_1 and the wall-rock alteration is less extensive.

The A_4 ore is a pyrite ore; it is about 145 m long and averages about 8 m wide. It dips 39°S and pitches about 45° to the SW. It is irregular in mineral content, consisting of intercalations of sulfide-bearing sericite-chlorite schist in unmineralized wall rock. Zinc is present in only vanishingly small quantities and the copper content is not much better, although some streaks of minor size on the footwall of the ore are rich in chalcopyrite. The copper content is only a few tenths of 1 per cent, but the sulfur amounts to more than 30 per cent. The usual gangue minerals, quartz, sericite, and chlorite, are common. The ore is separated from the ubiquitous quartz-sericite schist by a 1-m wide band of quartz-chlorite schist. Concretions of quartz-tourmaline are present on the footwall side of this ore.

Gavelin divides the B ore into three types, B_1, B_2, and B_3; they are some 100 m into the hanging-wall of the A_1 ore. The B_1 and B_3 ores are pyrite ore inter-mixed with dry ore (see above); the B_2 ore is entirely pyritic. In the eastern part these ores are quartzitic (dry) ore and chalcopyrite-bearing pyritic ore in the west; layers of pyritic ores are separated by bands of sericite-quartz or chlorite-quartz schist.

The rock complexes of the Kristineberg mine are strongly schistose and dip south in the middle and eastern parts of the mine and to the SSW in the western. The dip ranges between 30°S and 90° with the dip being conformable with that of the schistosity of the wall rocks. The pitch of the fold axes of both the folds and the linear structures of the rocks and of the ores are conformable; the dip of the pitch ranges between 20° and 65° and it is in directions that range from S80°W to S40°W. A few transverse folds have been identified.

The character of the ore and gangue minerals indicates that the ore was formed most probably in the mesothermal range, although some of the sphalerite and the chalcopyrite may, on the basis of inclusions of the one mineral in the other, may have been formed in the hydrothermal range.

The information available to the general public on the geology of the Rävliden mine is far less abundant or satisfactory than that on Kristineberg. No paper compatable to that of Gavelin (1953) is in the public domaine for Rävliden or Rävlidmyran, so I depended on the brief summary in Grip's (1978) paper for an idea of what the Rävliden and Rävlidmyran ores are like. The two deposits are in the transition zone between the volcanic rocks (below) and the

Phyllite series (above); Rävliden is some 4 km to the WNW of Kristineberg; Rävlidmyran is about the same distance from Kristineberg but lies directly west of that latter deposit. Mining began at Rävliden during the period 1940 to 1944; that at Rävlidmyran commenced in the period 1950-1954. By the end of 1974, production at Rävliden had amounted only to 888,000 tons, whereas that at Rävlidmyran was 3,878,000 tons or almost 3 million more than Rävliden; this disparity almost certainly has grown since 1974. In the 20 years between 1955 and 1974, Rävlidmyran averaged 400,000 tons of production to Rävliden's average over the same period of only 19,000. Over the entire period from 1925 through 1974, Rävlidmyran was essentially tied with Renström for sixth place in total production.

The Rävlidmyran mine is composed of two parallel ore zones that lie about 200 m apart; the more southerly of these contains four ore bodies that range between 40 and 160 m in length and have widths up to 4 m. The total area of these four ore bodies is 2045 m^2, whereas the northern zone contains only one ore body, the Sture, that has a length of 350 m and averages 20 m in width; its total area is 7000 m^2.

All of the Rävlidmyran ore lenses strike in an east-west direction, have a vertical dip, and plunge 25° to 30°W. The average composition of the ore in this deposit is 0.9 ppm Au, 55 ppm Ag, 1.2 per cent Cu, 3.7 per cent Zn, 0.5 per cent Pb, and 20 per cent S.

The host rocks of the ore deposits are (from oldest to youngest) dark phyllite, sericite schist, chlorite-talc-phyllite, and impure domolite in which, in places, skarn has been developed. The rocks normally are quite hard and their mica content is small.

The Sture ore body is made up, in part, of massive sulfides and in part of disseminated chalcopyrite ore. The massive ore is composed of pyrite, chalcopyrite, sphalerite, and galena in amounts that differ largely from one place in the deposit to another. The copper-rich ore type (found normally in the footwall of the deposit) is predominantly chalcopyrite and pyrite. Pyrrhotite is present in both ores but is not evenly distributed in the ore-types. A zinc-rich ore is present in the hanging-wall and exhibits well-developed banding in which bands of pyrite alternate with ones of sphalerite; the thickness of the bands may be as much as 10 cm or as thin as 2 cm.

The location of the Rävlidmyran ores was controlled by drag folds and the intense shearing that accompanied their development. Grip believes that the overlying phyllites may have dammed the upward flow of the ore-forming fluids. The ores also were deposited in the silicate and carbonate rocks in varied amounts with replacement being the agent of such deposition. Grip suggests that certain breccias in the ore bodies were produced by the forceful intrusion of a mush of crystal ore, lubricated by a saturating amount of hydrothermal solution. It seems more probable that the ore fluids that formed these breccias were under such high upward-driving pressure that they fractured the rock and then deposited their loads in the fractures. The force of the upward-moving ore fluid well may have been sufficient to hold the fractures open and to maintain the brecciated fragments in the same relative positions to those which they originally assumed when they were broken.

The Rävliden deposit is 1500 m south of Rävlidmyran and occupies a position at about the same stratigraphic level as the Rävlidmyran ores. The deposit consists of several linear bodies that strike N65° to N80°W, and their plunge is 25° to 30° in a WSW direction. The ore is of two kinds, copper-rich and zinc-rich and the compositions of the two types are as follows:

Ore Type	Au ppm	Ag ppm	Cu%	Zn%	Pb%	S%
Zinc-rich	0.2	193	0.4	8.4	1.7	20
Copper-rich	0.6	67	1.9	3.1	-	22.

The country rock in which the Rävliden ores were emplaced was sericitic and chloritic schist, chloritic phyllites, calc-silicates, and graphitic phyllites. As is true at Kristineberg and Rävlidmyran and at many other deposits in the Skellefte district, the zinc and copper ores do not occupy the same rock volumes but are quite distinctly separated with the copper-rich ore in the footwall and the zinc-rich ore in the hanging-wall. This arrangement is not always distinct and sharp as the ores may be intertwined or grade gradually from one to the other.

This arrangement probably is one of the major reasons for Rickard and Zweifel's postulate that the ores were initially laid down on the sea floor with the ore fluids being volcanogenic in origin. The close relationship of the ore deposits, both in this sub-district and in the Skellefte district as a whole, to structures developed long after the host rocks had been lithified, argues cogently against this.

The character of the ores themselves and of the host rocks in which they were emplaced seems to argue strongly that they were deposited in the mesothermal range. More detailed study of those ores that contain both chalcopyrite and pyrite would show whether it would be necessary to add the term hypothermal to the classification in the Kristineberg-Rävlidmyran sub-district.

ADAK-RUDTJEBÄCKEN, VÄSTERBOTTEN

Middle Precambrian	Copper, Gold, Pyrites, Zinc, Arsenic	Hypothermal-1, -2 (minor) to Mesothermal

Edelman, N., 1963, Structural studies in the western part of the Skellefte district, northern Sweden: Geol. Fören. Stockholm Förh., v. 85, pt. 2, no. 513, p. 185-211

Gavelin, S., 1945, Arsenic-cobalt-nickel-silver veins in the Lindsköld copper mine, N. Sweden: Sveriges Geol. Undersök., ser. C., no. 469, Årsbok 39, 18 p.

____ 1948, Adakområdet. Översikt av berggrund och malmer: Sveriges Geol. Undersök., ser. C, no. 490, Årsbok 41, 54 p. (Engl. Summ.)

____ 1954, A telluride assemblage in the Rudtjebäcken pyrite ore, Vesterbotton, N. Sweden: Sveriges Geol. Undersök., ser. C., no. 536, Årsbok 48, 14 p.

Grip, E., 1978, Adak dome, in Bowie, S. H. U., and others, Editors, Mineral Deposits of Europe, Volume 1: Northwest Europe: Inst. Min. and Met. and Mineral. Soc., London, p. 159-161

Hübner, H., 1967, Klüftung und Vererzung im Grubengebiet Adak: Sveriges Geol. Undersök., ser. C, no. 628, Årsbok 61, 12 p. (Engl. Summ.)

Ljung, S., 1974, Adak-Lindsköld-och Brännmyrangruvorna inom Adakfältet - Petrografi och malmer: Sveriges Geol. Undersök., ser. C, no. 701, Årsbok 68, 94 p. (Engl. Summ.)

Notes

The Adak dome, the area containing the Adak, Lindsköld, Brännmyran, Karlsson, and Rudtjebäcken ore bodies is located about 35 km N30°W from Rakkejaur with the only ore deposit reported in the area between the dome and Rakkejaur being the most unusual (for the Skellefte district) nickel-copper deposit of Lainijaur. The deposits in the Adak (65°23'N, 18°35'E) dome deposits are in a very different stratigraphic position from those of the Boliden-Rakkejaure district to the southeast. The succession in the Adak area (from top to bottom) is: (1) gray biotite-hornblende tuffite interbedded with graphitic phyllites; (2) upper banded series (tuffite formation); (3) andesite formation with intercalated limestone bands; (4) lower banded series (tuffite formation - volcanics and sediments - this is the ore horizon); (5) micaceous schist group, a transition zone which is a shear zone in the Adak mine and contains ore in the Lindsköld and Adak mines. In the southeastern portion of the dome, the upper formations were intruded by a red microcline (the Adak) granite - this granite more or less conformably covers the outer parts of the dome and completely encircles it.

The longer axis of the dome trends ENE, not a common trend in the Skellefte district proper where the general trend is to the ESE. The ore-bearing horizons are exposed for 4 km with a maximum width of 2 km. The Adak mine is at nearly the southwest tip of the domal structure with the Lindsköld mine being about 0.4 km about due north of the Adak. The Brännmyran mine is just over 2 km about E28°S from Adak and about 2.5 km S30°E from Lindsköld. The Rudtjebäcken mine is at the far northeast end of the dome N70°E from Adak at a distance of about 3.5 km. The Karlsson ore is located less than 1 km ENE of Adak. The beds in the dome dip outward at from 15° to 40°. The eastern border of the dome is very irregular, it probably having been appreciably faulted. The dome also has been strongly cross-folded in a NW-SE direction. The ore deposits of Lindsköld and Brännmyran are strata-bound ore bodies, and both are controlled by a cross-fold that cuts across the dome from the one deposit to the other; the strike of the axis of this fold is nearer S60°E than the S45°E called for by Grip's text. Lindsköld plunges to the northwest and Brännmyran plunges to the southeast. The Rudtjebäcken ore body also is strata-bound on a separate cross fold in the northeastern part of the dome.

Ljung (1974) points out that there well may be a direct relationship between the intrusion of a plagioclase granite found at depth in the Adak mine and the formation of the dome. The uplift of the granite (with its transitional gneiss and migmatite zones) may have pushed upward the volcanic-sedimentary rocks that now cover the dome. The rate of this uplift must have been very slow because there is no really noticeable thinning of these formations over the center of the dome. Further, there are no facies or discordances in trends that would indicate that intermixing of shallow-water sediments (such as might suggest continuous doming) at higher stratigraphic levels. Except, apparently, for the faulting mentioned by Grip as having taken place along the eastern border of the dome, the outline of the dome is very regular; there is no block-faulting in evidence anywhere (Ljung, 1974).

As is the case at Boliden, the favorite locale for ore deposition is around the boundary between two main stratigraphic groups, the lower of which normally is essentially volcanic, as can easily be ascertained in areas where the metasomatic alteration is slight. The upper unit is appreciably different, being composed of both volcanic and sedimentary rocks. This boundary also separated the rock volumes in which the plagioclases in the granite have been extremely or completely altered from those above in which the plagioclases are practically unaffected. The strain distribution within the zone of domal deformation is mainly shear in the upper part and tension in the lower.

The metamorphic rocks that are present in the Adak dome are of a much higher grade of metamorphism than any of those of the Boliden region proper, being of the cordierite-andalusite-muscovite subfacies of the amphibolite facies. The rocks of the dome are divided by Ljung into two major groups, the upper (and essentially unmetamorphosed) one of which is a tuffite group that is separated into four formations; these are (from bottom to top): (1) a banded tuffite formation with pyroclastics and clastics such as graywacke and graphite schist; (2) an andesite lava and tuff formation with intercalated hornfels horizons; (3) an upper banded tuffite formation [very similar to rock-type (1)]; and (4) a dark-gray biotite-hornblende formation. To the south and east the rocks of this upper group are overlain by a red microcline granite, the contact of which with the volcano-sedimentary formations is in conformity with the rock it intrudes.

The underlying schist and microgneiss group consists of three formations; these are (from bottom to top): (1) the Adak formation; (2) the cordierite-biotite schist formation; and (3) the cordierite-cummingtonite-anthophyllite schist formation (more concisely described as the cordierite-amphibole schist formation).

The rocks of the Adak formation have been divided into two quite different assemblages - the amphibole assemblage and the mica assemblage. The amphibole assemblage contains rocks of two types; (1) the cummingtonite-anthophyllite micro-gneiss with accessory biotite and chlorite and rarely cordierite and (2) cordierite-cummingtonite-anthophyllite schist with accessory biotite and chlorite. The mica assemblage is composed of: (3) biotite-chlorite micro-gneiss with accessory cummingtonite, anthophyllite, and cordierite and (4) cordierite-biotite-chlorite schist with accessory cummingtonite and anthophyllite.

Characteristic of the rocks of the Adak formation are stratified lime- and lime-skarn horizons as well as gray or white cummingtonite micro-gneiss. Ljung believes that, during metamorphism, a substantial exchange took place to produce an anorthite-hornblende-diopside (hedenbergite)-garnet skarn. During the later introduction of sulfide ore minerals, epidote and zoisite were formed from pyroxene, hornblende and/or anorthite; the skarn minerals in many places show a zonal arrangement.

The schists of the cordierite-biotite schist formation (4, above) include more plagioclase than does the cordierite-amphibole schist (2, above). Locally (the Brännmyran mine), these metamorphosed formations contain garnet. The cordierite-biotite schist formation is cut by concordant zones of cordierite-amphibole schists that, in many places, contain ore, but the cordierite-biotite schists are not suitable for ore deposition. Because the cordierite-biotite and cordierite-amphibole schists are erratically distributed in this formation; the sulfide mineralization also is erratic, thus, much of the mineralized Adak rock is not ore. The more intense the deformation, the greater the tension and shearing and the more of the cordierite-biotite rock is converted to cordierite-amphibole type and the more common the structures favorable for ore deposition. The cordierite-amphibole rock consistently overlies the cordierite-biotite variety.

Ljung thinks that there were several different phases of metasomatic alteration. He postulates an early hydrothermal phase in which plagioclase was broken down; the two upper schist types so affected have a bulk deficiency in Na and Ca and are much more altered than the lowest (and lime-bearing) beds of the Adak formation.

The second phase of the progressive metamorphism developed the original biotite schist, biotite micro-gneiss, and lime skarn; the latter two rock types were developed only in the Adak formation.

During a still later tectonic phase, in which fracturing and shearing were intense, the upper part of the original cordierite-biotite schist was converted to cummingtonite-anthophyllite-cordierite rock. Parts of the Adak formation also were affected by this type of alteration, converting the parts so affected to cummingtonite-anthophyllite schists and gneisses. Thus, the general effect of this tectonic phase was to add the cummingtonite-anthophyllite-cordierite to the cordierite-biotite type. Normally, the more abundant the sulfides in the rock involved in this alteration phase, the more complete was the conversion of biotite to amphibole. This brings Ljung to the conclusion that the solutions that achieved the biotite-amphibole conversion moved outward into the wall rock from fissures and fractures developed in the third tectonic phase. There are exceptions to this general rule that sulfides and amphiboles are found together, but usually, they are associated in space. This association of sulfides and amphiboles, Ljung considers, is not a temporal one, however. One reason for this belief on his part is that appreciable rock volumes that contain amphiboles are lacking in sulfides. This conclusion is strongly supported by the fact that the sulfides and the almandine garnet are essentially always closely associated with each other, and this is supported by the proof Ljung says he has that the garnets formed definitely later than the cummingtonite.

During the emplacement of the ore sulfides, some of the biotite in the host rock was altered to chlorite, and, where the iron content was high, almandine or magnetite was developed, depending on the oxygen fugacity of the system (the higher that value, the more likely magnetite was to have been formed). During this process, some epidote and sericite also were produced in the Adak formation.

Ljung points out that there is a close connection between the dome structure at Adak and the emplacement of the ores. This close connection between the structures resulting from doming is true throughout the Boliden district, the more complicated the domal structure in a given area, the more complex the ore pattern. The formation of the domes appears to be directly related to the formation and behavior of the syntectonic Jörn granite, an event demonstrated by the frequent occurrence of granite cores in the domes. The granite in the Adak dome is similar chemically and mineralogically to the Jörn granite as it is known in the Kristineberg area, some 35 km SSE of Adak. Ljung considers that the simplest way to explain the strain structures in the Adak dome is through the action of the stress-systems produced during the doming action (Edelman, 1963).

Differential movements produced drag folds along the flanks of the dome, and these created shear and tension fractures in a zone close to the boundary between the two major stratigraphic groups (schist and micro-gneiss groups). This resulted in the steep breccia zone in the Adak mine and in shear zones at depths that are related to the boundary between the Adak formation and the cordierite-biotite schist; these structures acted as favorable sites for ore deposition. Much of the ore so formed was concentrated in the lime horizons in the upper part of the Adak formation, mainly as chalcopyrite and pyrrhotite.

Ljung recognizes two main mineralogic types of ore: (1) pyrite ore with subordinate chalcopyrite and sphalerite (as recovered from the already mined part of the Rudjetbäcken ore body) and (2) chalcopyrite ore that is present in the Adak, Lindsköld, Karlsson, and Brännmyran mines and in which pyrite is a minor component.

He, however, has characterized six types of ore, based of different structures that contain them. These are: (Ia) compact pyrite or chalcopyrite-pyrrhotite ores (Rudtjebäcken, western part of the main Lindsköld ore body); these are located in shear zones in the lower banded tuffite formation and the andesite formation; these ore bodies are clearly discordant to their host rocks. (Ib) pyritic ores of the Fahlband type - chalcopyrite and pyrrhotite, plus subordinate galena and sphalerite - located in the upper shear zone in the Lindsköld mine; this zone includes the transition zone and the upper part of the cordierite-amphibole schist formation. (Ic) ores in tension fractures within most of the cordierite-amphibole schist formation - ore minerals are chalcopyrite, pyrrhotite and Co-bearing arsenopyrite - located in lower Lindsköld ores and the Karlsson ore body. (Id) breccia ores of the Brännmyran type; these are contained in two zones of cordierite-amphibole schist as fracture fillings by chalcopyrite, pyrrhotite, pyrite, and Co-bearing arsenopyrite; the ores include many pebbles of wall rock -

in this ore type the pyrrhotite is less abundant than in type IIa (following). (IIa) irregular ores of the Adak type that contain chalcopyrite, pyrrhotite, pyrite, and Co-rich arsenopyrite; the ores are included in a NE-striking steep and wide breccia body that cuts across the formations discordantly where the wall rock of the zone is heavily mineralized with magnetite and garnet in anthophyllite-gedrite skarn. The footwall of the breccia zone contains dragfolded lime and lime skarn horizons over a limited vertical range. In these Ca-rich horizons, replacement masses of chalcopyrite and pyrrhotite are heavily concentrated. (IIb) chalcopyrite and pyrrhotite ores are present in horizontal or gently dipping shear zones and lime-skarn horizons; locally the skarn contains much galena and sphalerite, but the arsenopyrite content of these ores is low.

The first stage in the crystallization of the various ore types is the development of Co-rich arsenopyrite (of course, a common feature in the ores of the Boliden district from Boliden proper on to the WNW). This is followed by pyrrhotite-pyrite-sphalerite mineralization that ends with the precipitation of chalcopyrite. The last stage of ore mineralization was Ag-Co-Ni arsenides and sulfides and sphalerite and galena in calcite in fissures.

In the central part of the Lindsköld mine, a distinct zoning of sulfides occurs in which the upper section is a pyritic zone with lenses of Co-poor arsenopyrite and a lower zone with idiomorphic, coarse-grained arsenopyrite with a higher content of cobalt and without pyrite. The cobalt content of the arsenopyrite increases toward the center of the dome. The ore generally is poor in gold and silver; there is a definite correlation between silver and chalcopyrite which indicates to Ljung that the Ag-component is bound up in the chalcopyrite lattice. The gold, on the contrary, shows no consistent relationship to the copper content, but gold is enriched in the copper concentrate, indicating that the gold occurs interstitially in the chalcopyrite. Exsolutions of cubanite and valleriite are common; exsolution of chalcopyrite in sphalerite is noted only in the pyrite mineralization without pyrrhotite. Metamorphosed dikes of mafic rocks in several generations cut through the dome; the majority of these dikes have been affected by the tectonically generated rock movements and are older than, or contemporaneous with, the formation of the ore. In places, the dike rocks are present as boudins within the ore body, demonstrating that the ore, in such cases, definitely is later than the dike rocks. One dike of mafic, quartz-free composition encircles the ore dome and dips at about 30° out from the dome center; Ljung says it is younger than the ore.

Grip reports that there are two types of ore in the dome: (1) massive, stratiform ore as at Lindsköld, Brännmyran, and Rudtjebäcken and (2) breccia-impregnation ore as at Adak and Karlsson. The breccia ore, according to Grip, occurs in micaceous schist and is discordant to it, whereas, the stratiform ores are more or less concordant to the bedding or are close to the contact between the micaceous schist and the overlying banded series.

Pyrite, pyrrhotite, and chalcopyrite are present in all the deposits with sphalerite occuring only at Rudtjebäcken in ore-grade amounts. Arsenopyrite is erratic in its distribution; locally it is quite abundant. A number of accessory minerals have been noted. In some deposits, several stages of mineralization can be observed.

The Adak deposit (Grip, 1978) is composed of a major number of small ore bodies that are elongate and plunge gently in a direction of S50°W. They are related to a system of joints and, to some degree, to the primary banding of the micaceous schist (the upper part of Ljung's Adak formation). The most important joint directions strike N45°E and N75°E and dip 50°NW and 10°SE, respectively. Where these joint systems intersect, ore concentrations are most common; these intersections appear to have the major channelways for the upward migration of the ore solutions. As the ore solutions traversed outward from intersections along the joints, they also moved out into the wall rocks where further ore emplacement was effected. As the ore is followed outward from the joints, it becomes lower in grade, a not uncommon phenomenon. Where the ore fluids entered the schists, the mineralization developed parallel to the banding in those rocks; where the solutions reached massive quartzite, however, the solutions moved away from the veins along crush zones to develop the typical breccia ore.

Chalcopyrite and pyrrhotite are the dominant minerals although pyrite is a common mineral, it is normally less so than pyrrhotite. Sphalerite is found only in minor amounts and is erratically distributed. Arsenopyrite occurs widespread through the Adak ores and is as idiomorphic crystals in masses of later sulfides; the FeAsS contains 1 to 3 per cent of cobalt.

The ore area of the Adak mine is 1200 m^2 and the grade was 1.9 per cent Cu.

The Rudtjebäcken ore body is a sheet-like mass 1000 m long and 300 m wide; it is present as massive ore below the 125-m level and reaches downward at least as far as the 275-m level. The ore usually is about 3 to 4 m thick but may be as much as 10 m or more. The sheet strikes N30° to N45°W and dips from 15° to 25° NE; it has an essentially level surface; at depth it is slightly folded. In the upper and western portions, the plunge is nearly horizontal, but gradually becomes steeper to reach, in the southeastern part, as much as 20°SE. Grip points out that this means that the ore lies in an anticlinal saddle and on its southeastern flank. The ore is poorer in its upper part, gradually passing upward into low-grade disseminations.

In contrast to the other deposits in the Adak district, pyrite is the main sulfide at Rudtejebäcken; pyrrhotite, sphalerite, and chalcopyrite also are present. The pyrite is quite coarse-grained with crystals of that mineral being greater than 1 cm in diameter. The ore grade is 0.8 per cent Cu, 3.0 per cent Zn, 0.1 per cent Pb, and 35 per cent S.

Rudtjebäcken is unique among those of the Skellefte district in that it contains an appreciable presence of telluride minerals in a deposit where the wall-rock alteration is less pronounced in degree than in any of the other Skellefte ore bodies, but the minerals developed in that process were, nevertheless, indicative of high temperatures of their formation and are much the same as are found in the other deposits of the Adak dome, these are: cordierite, cummingtonite, biotite, chlorite, almandite, andradite, diopside, and hornblende.

The first discovery of the telluride minerals at Rudtjebäcken was made in drill core, and no other example of such minerals had been found up to the time of Gavelin's 1954 paper. The main gangue minerals associated with the tellurides are quartz and microcline, and their presence and their texture lead Gavelin to consider the occurrence as a pegmatite. Some of the large microcline examples are perthitic, and all may contain small idiomorphic or hypidiomorphic crystals of plagioclase; other pegmatite minerals are biotite and lesser amounts of titanite and apatite. The sulfides are later than the non-metallic minerals and replace the silicates. Whether these sulfides and tellurides were the last of the pegmatitic fusion to crystallize or where introduced by hydrothermal solutions is not clear. At the same time that the sulfides were formed, potash also was brought in that converted some of the quartz to (secondary) microcline and muscovite. Gangue minerals introduced at the same time as the sulfides were calcite, epidote, and prehnite, and biotite was converted to chlorite. The late sulfide minerals are, in order of decreasing abundance, pyrite, chalcopyrite, arsenopyrite, galena, sphalerite, and tetrahedrite; they are very erratically distributed; where the tellurium minerals are present, in fact, tetrahedrite is more abundant than elsewhere. The tellurium minerals always are closely associated with each other and, in many places are intimately intergrown. Gavelin identified the following tellurium minerals in his microscopic and X-ray work: altatite (PbTe), hessite, petzite ($Ag_3Au^1Te_2$), and two other unidentified (but probably) tellurium minerals were noted by Gavelin.

Gavelin (1954) notes that the telluride minerals in his pegmatite contain those metals that are, in the Adak district ores, less common than they are in the generality of the Skellefte deposits - these elements are arsenic, lead, antimony, and the precious metals. Normally in the Skellefte ores, where these constituents are not abundant, gold and arsenic were deposited before the crystallization of the main pyrite-rich ore bodies, and lead, antimony, and the silver after the pyrite.

It seems of value to compare briefly the descriptions of the four ore bodies (Adak, Rudtjebäcken, Lindsköld, and Brännmyran) given by each of the two authors (Grip and Ljung) and let the reader synthesize them into a coherent whole, if he can.

Ljung's work is a general discussion of the Adak dome, whereas Grip, after a short, general introduction, discusses four of the Adak deposits (omitting Karlsson). For example, Grip recognizes two types of ore in the entire Adak dome: (1) massive stratiform ore in the mines at Lindsköld, Rudtjebäcken, and Brännmyran and (2) breccia impregnation ore at the Adak Karlsson mines. In contrast to this Ljung defines six ore types in the five mines (these six types have been detailed previously in this discussion).

At Rudtjebäcken, Ljung says there are massive pyrite or chalcopyrite-pyrrhotite ores, and Grip considers the ore to be massive; here there is not too great a difference between the two. Both authors recognize that this is the one mine on the dome where pyrite is the dominant iron sulfide.

At Adak, Ljung describes the ore (in part) as an irregular breccia in the NE-striking, steep and wide breccia zone that cuts the formations discordantly. The wall rock is characterized by Ljung as having a high magnetite and almandite content and anthophyllite-gedrite skarn. Along the footwall side of the breccia zone, there are drag-folds in lime and lime-skarn horizons where locally major concentrations of chalcopyrite and pyrrhotite have been developed by replacement of the skarn minerals to provide another part of the Adak ore. Grip mentions that, in massive quartzite, the ore fluids spread out into crush zones where typical breccia ore is found; he says nothing about the strike of the breccia zone, but he does mention N45°E as one of the important joint directions. Grip says nothing about the Adak ore bodies that are related to the systems of joints and the primary banding of the micaceous schists. The two authors agree on the Adak minerals with chalcopyrite and pyrrhotite as the dominant ones. Obviously these two partial descriptions of the Adak ores do not show much relation, the one to the other.

Ljung places the Lindsköld ores in three of his six categories: (Ia), (Ib), and (Ic). This means that he considers that: (Ia) chalcopyrite-pyrrhotite mineralization is in the shear zones in the banded tuffite and andesite formations; (Ib) chalcopyrite and pyrrhotite with subordinate galena and sphalerite is in an upper shear zone in the Transition zone rocks and in the upper part of the cordierite-amphibole schist; (Ic) chalcopyrite-pyrrhotite ore with Co-bearing arsenopyrite is in tension fractures in the cordierite-amphibole schist. Grip describes the Lindsköld ores as being stratiform ore at the contact between the micaceous quartzite and the lower-banded series that overlies it; this ore is part of a zone that is more or less mineralized around the margin of the dome as far as the Rudtjebäcken deposit on the north edge of the dome. This mineralization is described by Grip as being massive in the lower part and disseminated in the upper; between the upper and lower stratiform ore, there is ore rather similar to the breccia ore at the Adak mine. The conclusion I reach is that the authors might well be writing about two different deposits.

Ljung mentions Brännmyran only under his category (Id) in which he says that the ores at Brännmyran are breccia ores in the two horizons of cordierite-cummingtonite schist where chalcopyrite and pyrrhotite and Co-bearing arsenopyrite act as breccia cements. Grip says that the Brännmyran ores consist of two sheets of ore at slightly different horizons that overlap each other with a third body farther east. The Brännmyran ore, Grip states, lies on the flanks and crest of a minor anticline and are made up of pyrite, arsenopyrite, chalcopyrite, and pyrrhotite with the pyrite and the arsenopyrite early and the other sulfides later. Grip does not mention the brecciated character of the Brännmyran ore and Ljung emphasizes it.

I expect this summary of conflicting views by two highly trained and respected geologists to bring out the problems of telling the story of any ore body in brief; something must be left out under such circumstances, and what is mentioned by one writer may be given a different emphasis or omitted altogether by the other. This, in short, explains why the descriptions of ore deposits in this volume are longer than some geologists think they should be.

The major differences, from the point of view of genetic classification, between the Boliden-Rakkejaur area and the Adak dome area is the presence in the latter of an appreciable percentage of the ore in lime-rich skarn of the Adak formation. This probably requires that the category Hypothermal-2 be included in the classification of the ore bodies of the Adak dome, but the amount of ore in this category almost certainly is only a minor fraction of the total. The alteration minerals in the wall rocks of the Adak dome deposits are largely of a higher temperature of formation than those in the remainder of the Boliden district and suggest that a higher percentage of the ore minerals were deposited in the hypothermal range than in the rest of the Boliden district. The minor quantity of telluride minerals at Rudtjebäck are of much theoretical interest but were not introduced in sufficient amount to justify the inclusion of "leptothermal" in the classification of the Adak dome deposits. Thus, the Adak dome deposits are here categorized as Hypothermal-1, -2 (minor) to Mesothermal.

LAINIJAUR, VÄSTERBOTTEN

Middle Precambrian	Nickel, Copper, Cobalt	Magmatic-2b

Grip, E., 1942, Nickelförekomsten Lainijaur: Geol. Fören. i Stockholm Förh., Bd. 64, H. 3, No. 430, p. 273-276

Grip, E., 1961, Geology of the nickel deposit at Lainijaur in northern Sweden and a summary of other nickel deposits in Sweden: Sveriges Geol. Undersök., Årsbok 55, ser. C, no. 577, 79 p.

Grip, E., 1978, Lainijaur, in Mineral Deposits of Europe, Volume I, Northwest Europe: Inst. Min. and Met. and Mineral. Soc., London, p. 161

Notes

The nickel ore body of Lainijaur is located at 65°14'N, 18°59'E, some 13 kilometers northeast of the village of Malåträsk in the county of Västerbotten; it is 43 kilometers by road from the nearest railroad station at Abborrträsk. The surrounding country is quite flat and is covered with moraine ridges and intervening swamps; boulders of any size are largely lacking in the morainic material. Outcrops are rare; certainly not enough exist to permit the making of an even roughly accurate map of the Precambrian basement. To prospect the area, the Boliden Company carried out electrical prospecting in 1939, followed in 1940 by some test pitting, one of which uncovered copper-pyrrhotite ore at a depth of six meters. Chemical analyses of the ore material encountered showed that it contained even more nickel and some cobalt. Later diamond drilling, with detailed gravimetric and magnetic measurements were made, and the decision to sink a shaft was reached, the shaft begun in 1941 and production commenced; the mine was closed in 1945. The entire output was about 140,000 metric tons, which included 100,000 tons of ore that had an average grade of 2.20% Ni, 0.93% Cu, and 0.1% Co. The mine has been exhausted down to 100 meters below the surface, but ore remains below that depth. Since no mining is known to have been done since 1945, the ore boundaries are uncertain below 100 meters; whether or not mineable ore remains is not stated by Grip (1961).

The Lainijaur deposit lies on a direct line between the Boliden-type deposits of Rakkejaur and those around the Adak dome, about half way between the two areas. Lainijaur is about 80 kilometers west-northwest of Boliden, and lies at the northern edge of the upper Supracrustal rocks (the Phyllite group) where those rocks are cut off against the southeast edge of the Revsund granite that surrounds the Adak Dome. A window of the Phyllite reaches to the Adak Dome from the southwest but is only indirectly connected to the Phyllite rocks of the Lainijaur area.

The sedimentary rocks in the Lainijaur district probably belong to the lower part of the Phyllite group, as it is known in the Skellefte district, and these show a well-marked banded structure that is made up of alternating layers ranging from quartzite to limestone that are, texturally, classified as ranging from rudite to lutite. In addition, a dacite tuff material is common in the lower levels of the mine at least. In the rock volumes that are adjacent to the ore bodies, however, metamorphism is so strong that it is difficult to determine the origin character of the different rock types. Beside the rocks of the lower Phyllite series, the coarse clastic sediments under that series, arkoses and intermediate volcanics, also probably are present.

The sedimentary rocks in the mine area have been folded into a syncline and the beds at the bottom of the synclinal trough have been cut by a gabbro dike that strikes about N45°E and dips about 55°NW and cuts the schists at an angle about 90° to their bedding. The dike can be followed through the mine and throughout most of its length sent out wing-shaped sills on both sides at more or less right angles to the strike of the dike. As the dike is followed up through the sedimentary sequence, these wings increase in thickness until the gabbro assumes a phacolith shape, following the trough of the syncline and spreading out laterally to die out into serrated extensions into the sediments. The central part of the gabbro rose through the sediments to attain a thickness of some 50 meters. The sedimentary beds above the gabbro have been, particularly in the central part, largely replaced by diorite and quartz diorite. Whether these two more silicic rocks were the result of an actual invasion by magma or were produced metasomatically is uncertain, but it is certain that the feeding channel has not been followed above the top of the quartz diorite-diorite mass. There is no sharp contact between the gabbro and the igneous (-type?) rocks above it, and the amount of sulfide material in the diorite-quartz diorite is much lower than in the gabbro. Remains of calcareous beds, however, are found randomly scattered through the thickness of the diorite-quartz diorite, suggesting that the non-carbonate sediments were much more readily assimilated or metasomatized than were the limestones. Most of the limestone bed lenses in the diorite-quartz diorite have been converted to lime-silicate skarn which indicates that they were subjected to high energy forces, as the shaly material must have been if it was converted to rock of definitely igneous appearance.

The sedimentary rocks associated with the Lainijaur deposit differ greatly from one bed to another both in size of grains and in their composition, and the average composition corresponds quite closely to that of quartz diorite, so that not much chemical change would have been necessary to have changed the sediment into the igneous-appearing rock. Most of the sediments not altered (or intruded by) diorite-quartz diorite are now amphibolites, with the prominent banding in the sediments having been preserved in the different colors of the metamorphic minerals produced in one band as opposed to another. The limestone-banded portions of the sediments have been converted to lime-silicate skarn, providing additional evidence that the associated igneous-like rocks were the result of granitization and not crystallization from a magma.

Arenaceous beds, having a composition of dacite, have, in several places, been recrystallized to an igneous-appearing rock; even more intense "granitization" has converted these beds into quartz diorite; some mobility was imparted to this material, as quartz diorite is found as secretions and veins in both sediments and gabbros. Grip suggests that ore and mineralizing solutions were given off by the partially melted sediments during their later solidification - this matter is not discussed further, so what Grip thinks that these mineralizing solutions may have formed is not clear. Where granitization is associated with ore, the most intense aspects of that process are in the immediate area of the ore masses, and very fine disseminations of ore minerals occur in the granitized beds, particularly in secondary chlorite and biotite. The fact that the granitization is more prolific adjacent to an ore mass than to the gabbro strongly suggests the granitizing solutions were given off by the crystallizing molten sulfides and not from the crystallizing gabbro magma.

Below the 150-meter level, just beneath the ore, the sediments have been variously granitized, with the different beds in the banded sedimentary rocks having been recrystallized differently, depending on their initial composition which appears to have been dacitic, that is, dacitic tuffs and perhaps fragments eroded from similar lavas. Some minor bands, in the less banded portions, may be largely clinozoisite-epidote. Some of the dacitic sediments have been converted to a fine felted mass of chlorite, uralite, biotite, plagioclase, quartz, calcite, and disseminated ore minerals. Some of the recrystallized sediment has a somewhat granitic texture with plagioclase as the dominant mineral and several accessories including ore minerals. This granitization, taken in a broad sense, differs markedly from one location to another, a definite indication that it was the result of metasomatic processes and not full or partial melting.

The gabbro dike that marks the path followed by the gabbro magma is 1.5 to eight meters thick but widens where the gabbro body changes from dike to phacolith. The contacts of the gabbro dike with the enclosing normally schists are quite sharp although the schists are slightly amphibitolized along them and, in some places where amphibolite is well-developed, the exact contact is difficult to locate. The contact between gabbro and schists in most places is marked by variously definite slickensides along which movement has taken place in several directions at quite different times.

The gabbro intrusion took place in two, or in some places, three stages; the gabbro near the enclosing walls is fine-grained. In the usual case, the center portion of the gabbro is broken; following this additional magma was intruded, and some fragments of the fine-grained wall gabbro were incorporated in the second stage of the intrusion. This second, younger gabbro is much coarser than the first. Some of the gabbro shows an ophitic texture with the laths of plagioclase being as much as four millimeters long and being enclosed in a matrix of biotite and hornblende. In this rock, pentlandite-bearing pyrrhotite is disseminated in major amounts and normally occurs with biotite, but hornblende needles may cut the pyrrhotite. In most of the gabbro, the main femic mineral is augite, and it is this mineral that is altered to produce the hornblende.

Sills branch off from the dike, and these range in thickness between one and two meters. This gabbro is of the first intrusion and is not as rich in pyrrhotite as is the second-stage gabbro and may be even lower in that mineral than the border gabbro. In the sills, the gabbro is more strongly converted to amphibolite and, in many places, is difficult to distinguish from the amphibolite wall rock.

On both sides of the fault feeder, the gabbro is intruded outward conformably with the sediments, but the gabbro normally is blocked by wedges and fragments of schist that have been variably assimilated. In the vertical dimension, the gabbro varied in composition, mainly because of the different proportions of schist fragments that it contains. In places reactions, apparently between schist and gabbro, have converted both into fine-grained amphibolite, portions of which contain relict sedimentary structures. In the upper part of the phacolith, the normally sized gabbro may include several zones of very coarse gabbro. The gabbro on the footwall is very fine-grained and may include crystallized immiscible droplets of pyrrhotite. One mineralogical composition of the gabbro is 46 per cent hornblende, 41 per cent plagioclase, six per cent augite, one per cent biotite, and two per cent ore minerals, plus four per cent talc, chlorite, serpentine, and calcite. Other analyses range around these quantities. Only locally is olivine an important mineral (eight per cent in one sample); ore minerals may run as high as 17 per cent.

Diamond drilling has established that quartz diorite lies above ore-bearing gabbro phacolith; the contact generally is parallel to the lower contact of the gabbro and to the bedding of the sediments. The quartz diorite is at least 90 meters thick but changes in its upper portions into granodiorite; or, to a minor extent, even to granite; the quartz content of the rock decreases downward. The quartz diorite, appreciably more than the gabbro contains randomly distributed fragments of the sedimentary series, the fragments in many instances having been at least partly assimilated; light fragments are less affected than darker ones by this process. Fragments of lime-silicate skarn also are present in the quartz diorite, but these probably have made no appreciable contribution to the igneous-appearing rock. The contact of quartz diorite with gabbro is diffuse over distances of up to a meter. How much the sediments contributed to the total bulk of the quartz diorite has not been estimated by Grip, but the total is not small.

The grain size of the quartz diorite is varied, ranging around five millimeters, but much coarser material is known; plagioclase is the dominant mineral, and the zoned laths have cores of An_{55} and margins of as low as An_{10-20}. The core of the plagioclases always is saussuritized. The granodioritic phase contains some quartz and alkali-feldspar micropegmatite. Biotite and hornblende are the principal dark minerals, the quantities of which are roughly equal; some of the biotite was produced by alteration of the hornblende. Several ore minerals are accessories in the quartz diorite - pyrrhotite, arsenopyrite, magnetite, apatite, and titanite. The parent material of the quartz diorite probably was further on the mafic side than the final product, the less mafic minerals probably having been produced through the assimilation of sediments. It seems unreasonable to assume that the minor amounts of granite can be related to the huge mass of the Adak granite only about one kilometer away to the northwest.

The gabbro, in both the dike and the overlying phacolith contains disseminated pyrrhotite that has inclusions of pentlandite and minor amounts of chalcopyrite; this disseminated material locally is rich enough in Ni and Cu to form ore. This disseminated ore appears to have been introduced with the gabbro magma and to have crystallized in the same time span. At some stage, the Fe-Ni-Cu material appears to have separated immiscibly from the gabbro and to have formed droplets of sulfides that crystallized after the silicates.

The principal type of ore in the Lainijaur deposit is a breccia of gabbro cemented by pentlandite-containing pyrrhotite and chalcopyrite and is reasonably designated as breccia ore. Still younger than this post-gabbro solidification ore is vein ore of a complex type that is denominated as nickel-arsenic ore. Although it is quite small in amount compared with the breccia ore, its high nickel content allowed it to make an important contribution to the total nickel recovered.

The coarser, second-stage gabbro is the rock that contains the disseminated ore, the finer-grained gabbro being quite low in disseminated sulfides. The richest breccia ore is present in the dike and in the central portion of the phacolith. Directly above solid ore, a considerable amount of disseminated sulfide is found, with the amount of ore material gradually dying out upward; the amount of sulfide in the gabbro above the massive ore also decreases toward the margins of the gabbro phacolith. The highest concentrations of sulfides make up 50 per cent of the rock volumes containing them. In these, the principal mineral is pyrrhotite with strings and grains of pentlandite. The chalcopyrite, as is so common in deposits of the magmatic-sulfide type, is not normally enclosed in the pyrrhotite but appears to have separated from the iron-nickel-sulfide melt while both phases still were molten and crystallized after it (the pyrrhotite-pentlandite). Although the chalcopyrite crystallized later than the iron and iron-nickel sulfides, all three usually are in the same general rock volumes with chalcopyrite veinlets in many places cutting the other two ore minerals.

In the lower part of the sulfide-bearing gabbro are masses of magnetite of appreciable size; at least one of these masses of magnetite contained small inclusions of gabbro and considerable disseminated arsenopyrite. Cobalt in some areas is 10 per cent of the nickel content but normally is lower than that. Zinc averages 0.064 per cent, with the highest report assay being 0.39 per cent; the grade of zinc is completely independent of that of copper in the disseminated ore in gabbro.

On both the east and west sides of the gabbro dike, in the contact zone between gabbro-dike material and the underlying sedimentary rocks, are a total of two bodies of massive nickeliferous pyrrhotite. The contacts of these ore bodies against the gabbro are sharp and are marked by slickensides. Where the ore touches the gabbro dike, the contacts are sharp, but between the hanging and footwalls of the ore on either side of the dike, both the hanging- and footwalls are very irregular because of the intense brecciation developed along them.

The eastern, and apparently larger, ore body, in its upper and most southerly part, has the form of a thin sheet, but, as it is followed downward and toward the north, it is compressed toward the west (the direction of the gabbro dike) and becomes steadily smaller until, near its northern end, it enlarges appreciably. Toward the east, the eastern ore body becomes thinner and gradually grades out into breccia ore. In this breccia ore, the sulfide content steadily decreases as it, in turn, is followed to the east. The plunge of the ore body undulates gently, but it generally follows the regular or irregular bedding plane of the sediments no matter how contorted the bedding plane may be. From the 70-meter level, down to the ultimate depth of the ore (at least 185 meters below the surface) the sedimentary rocks were compressed in the direction of the plunge of the ore body to produce an imbricate structure that is followed by the ore. This branch of the ore, on the 100-meter level flattens and dies out. At 110 meters down, however, an en echelon continuation appears which lasts for only a short distance to the 115-meter level. On the 125-meter level, however, more ore appears that is wider and is located along the general direction of the plunge although it locally is discordant to the bedding. Ore is known as far down as the 185-meter level. The shape of this ore body changes markedly from one cross-section to the next.

The western ore body is generally narrower than the eastern one but in a few places may be thicker.

As is true of the eastern ore body, the western one has sharp contacts with the gabbro dike but feathers out into the schist to the west. Both of the ore bodies are similar in plunge, and both have been interrupted by badly disturbed sediments about the 100-meter level. The upper (southern) part of the western ore body is very flat with the schist forming the footwall, but a tongue of ore is contained between two gabbro sills that extended westward from the main dike. Ten meters farther north, the ore body is larger and more compact and has a quite straight boundary with the gabbro dike. In the next more northerly of Grip's sections the ore body is shown to be divided into several units separated by schist. The position of the ore body is partly controlled by the bedding in the sediments but also by tectonic structures that cut that host rock. Still farther north and farther down, the ore body is appreciably smaller and more regular. Even farther north and deeper down, the ore is five meters thick and has a sharp contact with the gabbro dike to the east; to the west, the ore tapers out into gabbro, a most uncommon feature.

In connection with these two rather elongated ore bodies (eastern and western), veins and lenses of nickel arsenic minerals and of chalcopyrite are present. These less common ore masses are most prevalent in the lower part of main ore bodies and in the schists adjacent to these lower portions of the ore. In some places, these veins cut the massive ore; rarely, narrow nickel-arsenic veins are present in the higher reaches of the gabbro. Both the nickel-arsenic and the chalcopyrite bodies appear to have been late segregations from the main pyrrhotite-pentlandite ore and to have remained molten until the pyrrhotite was solid enough to fracture. Many of these veins follow pronounced fissures in the gabbro where it is appreciably mineralized by disseminated sulfides; appreciable replacement of the host rock has taken place along the margins of these veins through reaction between molten material and the solid gabbro. Nickel-pyrrhotite ore quite commonly replaces brecciated schist and some chalcopyrite has been expelled (probably by molten unmixing to the contacts between pyrrhotite and schist). Sulfide ore also replaces both fine-and coarse-grained gabbro in the dike. In this replacement process, where it takes place on a major scale, the gabbro remnants are converted to well-rounded fragments, all contained in a sea of pyrrhotite and known as pellet ore. The gabbro fragments are not much altered, and their pyroxene still is partly preserved. Shale fragments, where they are replaced by ore, normally retain their angular forms no matter how intense the replacement. Some of the pyrrhotite in the gabbro is there as rounded droplets that appear to have separated in the molten state but not to have crystallized until the silicates of the gabbro had done so. Thus, the attack on gabbro was by a molten mixture of pyrrhotite and pentlandite rather than by any form of hydrothermal solution. Most of the pyrrhotite contains fragments of the rock in which it is contained. The pyrrhotite grain size ranges from fine in the apophyses and along contacts to quite coarse in the centers of ore veins and masses. Ordinarily any structures in the pyrrhotite, such as zoning, are parallel to the contacts, although in rock volumes in which tectonic structures are impressive, a definite vertical linearity may be present.

The essential minerals of the pyrrhotite ore are: pyrrhotite, pentlandite, chalcopyrite, and magnetite, typical of ore formed from magmatic sulfide melts. The accessories are arsenopyrite, cobaltite [Co,Fe(AsS)], and other Ni-Co arsenides. The pyrrhotite ore grades gradually into nickel-arsenic ore (where that ore exists), sharp boundaries being essentially lacking.

The first mineral to crystallize in the pyrrhotite ore was magnetite (the normal relationship in all magmatic sulfide deposits); ordinarily it does not show crystal form, being corroded by the sulfides that remained molten after the magnetite had precipitated. Cobaltite occurs in random fashion in sharp idiomorphic crystals and may contain inclusions of silicate minerals, magnetite, and pyrrhotite. Pyrrhotite is, of course, the dominant mineral in this ore type; locally it shows twining lamellae; it is strongly magnetic. Pentlandite is, of course, the major nickel mineral and, almost everywhere is directly connected with pyrrhotite. Pentlandite may be scattered through pyrrhotite as idiomorphic strings crystals or in gashes in the pyrrhotite; these appear to have resulted from unmixing from solid solution of pentlandite from pyrrhotite. Replacement pentlandite in feathery form also appears in pyrrhotite. Some pentlandite was completely excreted from pentlandite and may appear surrounding magnetite or cobaltite.

Chalcopyrite is found only in (or in close association with) pyrrhotite-pentlandite ore; it occurs along grain boundaries or in fissures and, in places, as actual veinlets cutting both the ore and the bordering schists. All of this chalcopyrite appears originally to have been solution in the molten pyrrhotite-pentlandite solid solution; it was separated from the Fe-Ni-S melt as molten chalcopyrite that normally did not solidify until after the pyrrhotite-pentlandite molten solution, thus accounting for its relationships to the Fe-Ni-S minerals described above. Bravoite (Fe,NiS_2) is an alteration product of pentlandite and is located along fissures and cleavage faces in pentlandite.

The frequency distribution of copper-nickel ratios in pyrrhotite-pentlandite lies around a mode of $\frac{Cu}{Cu + Ni} = 0.1$, whereas this ratio is appreciably higher in the disseminated ore.

Nickel-arsenic ore also separated from the Fe-Ni-S molten solution and is the last of the three ore types to crystallize. This ore was emplaced as fissure fillings that cut both the older ores and the wall rock; in fact, most Ni-As ore is in fissures in the sedimentary rocks (schists) immediately below the Fe-Ni ore and most particularly at the ends of the ore bodies. The minerals in the Ni-As ore (in the general order of their crystallization) are: ilmenite and magnetite in minimal amounts, cobaltite, skutterudite $[(Co,Ni)\text{-}As_3]$, rammelsbergite-löllingite $[NiAs_2\text{-}FeAs_2]$, arsenopyrite, tourmaline, niccolite [NiAs], pyrrhotite (as a result of minor amounts of that mineral being dissolved in the Ni-As solution when it separates from the Fe-Ni-S solution), chalcopyrite (it is questionable if this mineral, though closely associated with the minerals of the Ni-As ore, actually is present in the Ni-As solution; rather the chalcopyrite in a separate molten phase may have mixed up with the Ni-As molten solution), valleriite $[4(Fe,Cu)\cdot 3(Mg,Al)(OH)_2]$, probably the most peculiar copper mineral known, very rare scheelite and molybdenite, sphalerite and galena at the ends of the Ni-As veins where Pbs appears to have been the last mineral to have crystallized from the Ni-As melt, and minor quartz and carbonate minerals.

Grip believes that three ore types can be distinguished, as already has been discussed. These are: (1) disseminated ore, (2) nickel-pyrrhotite ore, and (3) Ni-As ore. The disseminated ore came out of solution in the gabbro magma before either melt had begun to crystallize. The gabbro magma crystallized before the separate sulfide phase, some of which was trapped by the silicate minerals as they crystallized and never fully separated from the gabbro. On the other hand, much of the molten Fe-Ni-S melt was able to separate completely from the silicate portion of the parent magma and collected at the bottom of the magma chamber (normally in contact with schist). There it remained in the molten state until after the silicates had crystallized. Much of this melt then crystallized where it had collected, but other amounts of it were driven or flowed into fractures in the schists (and to a lesser extent in the gabbro). Thus, it appears that the Fe-Ni-S melt was generated in the gabbro magma, some of it crystallizing among the mesh of silicate minerals and the remainder of it reaching the bottom of the magma chamber where part of it crystallized and the remainder was driven into fractures developed in the adjacent schist and (to a lesser extent) in the gabbro. None of the Fe-Ni-As melt, so far as can be seen from Grip's description, came out of solution at depth and was forced up into the fractured volumes of schist and gabbro above.

The composition of the disseminated ore is appreciably different from that in the massive Fe-Ni-S ore and from that in the Ni-As ore. In fact this difference seems to be such as the two varieties just mentioned could have been produced from the first by splitting an Fe-Ni-As-S ore into two immiscible halves, with the Fe-Ni-S ore having crystallized first of the two and the Ni-As ore later. To complicate the picture even more, the Fe-Ni-S molten solution contained enough Cu that some time before it began to crystallize it separated into an iron-nickel-rich fraction and into a copper-rich one, the former crystallizing before the latter. Grip believes that "during the crystallization of this nickel-pyrrhotite ore with a very simple composition the solutions rich in arsenic and metals were driven out peripherally and from these the minerals then crystallized, forming nickel-arsenic ore and ore pegmatites in the periphery of the nickel-pyrrhotite ore-bodies. Formed during the pneumatolytic-hydrothermal stage, here occur all transitional forms from nickel-pyrrhotite ore to low-temperature formations with sphalerite, galena, quartz and calcite. A surplus of volatiles had gone out from the distal extremities of the ore-bodies and migrated along fissures causing granitization and other alterations in convenient parts of the sedimentary series".

This scheme of Grip's seems overly complicated for the problems that it is required to explain. It seems much more reasonable to me that the minerals of the Ni-As ore were, at some stage, completely dissolved in the Fe-Ni-S molten solution, separated from that prior to, or early in, the crystallization cycle of the Fe-Ni-S melt, and remained as a separate molten phase until after the Fe-Ni-S phase had crystallized and then began itself to crystallize. This Ni-As phase may have contained some water (at the end perhaps of a separate phase) but probably nothing but the quartz and calcite crystallized from it. The difference in the composition of the disseminated Fe-Ni-S globules in the gabbro from that of the massive Fe-Ni-S ore probably was due to the separation and crystallization of the disseminated globules having taken place at a high enough temperature that this molten solution had not been separated when it came out of solution in the gabbro and crystallized quickly enough after that separation that it did not have time to separate into: (1) an Fe-Ni-S phase and an Ni-As phase as did the molten solution from which the massive Fe-Ni-S ore and the Ni-As ore materials later separated in the molten state with the Fe-Ni-S solution crystallizing first and the Ni-As crystallizing later.

The Lainijaur area was subjected to some metamorphism after ore deposition, but it was low grade, not having advanced beyond the chlorite-sericite stage; carbonates were unaltered and no lime-silicate skarn was formed.

All the products of metamorphism found in the mine were the result of the intrusion of the mafic magma and its associated ore formation. Assimilation of the wall rock produced the more silicic rocks that now overlie the gabbro and ore with the limey beds in the original sediments having been converted thereby into lime skarns (produced hydrothermally and not by regional metamorphism) both in the gabbro and the quartz diorite and diorite. What proportion of the gabbro was so produced from sediments and what was true gabbroic magma it is impossible now even to estimate. Certainly hydrothermal fluids must have been given off in some quantity from the truly magmatic portion of the gabbro magma because evidences of granitization are found in the schists around vein terminations. As would be expected from hydrothermal solutions given off from a gabbro magma, these were so rich in ferrous and non-ferrous metals, sulfur, and arsenic that some deposition must have resulted from them. But this does not mean that the Ni-As ores were formed by such solutions. It must be emphasized that the gabbro was intruded in several stages, so that there may have been appreciable differences between the sulfide mix at one stage from that obtaining during the others, so some of the differences in elemental composition from one ore type to another may have been, at least in part, related to this factor.

It appears almost certain that the igneous and ore-forming events in the Lainijaur area occurred during Svecofennian phase of middle Precambrian time and that the age of the ore emplacement must have, therefore, been middle Precambrian.

The manner in which the ores were produced in the Lainijaur area must have been such as to justify their classification as magmatic early separation, late solidification so they should be classified as Magmatic-2b.

STEKENJOKK, VÄSTERBOTTEN-JÄMTLAND

Middle Paleozoic	Zinc, Copper, Gold, Silver, Sulfur	Mesothermal, Metamorphic-C or Sedimentary-A3, Metamorphic-C

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Notes

The Stekenjokk ore area is located in west-central Sweden, and at no point is it more than 7 km from the Norwegian border. The area contains three ore bodies, Stekenjokk proper, Levi [which lies on the opposite (north) side of the antiform that separates Levi from Stekenjokk to the south] about 1.7 km northwest of the northeastern tip of Stekenjokk. The third ore body is Tjokkola that lies parallel to the Stekenjokk ore body and is about 1.5 km southwest of it. The antiform between the first two bodies mentioned is the Stekenjokk antiform, a structure that follows a slightly curved trace that strikes generally northeast-southwest. The length of the horizontal projection of the Stekenjokk deposit is about 4.5 km and that of the Levi deposits is about 3 km. The Tjokkola body must have a much shorter length than the other two since it contains only about 1/150th of the tonnage of ore in Stekenjokk and about 1/46th of that of Levi. The boundary between Jämtland (SW) and Västerbotten (NE) cuts through the Stekenjokk ore body about 1.1 km from its southwestern end.

Grip (1978) reports that mining began at Stekenjokk in 1976 and that it was planned to produce about 400,000 tons of ore each year. Whether all mining was to be from the Stekenjokk body or some was to come from Levi is not clear, although the much lower grade of Levi argues against it. The estimates of grade and tonnage are as follows:

Deposit	Million tons Reserves	Au ppm	Ag ppm	Cu %	Zn %	Pb %	S %
Stekenjokk	15.1	0.3	55	1.5	3.0	0.3	20
Levi	4.6	±0.1	20	1.2	1.6	±0.1	16
Tjokkola	0.1	---	--	1.1	1.9	---	6

The unmineralized gap between Stekenjokk and Levi is about 1.2 km. The mine is located about 20 km west of the village of Vilhelmina (64°38'N, 16°40'E).

The major tectonic unit in the area of the Stekenjokk ores is the Seve-Köli nappe that is known from as far north as Sulitjelma and as far south as Røros and has been thrust more or less eastward for more than 100 km toward the foreland. This nappe, at its eastern extremity, lies on Eocambrian sparagmite (arkoses) and quartzites or lower Paleozoic rocks of miogeosynclinal affinities. These Paleozoic rocks, at least, belong to lower tectonic units and contain, in some places, slices of crystalline basement.

The Seve-Köli nappe complex is, in its lower Seve part, composed of high-grade metamorphic rocks, whereas the Köli (upper) portion is made up of lower Paleozoic eugeosynclinal rocks that have been metamorphosed to the greenschist or, at most, to the lower amphibolite facies. Both, however, are part of the same major tectonic unit. Within the Köli nappe several minor (second order) nappes, such as the Leipek and Gellvernokko nappes, have been identified; the latter is directly involved in the Stekenjokk area, whereas the leading edge of the former is a short distance (less than 2 km) to the west.

The Stekenjokk ore body (in the narrow sense) ends (to the north) against the first synform east of the Stekenjokk antiform. The ores of Stekenjokk are contained in the lower part of the silicic and mafic volcanics that form the upper part of the Lasterfjäll group. By correlation with rocks farther to the east, the corals and brachiopods that these latter rocks contain date the Stekenjokk volcanics as uppermost Ordovician and the graptolites place them in the lower Silurian; the latter age is assigned to the Stekenjokk volcanics.

The complete stratigraphy of the rocks in the Köli portion of the major thrust is as follows (from top to bottom) (Zachrisson, 1971):

Group	Formal Lithostratigraphic units	Lithologies
Remdalen group	Frems phyllite	Dark, quartzose and graphitic phyllites
	Remdalen greenschist	Mafic volcanics, quartz porphyry, limestone afic volcanics
	Portfjäll conglomerate	
	Remdalen qtz conglomerate	quartz conglomerate, quartzite, or quartz phyllite
Lasterfjäll group	Basalt & qtz-keratophyre formation and Lasterfjäll qtz-keratophyre-bearing formation and Stekenjokk qtz-keratophyre	graphitic phyllites, mafic and (dominantly) silicic volcanic rocks, intercalated seds (tuffites and graphitic phyllites)
	Blåsjö, Lasterfjäll, and Lövfjäll phyllite	calcareous phyllites (in western synform with gabbro intrusions)
	Broken series	dark, graphitic phyllites, mafic volcanics, and green tuffites
	Slätdal ls	limestone
	Bellovare fm	
	Vojtaja cong	qtzite or qtzite
Tjopasi group	Gilliks "series"	dark phyllites or coarse clastic rocks locally polymict conglomerates
	Seima "series"	mafic and silicic volcanics dark phyllites
	Rotic conglomerate	serpentinite conglomerates
	Ro conglomerates	qtzite conglomerates
		varied sedimentary, tuffitic and volcanic rocks

The rocks in the Stekenjokk area have been affected by two periods of deformation. The second period produced four major, north-trending folds (from west to east): (1) the Riksgränsen antiform; (2) the Western synform; (3) the Fjällfjäll antiform; and (4) the Eastern synform. They have produced folding of the compositional banding, the regional schistosity, and the thrust sheets. The S_1 schistosity (a deformed penetrative variety) occurred as an early phase of the F_1 isoclinal folding; locally, minor F_1 folds were preserved - at times these are nearly regional in dimension. In the north, the axes of the F_1 folds are nearly east-west but they turn to the NW-SE as the Western synform is followed to the south. It also is possible that the low and even more regional low-angle thrust faults (e.g. the Gellvernokko nappe) are associated with the F_1 folding. The entire Seve-Köli nappe complex was thrust east before the F_2 folding that developed more open folds that are oriented NE-SW to NNE-SSW; these folds normally have steep-dipping axial surfaces.

The Stekenjokk area proper is part of the Western synform and is located in a major culmination zone to the east of the Børgefjell window (in which the basement is exposed). A pronounced axial plunge toward the Remdalen synform takes place to the NNE of the window, whereas to the southwest, the unfolded Gellvernokko and Leipik nappes both are in the central part of the Western synform.

The Köli rocks in the Stekenjokk area proper include the lower two groups of the stratigraphic sequence given above - the Tjopasi and Lasterfjäll groups. In the regional picture, however, the calcareous phyllite of middle Lasterfjäll age (that has been intruded by gabbro) is a most important key unit. The most common rock type in this phyllite is a sericite/chlorite-rich pelite or psammo-pelite that may be banded or may be too poorly sorted for that structure to be visible. The calcite content ranges from a few to as much as 30 per cent. Into the upper layers of these phyllites, a considerable number stratiform lenses of metagabbro were intruded; the border zones of these gabbros are strongly schistose and much resemble volcanic greenschists. In the centers of the larger gabbro lenses, the coarser-grained primary texture has been preserved with the plagioclases converted to saussurite and the hornblendes largely altered to actinolite or even to chlorite. The very highest phyllite lenses are appreciably different from those below them, containing grayish graphitic horizons that cause major electromagnetic anomalies (Zachrisson, 1971).

The ore mineralization is composed of stratiform layers of sulfides that are interbedded with a series of silicic volcanic tuffs, tuffaceous sediments, and fine, in many places, carbon-rich detrital and chemical sediments. The gabbro sills also are stratiform but with strongly foliated margins. The sulfides do not appear to favor one host rock over the others, but they do tend to be on or near the borders of different depositional facies of the host rocks. The gangue associated with the sulfides differs appreciably in composition so that it is difficult to distinguish the exact contact between hanging- and footwall units as the contact between layers may actually be within the ore. In other words, the ores were emplaced, whenever they were emplaced, outward from the contact between host-rock layers in both the directions of foot- and hanging walls. Thus, if the sulfides were syngenetically deposited, they began deposition in the later stages of the formation of one sedimentary layer and continued to be laid down after the character of the host sediments had changed appreciably. If, on the contrary, the sulfides were introduced after the sediments had been lithified but before they had undergone their most intense folding, the replacing solutions must have moved along the bedding planes and penetrated more or less equally above and below the plane dividing the two rock types. As the mineralization is followed out from the center of the sulfide band, the bands become less frequent and thinner. At Stekenjokk, the main ore horizon lies near the contact of silicic pyroclastic rocks and a black phyllite unit. All of the sulfide layers appear to be remarkably persistent parallel to the primary depositional layering of the host rock (Juve, 1974).

Zachrisson (1971) make the point (quoted by Juve, 1974) "that the regional major fold phases are of fundamental importance in understanding the geometry of the ore bodies". Thus, the general shape and thickness of the ore body in cross section and its extent along the main axes of the ore body are controlled by the F_1 phase of folding with the ore bodies being situated in the hinges and limbs of the major F_1 folds. The later (F_2) folding is superimposed on, and refolds, the F_1 structures that contain the ore bodies. Juve's investigations confirm Zachrisson's observations as to the fold and ore relationships. The F_1 folds are isoclinal and have quite flat-lying axial planes with amplitudes that range from hundreds of meters to those are measured on a micro-scale. The locations of stress minimums around the crests of folds and in drag folds also are the places where the ores are thickest. Thus, a thin sulfide layer may be followed from its position on a fold limb into sulfide masses 30 times as thick in the hinge zones. Juve points out that the main Stekenjokk ore body is situated in the crest of a major F_1 fold so that it is sickle-shaped in cross section with its axis coinciding with the axis of the F_1 fold. Juve says that all mineral layers that he thinks pre-existed before the folding have maintained the same mineralogy and banding. He does say, however, that, in detail, he can detect what he calls a selective mobilization and redeposition by which the more mobile minerals, chalcopyrite and sulfosalts, have moved more readily into these zones of stress minimums. Zachrisson (1971), on the contrary, concludes that the strong tectonic control of the ore by the F_1 folding and certain alterations to the minerals of the host sediments suggest an early hydrothermal epigenesis for the ore bodies. Zachrisson, however, also said that the apparent pre-F_1 age, the distinct banding, and the nearly perfect stratigraphic control indicate a syngenetic emplacement of the ore sulfides. This matter will be considered further below.

In all the wall rocks, there is a marked penetrative schistosity parallel to the compositional banding and the F_1 axial plane. This schistosity is the result of the preferred orientation of sheet silicates and is most impressive in the softer rocks such as black phyllite and chlorite-sericite schist. Locally, an older schistosity can be seen; it also is parallel to the compositional banding but has been deformed around the hinges of the F_1 folds. This older schistosity is best developed in the more massive quartz keratophyres.

During the F_1 stage many small thrust faults developed along the borders of rocks of different degrees of competence; locally brecciation was produced along such contacts. Fold-thrusts fault along the limbs of the greater isoclinal folds also are seen in some of the ore sections.

The second fold phase (F_2) refolded the isoclinal structures of F_1 stage into open symmetrical folds that have vertical to subvertical axial planes. Where these folds are studied in cross-sections perpendicular to the ore axis, it is seen that the amplitude of these F_2 folds ranges from meters to several tens of meters. Smaller parasitic folds repeat this folding pattern in minor and micro scales. The gently undulating course of the main Stekenjokk ore body shows an even larger amplitude of F_2 folding. Juve does not believe that this F_2 folding gave rise to any ore- or rock-thickening of importance nor did this folding cause internal secretions of ore minerals. Some small fissures did form parallel to the F_2 axial cleavage these are filled by quartz, calcite, and some sulfides, principally pyrrhotite. Juve thinks these minerals were deposited at low temperatures from aqueous solutions. In comparison with the F_1 phase, little mechanical transport or metamorphic grain-growth was accomplished; some pyrite cubes, however, were rounded, fractured, and shattered during the F_2 deformation.

Juve summarizes the effect of the two stages of folding by saying: (1) that the first fold-stage produced isoclinal folds with flat-lying axial planes these resulted in important thickening of the ore in zones of minimum stress that were developed around crests of folds and in parasitic drag structures - the main axis of the deposit coincides with the F_1 fold axis and (2) that the second fold phase (F_2) restructured the ore bodies through open folding that produced little, if any, remobilization of sulfides.

A considerable proportion of the total volume of the Lasterfjäll group is made up of the Lasterfjäll and Stekenjokk quartz keratophyres. Much of what appears originally to have been quartz keratophyre is now chlorite-sericite quartzites. The quartz keratophyres are considered by Juve (1974) to be silicic, fine-grained pyroclastic rocks to which the term - quartz keratophyre - can be reasonably applied. He describes three types of quartz keratophyre, as follows:

(1) hard, dense, fine-grained, fine-banded rock in which foliation is marked by chlorite and sericite; some carbonaceous material may be present. The color of the rocks is white to light gray with greenish, bluish, and yellow tints. This rock type grades into other varieties with higher sericite-chlorite contents.

(2) porphyroidal varieties of the following types: (a) with feldspar laths, locally having anhedral outlines; (b) with quartz megacrysts that normally are of a clear blue color; and (c) a combination of (a) and (b).

(3) greenish quartz-keratophyre, low in quartz and gradually changing to green tuffite.

The porphyroidal horizons are of different thicknesses, occur on several levels, and usually show good persistence along strike.

The main minerals of the quartz keratophyres are albite, quartz, chlorite, and muscovite (sericite). Minor minerals and accessories are: biotite, epidote/clinozoisite, calcite, actinolite, titanite, traces of apatite, zircon, rutile, magnetite, and ilmenite. Quartz megacrysts seldom amount to 5 per cent of the rock; if the percentage of these megacrysts reaches 10 per cent, they are appreciably larger than the grain size of the matrix.

The greenschists are fine-grained, green foliated rocks in which Mg, Ca, and Fe are relatively high; Al is medium, and Si is medium to low; Juve considers them to be metamorphosed mafic tuffs. The greenschists grade gradually into green tuffites. They (both greenschists and tuffites) are mainly composed of albite and chlorite with epidote/clinozoisite being noteable but highly varied in amount.

The tuffite makes up a large share of the Lasterfjäll group and appears to be composed of varied proportions of volcanic and terrigenous sediments with the different materials being present in alternating bands. The width of these bands may be on a millimeter, centimeter, or decimeter scale. Juve distinguishes the following types of tuffites: (1) green tuffite in which green chlorite is the main mineral and in which some carbonate and/or epidote is present; an important subvariety is mottled tuffite that is green with white patches and is high in carbonate with or without accompanying epidote; (2) gray tuffite which is mainly colorless or gray chlorite, micas, and/or minor carbonaceous material; and (3) graphitic tuffite that contains considerable carbonaceous material and appears to be transitional to black phyllite. Although some of these types can be followed over considerably distances, most of them are so mixed up as to be difficult or more likely impossible to follow. In many places, the host rocks of the ores are tuffitic in character. When these rocks are mineralized, the MgO and C contents have been considerably increased, the K_2O weakly increased, and the Na_2O content markedly reduced. Juve considers that these conditions may reflect differences in depositional loci or possibly the effects of metamorphism. He makes no mention of the possibility that these conditions are the result of hydrothermal introduction of the sulfides, combined with an appropriate altering effect on the minerals of the host rock.

The black phyllite is one of the two rocks (silicic pyroclastics are the other) near the contacts of which most of the ore is located. This rock is colored by organic material with only a small part of it being graphite. The organic materials have not been identified. The mean carbon content of this phyllite is 5.4 per cent; the highest assays are between 13 and 16 per cent, and the lower carbon limit for this type of rock is between 0.5 and 1.5 per cent. The phyllites are divided into pelitic, quartzitic, and calcareous types, and they may grade into most of the other rock varieties. The sulfur content of these phyllites is low, perhaps averaging about 3.0 per cent; the quartz-rich types are poorer in ore minerals than the others. The strongly deformed black phyllites have been given varied thicknesses and, in some hinge folds, they are completely chaotic masses.

The limestone normally is in thin (0.1 to 0.5 m) beds that are difficult to trace from one drill hole to another; they do not make good marker beds. The limestone is mainly $CaCO_3$ but contains magnesite and ankerite as accessories - no dolomite has been noted; locally small amounts of organic carbon are present, but usually it is absent, but, where present, it makes vague blue shadows that mark depositional layering. Where these bands have been studied, they show that the limestone has been plastically deformed by folding.

The gabbro is in quite thin sills with apparently few, if any, discordant structures. The gabbro is markedly schistose with the schistosity being parallel to the foliation of the surrounding rocks. The border zones between the gabbros and their host rocks are characterized by a lighter border facies probably developed by deformation and metamorphism. Juve thinks that, in many instances the contacts give no evidence of intrusion. Although the schistosity may be developed throughout a given sill, its intensity is normally less away from the contacts. The texture in the centers of the sill, plagioclase laths and randomly oriented or pilotaxitic hornblende, suggest to Juve that they had an eruptive origin. The alteration of these minerals already has been mentioned. Some accessory pyrrhotite and pyrite usually are found in the gabbro; locally richer pyrrhotite masses with some chalcopyrite are present in the contact zones of the gabtro sills. Juve says that the sulfide content may be interpreted in the following ways: (1) a primary constitutent of the gabbro and (2) derived from the host rocks by: (a) contamination during the intrusion; (b) contamination by remobilization and redeposition during metamorphism. Most gabbros contain less than 0.5 per cent sulfur.

The Stekenjokk deposits also contain two types of quartz veins. Those of the first generation are concordant to subconcordant and are composed in part of primary material and in part what Juve calls mobilized wall-rock material. Locally, these veins have been intensively reworked by (Juve believes) metamorphic processes, and direct transitions are known between these veins and quartz-rich volumes of the keratophyres that surround them. Veins of the second generation fill cross-cutting tension cracks that were produced by late-stage metamorphism. Juve considers that this quartz was deposited from aqueous solutions.

Juve (1974) divided the Stekenjokk ores in four main types : (1a) massive, banded, fine-grained pyritic ores; (1b) massive to disseminated, brecciated pyrrhotite-chalcopyrite ore with or without pyrite; (2a) disseminated in light-colored rocks; and (2b) other disseminated ores in black phyllites, in many places adjacent to massive ores or in tuffitic rocks.

For type (1a), Juve says that to be massive, the ore must contain more than 50 per cent of sulfides by volume. The texture and the composition of this ore is entirely dominated by the contained pyrite more often subhedral than euhedral but is essentially never anhedral unless it has been brecciated. In this ore, sphalerite and chalcopyrite are common and important as matrix minerals. Pyrrhotite ranges widely; galena is quite constant in amount but is definitely minor and normally is not seen by the naked eye. Juve found the mean massivity of 54 samples of massive ore to be 69-45 per cent and the sulfur content of these samples to range between 35 and 41 per cent. The degree to which banding can be recognized depends on the extent to which the gangue minerals are intermixed with the sulfides; the gangue minerals most commonly are chlorite-sericite-bearing keratophyre, but considerable amounts have tuffitic or even graphitic gangue. The most massive ores are best banded and the banding is caused by changes in the proportions of the gangue minerals.

The massive to disseminated, brecciated pyrrhotite-chalcopyrite ore differs appreciably from one occurrence to another, but all kinds have much the same general characteristics. Although the minerals present are those found in all other ore types, the various amounts are quite different. The principal minerals are pyrrhotite and chalcopyrite, are intimately intergrown, and mutually include each other. The pyrite content differs widely, and the gangue is richer in quartz than is true of other ore types. Brecciation is nearly always important, and the grain size is determined by the degree of fragmentation. Fragments of wall rock, gangue, and particularly of pyrite are angular to rounded; the harder minerals have been rolled, bent, and cracked whereas the softer sulfides are much less fractured but normally bear some evidence of deformation. Despite the general breaking of pyrite, in places it may be barely affected; the exterior of such pyrite may be much affected but cracks in the interior are quite rare. The presence of embayments of pyrrhotite and chalcopyrite in pyrite in this ore type suggest replacement by the former two of the latter. How this replacement is effected Juve does not say. It may have been the result of instability created by metamorphic pressure or it may have been caused by hydrothermal replacement reactions whether the solutions involved were metamorphic or hydrothermal in nature. The harder gangue fragments in this ore are, in most places, well rounded, but the softer minerals may be smeared on the surfaces of the rounded ones; such a situation seems much more likely to have been created metamorphically than hydrothermally. This ore type makes up only a small fraction of the total ore volume. The brecciated ore does, however, show gradual transitions to the massive or disseminated types. The brecciated ore is not conformable with the structure of the host rock, showing that the processes, at least as far as the production of fragments was concerned, was tectonically controlled. In this ore type, chalcopyrite is enriched over sphalerite, probably due (Juve, 1974) to the greater mobility of the former. Quartz, however, also is enriched in such ore over the usual gangue minerals, and this hardly can be charged to the greater mobility of quartz as compared with that of the other gangue minerals. Juve does not think that the heavy concentration of pyrrhotite in this ore type is due to its relative mobility but to the metamorphic conversion of pyrite to pyrrhotite. Not enough work has been done to know if the other mobile sulfides (galena and sulfosalts) are similarly enriched in the breccia ore; probably not because of the much greater difficulty of changing pyrite metamorphically to these mobile minerals than of converting pyrite to pyrrhotite.

In disseminated ore in general, chalcopyrite is enriched over the other sulfides; this matter needs explanation. As has been pointed out above, the principal economic ore horizons at Stekenjokk are emplaced at the borders of different rock types, one with another. Generally, these ore horizons are located between quartz keratophyre and black phyllite or tuffite. Actually, Juve's detailed study has shown that each massive layer is made up of a huge number of micro-width horizons that repeat both the composition and the texture of the macro-sized massive layers. For some reason, as Juve says, the individual layers of disseminated ores increase in number and richness as such keratophyre-phyllite contacts are approached. Juve points out that the disseminated ores form a continuous transitional series, the two extremes of which are: (1) anisotropic disseminations which are composed of thin, concordant grains of sulfide aggregates or bands alternating with thicker silicate bands. Obviously, the thicker the sulfide bands become and the nearer they approach each other (i.e. the narrower the silicate bands separating them), the more closely the disseminated ore approaches the massive type until finally the sulfide bands are close enough to each other that the sulfide content of a given vertical section becomes greater than 50 volume per cent and (2) isotopic disseminations which are made up of small, isolated sulfide aggregates that are distributed with equal density in all directions (i.e. they lack all evidence of banding) but nevertheless are generally concordant with their main foot- and hanging wall borders. Most Stekenjokk disseminated ores are of the anisotropic type; only locally are the unusual isotropic disseminations of any economic importance.

Nevertheless, the change of weakly banded (i.e. weakly disseminated) ore to massive ore normally is the result of a sufficient increase in the numbers of bands over a given vertical distance.

The more widely the bands are separated, the greater the frequency of perfect pyrite cubes in a given band; the smaller the crystal the more perfect its outline is likely to be. The relative frequency of pyrrhotite as opposed to pyrite is somewhat lower in pure albite-quartz (keratophyre) rocks than in green tuffite or black phyllite. This suggests to Juve that there the conversion of pyrite to pyrrhotite is greater in the disseminated ore because of the smaller grain size in that ore type which provides a greater surface area over which the necessary reactions can take place. The rhythmic alternations of copper, zinc, and iron sulfides appear to be the same in massive ore as in the lesser numbers of micro bands in the disseminated ore. Pyrrhotite is the only sulfide that is rarer in keratophyre than it is in the black phyllite.

The two types of disseminated ore are those in light-colored rocks and those in black phyllites (mainly adjacent to massive ores) or in tuffitic rocks. In the former type, the behavior of individual crystals or the development of their aggregate geometry is widely varied. In weak disseminations, sphalerite and chalcopyrite aggregates are most randomly distributed in relation to grains of pyrite. All of the softer sulfides are present in aggregates that are flattened parallel to the prevailing schistosity. In denser disseminations, the softer sulfides, in the more deformed portions, are concentrated in pressure shadows around pyrite crystals. As the pyrrhotite content increases in this ore type, the intergrowths of chalcopyrite also increase; sphalerite does not behave in this manner. In the more deformed layers, pyrrhotite-chalcopyrite agregates are increasingly present in discordant fractures and cracks. Juve appears to be certain that this phenomenon is due to metamorphism-induced deformation.

The disseminations in black phyllites of tuffitic rocks have the same composition as those in light rocks; pyrite dominates over chalclpyrite and sphalerite, but pyrrhotite is more important than in the light rocks. At Stekenjokk, there are gradual transitions in both texture and composition, ranging from a monomineralic pyrite nodule type to base-metal iron-sulfide assemblages with intimate intergrowths taking place among all the sulfides. This, Juve thinks, shows that all the sulfides were produced contemporaneously.

In places, the disseminations of aggregates are quite irregular; these consist of flattened aggregates that have no concentration at any particular horizon. This arrangement also is true, in places, of pyrite, except that the pyrite is much more likely to be in well-outlined crystals no matter what the size of the grains may be. Where chalcopyrite and sphalerite are intergrown, they usually are so abundant that pyrite grains seldom are in contact. Juve finds that the more the rock is deformed, the more common are the accumulations of pyrrhotite-chalcopyrite to the exclusion of pyrite. Juve thinks that such arrangements are preliminary to the formation of independent, brecciated chalcopyrite-pyrrhotite ores. He also holds that the generally fine-grained and incompetent chalcopyrite is the result of mobilization and redeposition of that sulfide, in many places as rims along the contacts with massive ores. Along the contacts of massive ore with black phyllite, the margins of that rock may contain as much as 15 volume per cent of chalcopyrite up to a thickness of 20 to 40 cm but normally less. Juve believes that much chalcopyrite has been tectonically moved from the massive ore into the bordering phyllite. Veinlets of chalcopyrite, small and in most places disconnected and with or without other sulfides, do not reach more than a few centimeters to as much as 2 dm in to the phyllite.

What Juve calls meta-pegmatites are present in stress minimum zones in the phyllite and the light-colored rocks as well. Concentrations of rare sulfides and sulfosalts are found in such situations.

Juve presents four rules to explain the locations and intensities of chalcopyrite impregnations marginal to massive banded ore. These are: (1) the more intense the deformation, the broader the impregnated zone with these impregnations being directly related to deformation-induced fracturing of the phyllite; (2) impregnations normally reach farther into [more readily fractured] keratophyre than into the phyllites; (3) the best developed low-pressure zones contain the greatest concentrations of the more mobile sulfides, mainly chalcopyrite; this rule applies principally to hinge zones and drag structures in which [rather unexpectedly] gangue minerals such as quartz and carbonate also are concentrated; and (4) in loci of structural attractiveness for remobilization, remobilization is greater for chalcopyrite than adjacent to less deformed locations - the corollary to this is that the less the deformation the less likely the chalcopyrite is to be remobilized.

Juve believes that metamorphic movement (and to some extent change in composition) produce such changes in mineral geometry and in the character of the minerals themselves as always indicating that they were produced where they originally were deposited at, or a short distance from, those places (measured in millimeters to a few meters) where they originally were emplaced. Pyrite's conversion to pyrrhotite is stronger in disseminated mineralizations than in massive ones, adding to the zoned appearance of the ores. He refers to this mobilization and redeposition as internal and lateral secretion. The sequence of mobility was fahlerz, galena (both minor in amount) and chalcopyrite; pyrite is least affected by these processes. Redeposition is opposite to the mobility is relative effectiveness. He is convinced that the ore textures were produced entirely (or essentially so) by metamorphism; the depositional sequences that appear to exist were brought about by grain growth, selective mobilization, and redeposition induced by metamorphism. The fillings in cracks and fractures caused by the F_2 folding was accomplished by lower-temperature solutions [of what composition and chemical character Juve does not say]. He believes that locally the sulfides, under metamorphic effects, were converted to a highly mobile sulfide mush as indicated by the fine-grained texture exhibited in places along ore-country-rock contacts. These mushes, however, were volumetrically unimportant. Juve thinks, therefore, that metamorphism was at its peak of intensity during the F_1 fold phase. Such retrograde metamorphism as occurred was only in the filling of fissures in zones parallel to the axial planes of the F_2 folds.

He considers that the mutual relations of the sum total of the metal content of the ores show: (1) copper and zinc have no distinct correlation; (2) copper and lead probably have no distinct correlation; (3) zinc and lead show a positive co-variation; (4) silver and lead show a definite positive co-variation; (5) silver and copper show a positive co-variation; and (6) silver and zinc show a weak positive co-variation which probably is not more than the effect of the intimate crystal association of sphalerite and galena.

As is true of most Caledonian stratiform sulfide deposits, the total number of ore minerals at Stekenjokk is quite large (more than 30), but the number of abundant and common minerals is small with pyrite and pyrrhotite, sphalerite and chalcopyrite being the only ones that are quantitatively important. Only a few of the minor minerals have economic importance; sulfosalts and tellurides do carry worthwhile amounts of silver. Pyrite is the really dominant iron sulfide as pyrrhotite (in Juve's opinion) is a metamorphic transformation product as much as it is a primary mineral. Juve believes that the mineral textures demonstrate that all of them were deposited at essentially the same time. What seem, if the metamorphic effects are neglected, to be mineral textures showing apparent paragenetic order, actually are to be seen as the result of the almost complete recrystallization produced by the regional metamorphism. What the observer really is seeing are the result of selective enrichment and impoverishment of sulfides under the control of a standard sequence of relative mobility. What minerals are seen as filling fractures are not late minerals in reaching the Stekenjokk scene but simply are those that, although deposited in the same primary mineral melange, were most readily moved into such open spaces by the forces generated during metamorphism.

The more gangue minerals are present, relative to sulfides (Juve believes), the greater the amount of pyrrhotite generated from pyrite during metamorphism with the grain sizes in disseminated ores being greater, the greater was the silicate content of the primary. For Juve's explanation of why this change should occur, see below. But to produce pyrrhotite from pyrite requires that some of the S_2^{-2} in pyrite be reduced to S^{-2}, and what process as a part of the metamorphism can be held responsible for such reduction Juve does not say.

Juve reported in 1977 the results of his efforts to unfold the Stekenjokk beds. He described the result as showing that the original bed normally was (from footwall to hanging wall) as follows: (1) banded disseminations (pyrite-pyrrhotite-sphalerite-chalcopyrite) with the pyrrhotite almost certainly having formed entirely, or nearly so, during the F_1 stage and contained in quartz-keratophyre rock (uneconomic); (2) irregularly to coarsely banded disseminations that are poor in sulfur but relatively copper-rich (py-po-cp-sp) in quartz-keratophyre rock rich in alteration minerals (chlorite and sericite), medium to good economic grade - alteration due to reactions mainly in F_1 stage; (3) fine-banded, massive pyritic ore, largely Zn-dominated (py-sp-po-cp), host rock normally light-colored quartz-keratophyric tuffite - locally dark as an ore type may continue into overlying black phyllite, best economic type: (4) black phyllite with banded to irregular disseminations (py-po-sl-cp) with which are mixed irregular, mostly copper-rich disseminations, economic to sub-economic type - difficult to distinguish between primary chalcopyrite and that mineral later laterally secreted from massive ore; (5) upper banded disseminations (po-py-sl-cp) in tuffitic rock, uneconomic. Juve lacks independent evidence that this sequence is "right-way-up", but argues that this must be correct because: (a) the unilateral position of the alteration assemblage zone [subject to question - see below], and (2) the metal zoning. He goes on to say, however, that a genetic connection between the unilateral alteration assemblage and the ore is debatable because all minerals were deposited at the same time. Further, there are zones of such alteration without any ore, and large amounts of massive ore lack the K-bearing minerals.

The argument he puts forward for the remobilization of primary sedimentary-exhalative to form the Stekenjokk ores is that, during the folding, the sulfides were selectively mobilized and transferred to, and redeposited in, zones of minimum stress, that is, in the thickened parts of folds (such as drags and hinge zones, along cleavage and foliation planes, and in small cracks and fissures). He judges that the material mobilized was initially formed at the same time as the rocks in which they are now incorporated. Although the ores normally are located with quartz keratophyres on their footwalls, it must be emphasized that these rocks were deposited as volcanic clastic debris that later was cemented into solid rock. The cementation, according to Juve, took place as a result of diagenesis and the later metamorphism engendered during the F_1 stage of folding in particular. These processes also, of course, affected the overlying sediments, usually black phyllites; further, at some stage, the sericite-chlorite alteration of the sulfide-containing rocks was produced, also probably during the F_1 folding.

These metamorphic processes must have found the ores already in place, that is, the ore must have been formed before the F_1 folding. Juve argues, implicitly if not explicity, that, granted this relative time of ore emplacement is correct, it is more probable that the ores were introduced during sedimentation than during an episode of hydrothermal activity that followed lithification but preceded the F_1 stage of folding.

There seems to be no question but that all the metamorphic effects described by Juve occurred after the ore had been emplaced; the structures exhibited by the ores at present cannot have been produced by the F_2 folding, except to make minor modifications through the gentle F_2 folding of the already metamorphosed ores. All this will be considered further below.

Juve (1974, 1977) defines the massive ore as having 50 per cent or more by volume. Fifty per cent sulfides corresponds to a sulfur content of 26 to 29 per cent by weight. The average sulfur content of the 15 million tons of economic ore is 20.1 per cent. This means that well over half of this ore is considered as of the disseminated type. So far as the meta proportions in mineralized rock with various percentages of sulfur, the following were noted: (1) the sterile portion of an ore-bearing section of ore (less than 0.5 per cent sulfur) has essentially the same average metal ratio as the proportions in the entire economic ore; (2) with increasing sulfur content, the Cu/Zn ratio first increases and reaches its maximum in the range of 5 to 10 per cent sulfur (the approximate lower limit of economic ore); (3) from this point on (with rising sulfur content and increasing sulfide density) the Cu/Zn ratio steadily decreases and passes near the ratio for the point of average composition of total economic ore (Cu:Zn:Pb being 1:2:0.2) to reach 1.3:6.0:0.21 for the most massive ores. Thus, the principal characteristics that the various ore and non-ore types have in common are: (1) similar base-metal proportions in uneconomic disseminations and in the average composition for the entire ore body; (2) copper is present in relative abundance in sulfur-poor economic ores; and (3) zinc is relatively abundant in sulfur-rich economic ores. This indicates that, through the entire vertical dimension of sulfide mineralization, the mass of mineralized rock is quite homogeneous; the composition of the ore across the vertical dimension, however, differs widely between the limits set in the average of ore mineralization. Juve, however, believes that the ore in a given stratum has a quite uniform composition along strike. Thus, the economic value of any particular sulfide layer depends on the degree to which it is diluted by silicate minerals.

It should be noted that, in addition to the main ore layer, uneconomic sulfide layers are present throughout the entire stratigraphic sequence; the uneconomic layers also tend to be emplaced at the transitions between depositional phases of the host rocks.

Juve outlines how the transition of pyrite to pyrrhotite (more intense in disseminations than in massive ore) is accomplished. The change from pyrite to pyrrhotite releases sulfur which Juve then combines with iron released from gangue silicates to produce pyrrhotite additional to that formed directly from the pyrite, but remember that the sulfur so released is S_2^{-2} and must somehow be reduced to S^{-2}.

Juve, however, concludes that an elongated sulfide lens was formed syngenetically in a shallow sea. Into that sea the first material in a given rock-ore sequence was airborne silicic clastic material; this was followed by subaqueous sulfide-rich volcanic exhalations that overlapped to some extent with the silicic clastic material and were contemporaneous with, and followed by, the sedimentation of black phyllitic material. Juve believes that the volcanic exhalations that produced the sulfides probably also were responsible for the alteration of the silicate materials in the black phyllite, particularly, to a chlorite/sericite mixture that would be preserved through the two following stages of folding. The ore did not have a chemical or physical preference for the contact of the keratophyre-to-be or the black shale; it was the accident of when volcanic exhalations entered the basin that determined where the sulfide layers would lie in relation to the host rocks. The position of the sulfides was, in each case, a matter of coincidence. This point Juve makes but does not emphasize, and this is the major weakness in his otherwise excellent concept.

It would be most remarkable if volcanic exhalations could be so exactly timed that they always were laid down with their maximum development just as the formation of the stratigraphic sequence changed from the sedimentation of silicic clastic volcanic debris to that of surface-supplied sediments that would lithify to black phyllitic shale. The only way I see out of this impass is to assume that the sulfides were brought in by magma-generated sulfide-rich solutions that entered the lithified rock sequence before the beginning of the F_1 folding. These solutions followed the contacts of the various keratophyre-phyllite contacts, replacing those materials, but doing so more extensively in the black phyllite than in the keratophyre. On the contrary the alteration effects of these solutions were more impressive in the keratophyre than in the phyllite, thus accounting for the greater wall-rock alteration in the footwall than in the hanging wall of the ores.

I agree completely with Juve that the realtionships of sulfides to folds and fractures was due to the tectonic and metamorphic effects of the two phases of folding.

If Juve's concept is correct, the age of the ores is Silurian (middle Paleozoic), the age of the encosing rocks, and the classification is Sedimentary-A-3, plus Metamorphic-C. If my ideas are nearer correct, then the ores also are middle Paleozoic because they were introduced in Silurian rocks before they were affected by the Caledonian orogeny. The Stekenjokk ores, however, should probably be classified as Mesothermal with, perhaps, some early overlap into the hydrothermal range, but neither the mineralogy of the primary ores, so far as can be deduced, nor the type of wall-rock alteration actually justifies the inclusion of "Hypothermal" in the classification. The term "Metamorphic-C" also must be added to account for the present mineralogic and structural characteristics of the Stekenjokk ores.

VARUTRÄSK, VÄSTERBOTTEN

Middle Precambrian	Lithium, Feldspar Cesium, Rubidium	Magmatic-3a Magmatic-4

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* Papers in the series: "Minerals of the Varuträsk pegmatite" run from I to XXXIX and are to be found in Geol. Fören. Stockholm Förh. from 1937 (Bd. 59) through 1962 (Bd. 84), with XXXVII appearing in 1946 and XXXIX in 1962.

____ 1952, The paragenesis of the Varuträsk pegmatite: Geol. Mag., v. 89, p. 49-60

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Notes

The pegmatite mass of Varuträsk is located 22 km southeast of the great (but now abandoned) copper mine of Boliden; it is some 15 km WNW of the major port of Skellefteå on the Gulf of Bothnia. The deposit was found in 1933 by a farmer who, on speculation, blasted an outcrop of pure quartz he had discovered in a forest clearing. After the blast, a fine-grained pink rock was exposed that was identified by O. Bäckström, then Chief Geologist of the Boliden Company, as lepidolite. Further balsting by the Boliden Company confirmed the presence of lepidolite and other lithium-bearing minerals, and mining was begun on a considerable scale, with the major efforts being expended during World War II when lithium was in major demand. Although mining has been suspended, at last report, large tonnages apparently remain in stockpiles and as yet unmined rock.

Production, during the life of the mine, amounted to 400 tons of feldspar, 4400 tons of lithium-bearing minerals (1400 of petalite, 900 of spodumene, 2000 of amblygonite, and 100 of lepidolite), and smaller quantities of rarer minerals, such as the cesium mineral pollucite.

The pegmatite is enclosed in an amphibolite that is, in turn, intercalated in an extensive belt of what is known (at the mine, at least) as the Phyllite (schist) group; the phyllite originally was composed of fine-grained sediments. Amphibolite is quite common in the phyllite and may have been metamorphosed into its present state from mafic volcanics, lavas, or tuffs, or from calcareous or marly sediments; from which material the amphibolite was produced cannot always be determined. Gavelin (Quensel, 1957) believes that the presence of cordierite in the amphibolite indicates that it (the amphibolite) was initially, at least in part, a sediment. Some of the amphibolite, however, definitely appears to have been from mafic volcanics.

Quensel (1957) believes that the pegmatite is related genetically to the Skellefte granite (a fine-grained variety of the Revsund granite that is so common in the general area). The widespread Revsund granite occupies huge areas in northern Sweden where it has intruded Svecofennian rocks. Welin (1970) has determined the age of the palingenic Revsund granite to be

1785 ± 40 m.y. or latest middle Precambrain. The intruded Svecofennian rocks (the Phyllite group) are correlated with similar rocks to the east that once were known as Karelian and considered to be younger than the Svecofennian and even than the Revsund granite. More recent work, however, has shown this to be untrue, and the two rock types [Svecofennian (or Svionian) and Karelian] are considered of essentially the same age, having been laid down between 2100 and 1900 m.y. ago. Herzog (1961) dated the lepidolite from Varuträsk as only slightly younger than Welin's determination for the Revsund rocks. Quensel rounds off the age of the Varuträsk pegmatite as 1800 m.y. which cannot be far from the mark; this age is in agreement with the ages assigned to equivalent formations in middle Sweden.

The surface plan of the Varuträsk pegmatite has the shape of a boomerang, convex to the south. The width across the open (northern) end of the arcuate body is about 350 m, whereas the length of the broader (eastern) limb is about 275 m. The total length of the boomerang is about 500 m; the exposed length when the overburden was removed was about 350 m. Lithium-rich units occupy a considerable portion of the plan area of the pegmatite, perhaps about 20 per cent and about equally distributed between the two arms of the boomerang. The three areas containing cesium-rich rocks are far smaller in total than that of the lithium units, and each of these three lies within an area of lithium-rich rock. On the surface, the eastern arm of the pegmatite dips about 30° WNW, whereas the western segment is essentially horizontal. The thickness of the pegmatite ranges from a few up to 30 m. Grip (1978) says that the single, connected outcrop divides with depth into two sheets. The western sheet dips 10°E and the eastern about 20°NW. These dips appear to be different than those reported by Quensel (1957) for the surface exposures. The widest part of the pegmatite outcrop at the surface is nearly 90 m; it is located near the northeast end of the eastern limb of the boomerang. The narrowest part is near the north end of the western limb, where the pegmatite is no more than 10 m wide.

Quensel divides the crystallization history of the pegmatite into two parts: (1) the pegmatite stage proper and (2) the younger "pneumatogenic" stage by which he means that the minerals produced during this stage were formed by replacement of the first-stage minerals through the action of hydrothermal fluids above their critical temperature. The pneumatagenic fluids were subjected to such high confining pressures that they were dense enough to have carried the ions they contained in true solution. He also recognizes a minor hydratogenic stage and a stage of superficial alteration; these two stages are of essentially no economic importance.

During the pegmatitic stage, four zones of deposition from the melt were developed: (1) the border (outer) zone; the wall zone (next to the border); (3) the intermediate zone (divided into outer and inner portions); and (4) the core. Because the quartz in the core is milky white whereas that in the outer three zones is glassy, Quensel believes that the milky quartz is of the low-temperature variety and the glassy of the high-temperature type. He thinks, therefore, that the temperature of formation of the three outer zones must have been above 600°C and may have been, initially, as high as 800°C. This is probably rather high for a pegmatite, and his evidence for these temperatures is not conclusive (see below), and his estimate may be high by as much as 200°C.

The four zones of true pegmatite formation (border, wall, intermediate, and core) are believed by Quensel to have been formed by fractional crystallization of the pegmatitic fusion from the outside inward with all the reactions taking place in a closed system except for minor reactions with the enclosing walls.

The minerals in the border zone are a fine-grained mixture of quartz and muscovite; this zone is seldom more than 10 cm thick. The wall zone has a thickness of 0.5 m to several meters and is composed of the same minerals as the border; large crystals of single minerals, however, are common. Although microcline perthite is found in the wall zone, most of it probably resulted from replacement by late-stage hydrothermal fluids generated within the intermediate zone of the pegmatite.

The intermediate zone includes all the true pegmatite between the wall zone and the core; it forms by far the major portion of the mass of the four divisions into which the three outer zones are divided (the intermediate zone consists of two parts, the outer and the inner). The outer part of the intermediate zone has a more simple mineralogy than the inner one. Although quartz and muscovite still are present in the outer portion, the most important mineral there is microcline perthite as large (up to 3m) crystals and anhedral masses. This perthite is remarkable for its high rubidium content; in the eastern arm of the boomerang, the microcline perthite averages 1.55 per cent Rb_2O, whereas, in the western limb, the average is 0.72 per cent. The mica minerals of the intermediate zone also include appreciable rubidium. In the outer part, this zone also includes beryl and manganapatite. In the inner intermediate zone, abundant spodumene ($LiAlSi_2O_6$) and amblygonite [variety montebrasite ($LiAlPO_4OH$)], and verdelite (a green tourmaline) are present in addition to the minerals of the outer part of this zone. The two sections of the intermediate zone merge gradually into each other. The huge size of many minerals in this intermediate zone is true of both of its subdivisions. The microcline perthite in this zone is not a product of replacement (as in the wall zone) but appears to have been a result of primary crystallization from the molten phase.

The core of the pegmatite, as far as can be seen on original surfaces exposed by the removal of the drift cover, is only a small portion of its total mass; no more core was exposed by quarrying or underground workings. The only mineral in the core is a pure milky white quartz. This quartz (Quensel, 1957) is thought to be of the low-temperature variety, although it appears from his 1942 paper that the principal evidence for this is that the quartz associated with the late (hydrothermal) cleavelandite (small, platy crystals of albite) is glassy. From these quartz characteristics, Quensel reasons that the earlier (milky) quartz must have been formed as the high-temperature variety at some temperature above 600°C. I am not certain that this difference between the two types of quartz provides a sufficient base for declaring one to be of the high-temperature kind and the other of the low-. If this is a reasonable conclusion, then his upper limit of 800°C on the formation of the minerals of the pegmatite is high by perhaps as much as 200°C or so. I doubt if the milky quartz of the core formed in a higher range than 400°C to 500°C or perhaps even lower. The lack of essentially any quartz crystals in the Varuträsk pegmatite prevents a study of quartz morphology and, thereby, a determination of what variety or varieties of quartz are present.

The deposition of the core completed the formation of what Quensel considers as the pegmatitic development cycle proper; the minerals introduced into the pegmatite body after the consolidation of the intermediate zone almost certainly were the result of replacement reactions by water-rich fluids generated during the crystallization of the third (intermediate) zone and during that of the core. These replacement reactions make up what Quensel designates as his pneumatogenic stage. These replacing fluids were forced out into the pegmatite zones themselves along fractures in the solid portion of the body probably both before and after the consolidation of the core. One of these replacement bodies completely surrounds the core (in the east arm of the boomerang); the remainder of these replacement bodies are rather irregularly distributed through the mass of true pegmatite, but Quensel's maps show no relationship in space between them and any core material. This would seem to confirm the concept that the bulk of the replacing fluids were generated during the crystallization of the third (intermediate) stage in the main rather than solely during core formation. At any event, it seems probable that the water-rich fluids that caused the replacement were developed in the pegmatite fusion after it had become saturated with water (through the earlier crystallization of mainly anhydrous minerals). With further crystallization, water was given off from the fusion in major quantities, the ions contained in which reacted with the already solid pegmatite to produce the minerals of Quensel's pneumatogenic stage.

Quensel divides his pneumatogenic stage into: (1) a high-temperature phase and (2) a lower-temperature phase. He separates this high-temperature phase into (a) a lithium unit and (b) a cesium unit; the lithium unit was formed at higher temperatures than the cesium one. The lower-temperature phase consists only of a sodium unit.

The lithium unit includes some of the usual minerals of the pegmatite - quartz (glassy), muscovite, lithium-bearing micas, and spodumene but lacks microcline perthite. Among the new minerals, the lithium unit contains lepidolite [$K(Li,Al)_{2-3}(AlSi_3O_{10})(O,OH,F)_2$]; both Al^{+3} and Li^{+1} occupy octahedral coordination sites - any formula for lepidolite is a compromise among conflicting data. Also newly present is petalite ($LiAlSi_4O_{10}$), plus minor amounts of verdelite, beryl, amblygonite, manganese apatite, manganvoelkerite (Mn-oxyapatite), triphylite [$Li(Fe,Mn)Po_4$], lithiophylite (an iron-poor analog of triphylite), and varulite (an Na-Ca-Fe-Mn- phosphate). The various bodies of the lithium unit cover as much as 20 per cent of the area of the pegmatite shown in Quensel's (1957) map and indicate that a very considerable volume of lithium-rich, water-rich fluid must have been available for their production.

Quensel makes a definite distinction between lithium-bearing micas and lepidolite with the rock containing them being mauve-colored and fine-grained and concentrated in the western arm of the boomerang. Where the lepidolite is cut and splotched by replacement veins and patches of cleavelandite, small grains of cassiterite also are present and apparently were introduced at the same time as the cleavelandite, both as part of the mineralization developed during the time of the later sodium unit.

Spodumene is the third lithium silicate of importance in the lithium unit; although this mineral is present in the inner intermediate pegmatite zone, it also occurs in the lithium unit but with an obviously different habit - slender laths here instead of the earlier compact tabular masses. While lithium-unit spodumene locally is altered to a mixture of clay minerals, this alteration does not happen to the spodumene in the inner intermediate zone of the pegmatitic phase. Spodumene and petalite are found together in only one place in the lithium unit, where they are closely intergrown with quartz; so they probably were deposited at essentially the same time. Although petalite is in places in contact with pegmatitic spodumene, a narrow rim of recrystallized spodumene divides the two, indicating that the petalite is not part of the pegmatitic stage but is appreciably younger.

Other minerals of the lithium stage occur only in minor amounts. It includes a younger generation of montebrasite; here in much smaller crystals than in the inner intermediate pegmatite zone; it is, however, younger than the cleavelandite and probably should be part of the lithium unit.

The cesium unit consits of three separate bodies large enough to appear on Quensel's 1957 map, and these are each completely surrounded by larger bodies of the lithium unit. Two of the cesium-unit bodies are in the eastern arm and one in the western. Of the four lithium-unit bodies in the eastern arm, only the one surrounding the core contains cesium units (two). The third cesium unit is near the southern end of the long, narrow lithium unit in the western arm. In addition to glassy quartz, the cesium units also contain large masses of pollucite ($CsAlSi_2O_6$), polylithionite (a variety of lepidolite), a rubellite (a lithium-bearing, red tourmaline).

Quensel thinks that the cesium replacement units were deposited through new replacement channels and that the replacing solutions must have been of quite different composition and probably were from a different source than those that formed the lithium units. This does not follow automatically, as the cesium bodies are much smaller than the lithium ones and may have been formed by the last of the solutions that formed the lithium units. Although Quensel says that the cesium-unit solutions must have, to a major extent, replaced the minerals of the intermediate (pegmatite) zones, most of the pollucite (in plan at least) is within lithium-unit replacement rocks. Only in the underground workings on the 37-m level is pollucite shown as replacing intermediate zone rocks that probably had not first been replaced by lithium-unit rocks. Certainly, the quartz core was not attacked by any of the pneumatogenic solutions in existence after the core had begun to crystallize.

In a few places, pollucite of the cesium unit is observed to have replaced petalite, but normally the order of deposition of the two cannot be observed, but the normal order of events almost certainly was to place petalite in the lithium unit and pollucite in the cesium one. In one area, however, microcline perthite is seen to be replaced by spodumene, then by petallite, and then by pollucite. That the pollucite is earlier than the minerals of the sodium unit is shown by large masses of pollucite cut by veins of albite (cleavelandite) that strike in all directions. In the cesium unit, the only lithium mineral present is a light, grayish-white polylithionite in coarse masses in association with pollucite in the western-arm occurrence of the cesium unit.

The minerals of the sodium replacement unit (the lower-temperature stage of Quensel's lower-temperature phase of his pneumatogenic stage) cut and replace those of all the earlier minerals and zones. The principal mineral of this stage is almost pure albite (cleavelandite), and it includes many others that were not found in the earlier stages and zones. By the time that the sodium unit was being deposited, the contents of the ore fluids in the elements that had made possible the unusual suites that had preceded this zone had been depleted so that these minerals did not deposit at all or did so in much lessened amounts. Enough lithium ion, however, was available for this stage to include lepidolitic micas and lepidolite, plus saacharoidal albite (in addition to cleavelandite), verdelite, rubellite, and indicolite, beryl, and minor amounts of several phosphates, and several metallic ore minerals - cassiterite, columbite, tantalite, microlite, the allemontite group, and uraninite - all in quantities to be of scientific interest only.

In the sodium unit, the lepidolite (or lepidolitic muscovite) is in veins cutting the older minerals, including petalite; locally this mica is replaced by cleavelandite. The larger masses of cleavelandite occur as spheroidal sun-bursts or as large, radiating sheaves; the fine-grained albite is confined to the outer reaches of the pegmatite. A second type of lepidolite is in large purple crystals; it has the highest Li_2O content (5.95 per cent) of any such micas from Varuträsk that have been analyzed. This purple lepidolite forms concentric bundles and is associated with a rose-colored, fine-grained muscovite. Another mica mineral in the sodium unit is the oncosine variety of muscovite that cuts large masses of pollucite.

As the water-rich fluids responsible for the pneumatogenic stage became cooler and depleted in most of the elements that were responsible for the minerals of this stage, kaolin-group minerals, montmorillonite, vivianite, and cookeite $[LiAl_4(Si_3Al)O_{10}(OH)_8]$, a member of the chlorite group, were deposited. These constitute Quensel's hydratogenic group and are of no economic importance. Some surficial alteration also took place long after the pegmatite had solidified and after it had been brought near the surface by erosion.

It appears obvious that the huge volume of solutions needed to produce the lithium, cesium, and sodium units could not have come from the crystallization of the core material alone; rather the development of water-rich fluids must have taken place during the same period over which the intermediate zones were being formed, and this production of water-rich fluids must have continued for a very considerable period of time to have accomplished all the hydrothermal changes that took place in the pegmatite.

Quensel seems to think (1957) that the solutions that modified the original pegmatitic material were separated sharply, the one from the next. This seems to me an incorrect assessment of the situation. It seems probable to me that the fluids, once they had begun to separate from the pegmatitic fusion, were supplied continuously from the same source and changed as result both of reaction with the already solidified pegmatic material and of reactions within the molten pegmatitic fusion. The development of the late sodium unit almost certainly was produced by reactions that took sodium out of the wall rocks and redeposited it as veins and masses of cleavelandite. The large number of minerals in the sodium unit resulted from the most uncommon ions finally being assembled in sufficient quantities that they could be precipitated in, for this pegmatite, most unusual minerals.

The Varuträsk pegmatite appears to be a relatively normal zoned pegmatite, modified by self-generated hydrothermal fluids. It is here classed as magmatic-3a for the material deposited from the pegmatitic fusion and as magmatic-4 for the deuteric alteration that produced the lithium, cesium, and sodium units (Quensel's pneumatogenic phases).

AITIK, NORRBOTTEN

Middle Precambrian	Copper, minor Silver	Hypothermal-1 to Mesothermal Metamorphic-C (minor?)

Grip, E., 1978, Aitik, in Bowie, S. H. U., and others, Editors, Mineral Deposits of Northern Europe, Volume 1, Northwest Europe: Inst. Min. and Met. and Mineral. Soc., London, p. 180-182

Zweifel, H., 1972, Geology of the Aitik deposit: 24th Int. Geol. Cong., Sec. 4, p. 463-463

____ 1976, Aitik - geological documentation of a disseminated copper deposit: Sveriges Geol. Undersök., ser. C, nr. 720, 79 p.

____ 1980, Aitik copper ore: 26th Int. Geol. Cong. Guide to Excursions 078 A+C, p. 24-35

Notes

The Aitik copper mine is some 15 km east of Gällivare at 67°07'N, 21°00'E; the deposit is in the same county (Norrbotten) as the famous mines of Gällivare and Kiruna and the newly discovered (and geologically much different) copper deposit at Viscaria near Kiruna. The small (and currently) uneconomic copper deposit of Nautanen is about 15 km NNW of Aitik; other even smaller deposits are known in the Nautanen area. Also, a similar, but smaller deposit of the Aitik type (Liikavaara E zone) is located 5 km east of Aitik, so the mineralization process that produced Aitik is not unique to this one site.

The first indication of the Aitik deposit was the finding of boulders containing disseminated chalcopyrite in 1930 by geologists of the Boliden Company; investigations in 1932 found a small outcrop of this type of ore, and there the Aitik deposit later was developed; at the same time, the Liikavaara zone was localized. The drilling in 1933 and 1936 showed that the Aitik ore was not mineable at that time. After minor investigations between 1948 and 1956, work between 1960 and 1962 demonstrated resistivity surveys to be superior to EM methods for locating the Aitik ore. This work, supplemented by appreciably more drilling, resulted in open-pit production beginning in 1967. Project Aitik No. 1 developed a 2 million-ton-year operation. Project No. 2 raised this tonnage to a 5 million-ton-year venture with an average grade of 0.4 per cent Cu by 1970. The mine now produces about 6.5 to 7.0 million tons a year, again with an average grade of 0.4 per cent Cu. It is expected to take the pit down to 370 m in future years and to mine 11.3 million tons per year when that project is completed. The concentrates produced from this 0.4 per cent Cu ore runs 28 per cent, and the tailings contain only 0.04 per cent Cu. The concentrates are trucked 18 km to the rail station at Gällivare and are carried by rail 400 km to the Boliden Company's smelter at Rönnskar on the coast near Umeå on the Gulf of Bothnia. No other metals are recovered at Aitik except for minor silver.

The area is highly glaciated and is almost entirely covered by major swamps and moraines.

All the sub-soil rocks of the area are Precambrian and are metamorphosed sediments in a forty-kilometer-long zone that averages five kilometers wide; the strike of the zone is parallel to that of the sedimentary rocks. Completely enclosing the sedimentary zone are the Lina granite, migmatites, and gabbro. The igneous rocks are, in part at least, younger than the sediments and have been radiometrically determined to be 1540 ± 90 million years old. Whether the ores are genetically related to the granites or were formed contemporaneously with the sediments, these ores definitely are middle Precambrian in age.

What may be the stratigraphic relationship between the rocks of the Aitik area and the rest of the Norrbotten county remains uncertain. The metasediments, however, probably are younger than the Kiruna porphyry but of about the same age as the Kurravaara conglomerate of that area. Zweifel (1972) considers that the gray leptite at Malmberget well may be metasediments of the same age as those at Aitik. He also holds that the Malmberget red leptites were formed from the gray (?) by the infiltration of microcline into the rocks. He doubts the existence of volcanic rocks in the Malmberget area and, by inference, in the Aitik area also.

The major rock type in the eastern portion of the Aitik district is metaarenite, in the stratigraphically higher parts of which amphibolite layers are intercalated. In addition to the presence of conglomerates in the Aitik metasediments, graded and cross bedding are seen in many places; these eastern rocks have been called the Liikavaara group, although this is hardly a true stratigraphic term. In spite of some recrystallization, metamorphism, and alteration, the metamorphic grade is low. Structurally, the eastern portion of the Aitik area is a syncline, the axis of which dips south-southeast. As both limbs dip steeply east, the structure is slightly overturned to the west. The mineralization at Liikavaara (not described by Zweifel) is on the east side of this syncline.

The effects of metamorphism, recrystallization, and high-temperature hydrothermal alteration are intensive and extensive. This last form of alteration includes: tourmalinization, scapolitization, skarn schlieren, and veinlets. The metamorphic process (in the general sense) largely has obliterated the original sedimentary structures. The western part of this district is considered to be an antiform (if not an anticline) and the structural high is to the east of the mineralized area, but the Aitik ore occurs on the west flank of the antiform.

These western rocks (the Aitik group) are biotite to biotite-amphibole gneisses. These gneisses overlie rocks that are known as the Aitik formation and show low magnetic anomalies in comparison with the overlying gneisses. Principally, the Aitik formation is made up of skarn-banded gnesses and fine-grained biotite gneisses that contain garnets and amphiboles and, in some places, are converted as far as to mica schists or quartzites, gneisses with skarn schlieren, amphibolites, and coarse-grained biotite gneisses. The original composition of these Aitik formation rocks has been further altered by sericitization, scapolitization, tourmalinization, and microcline infiltration. Here also, the metamorphism has so affected the rocks that it is uncertain if the rocks of Aitik formation form a true stratigraphic unit or not.

Still another of these uncertainly named formations, the Nautanen, lies farther east and is petrographically much the same as the Aitik formation but with higher magnetite content. The Lina granite is about 1540 ± 90 million years old which puts it near the upper boundary of the middle Precambrian. It appears definitely probable that the Lina granite was introduced, in part at least, before the ore entered the area. This means that it is older than the Lina granite but certainly cannot be so much older as to be categorized as any age but middle Precambrian.

The actual Aitik ore is in a layer of sericite schists and fine-grained biotite gneisses and quartzites that contain some garnet. These rocks are in the upper part of the Aitik formation and have skarn-banded gneiss in their hanging wall. The zone that contains the mineralization is nearly three kilometers long and is about 400 meters wide. The size of the areas of mineable ore normally decreases downward. The portion of this zone that contain the ore is about two kilometers long and up to 200 meters wide. Above the 150-meter level, the cutoff was 0.4% copper; no other base metal is present in amounts worth considering. Gold runs 0.4 grams per ton and silver four grams per ton. Sulfur averages only about 1.5 per cent which indicates the paucity of iron sulfides in the deposit.

The rock in which the ore and mineralized zones lie is a series of metamorphic rocks that are of a range from sericite schist to biotite gneiss, and gneisses with schlieren of skarn. The major portion of this rock is fine-grained biotite schist, some of which rock contains red garnet. The texture usually is lepidoblastic and the grain sizes are small (1.0 millimeters to 0.1 millimeters). The correlation between one drillhole and the next is difficult because the differences in the various rock types, though small, are too irregular to permit certain identification of marker beds. Only the hanging wall of readily identified skarn banded gneiss is definitely to be recognized. The quartzites grade into mica schists with increases in micas and feldspar. With further increases in feldspar, the schists grade into gneisses. The feldspar normally is oligoclase and, in places, may be altered to sericite; some late microcline may be present. The bedrock is a red aplite with some epidote; it consists principally of a secondary (?) microcline. The rocks of the ore-bearing section are considerably scapolitized and contain disseminated tourmaline; black tourmaline may be abundant enough to produce a tourmaline fels. The skarnschlieren-bearing rock is largely on the footwall, with the dark minerals (hornblende and biotite) surrounded by bleached areas made up of plagioclase and quartz; the dark minerals contain some chalcopyrite and pyrite.

On the west side of the Aitik deposit, pegmatites are common and lie between the granite and the ore; they are, Zweifel says, definitely related to the granite. The abundance of pegmatite is less as the contact is followed from west to east. Further, in the actual mineralized rock, such pegmatites are present; their numbers, widths, and frequence of occurrence being greater with depth. Between 400 and 600 meters down, the pegmatites may be up to 40 meters wide. The bulk of the pegmatites are oriented parallel to the general strike, but some of them cut through the bedrock of the mineralized zone. The pegmatites contain quartz, microcline, and albite and considerable black tourmaline. Zweifel believes that the tourmaline in the bedrock had its source in the pegmatite magmas. The pegmatites normally are not mineralized, but the ore minerals are larger in size the nearer they are to the pegmatites.

The only mineral that is worth recovering from the ore is chalcopyrite (and from the concentrates some silver is recovered) but some magnetite, pyrite, and pyrrhotite are present. The ore is cut by quartz veins that may include bornite and chalcocite; economically these minerals probably are worthless but they are recovered with the chalcopyrite and may add slightly to the copper content. In places, arsenopyrite, sphalerite, and galena have been identified. Silver is recovered from the chalcopyrite concentrates and probably is in inclusions of tetrahedrite in the copper mineral. Very minor ilmenite, with magnetite and hausmannite are in the ore, and uranitite, scheelite and molybdenite have been seen in the pegmatites. Quartz and barite make up the principal gangue minerals, and fluorite and calcite are rarely in small veinlets. Such tourmaline, scapolite, and apatite as are in the ore Zweifel considers to have been derived from the pegmatitic magmas. The commonest skarn mineral is hornblende associated with some biotite. The skarn minerals are in schlieren and veinlets; they include appreciable chalcopyrite; normally these are near the footwall of the ore. Sericite is spatially and probably genetically associated with the ore, especially where the ore carries more than normal pyrite as it does near the hanging wall.

The ore occurs in several varieties: (1) disseminations and stringers in gneisses, schists, and quartzites that are somewhat tourmalinized and scapolitized; this type provides the bulk of the ore; (2) disseminated pyrite and chalcopyrite in schists and quartzites or in partly sericitized biotite gneiss; this variety is near the hanging wall and is not economic; (3) chalcopyrite and pyrite concentrated in skarn schlieren and veinlets with stringers of magnetite and barite; this is a footwall ore and only in a few places is of mineable worth; (4) chalcopyrite and pyrite in quartz veins and veinlets (with some chalcocite and bornite); large veins are uncommon but in only a few places are worth mining; and (5) chalclpyrite in the red aplite-like gneisses; this ore is known in only one place.

The general mineralization contains from 0.15 to 0.70 per cent copper - but several centers of copper concentration are known where the copper grade is more than 0.70 per cent, locally being as high as more than 1.0 per cent. The principal zone of high-grade ore is between the center of mineralization and the hanging wall, but some concentrations of higher-grade material are in the northern and southern parts of the mineralized zone.

The pyrite is more abundant toward the hanging wall; in the north portion of the ore zone, the mineralization plunges north; pyrite is concentrated at the top of the plunge.

The mineralized zone is located on a minor anticline on the west flank of a major antiform that is almost three kilometers long and about 400 meters wide at a maximum. Toward the hanging wall to the west, a sharp contact exists between the ore and the skarn-banded gneiss. On the contrary, to the east, the ore grades gradually down and lacks a sharp contact. The ore and the mineralized zone both dip 45°W and strike N10-20°E. The dip of the limit of economic ore on the footwall usually is somewhat steeper than that of the hanging wall; in fact, the hanging wall, with depth, dips no more than 35°W. The mineable ore plunges 25° to 30° toward N20°E in the north part of the ore body; this plunge is parallel to the observed fold axes. In the center of the ore, the fold axes are almost horizontal, whereas, toward the south, they dip 20° toward S20°W. In the southern part of the ore body, it is not certain if the ore follows the fold axes.

The ore zone at the Aitik mine has a high contant of K_2O, whereas the rest of the district is high in Na_2O. MgO decreases in the ore zone (as compared with the surroundings) but increases again in the skarn schlieren of the footwall. Alkalies increase only to a minor extent toward the hanging wall of the mineralized zone. CaO content drops off appreciably between the skarn-banded rocks of the hanging wall and the mineralization. Although the Fe and Mg contents increase east of the hanging wall, they remain quite steady in the mineralized zone. The SiO_2 content rises slightly to the east of the hanging wall and then goes down toward the skarn schlieren rocks of the footwall. The barium appears to agree with the copper pattern. Of the trace elements, only Ag and Mo are more abundant in the ore than in the surrounding rocks. Only in the ore do Pb, As, Bi, and Sn increase relative to the rocks that enclose the ore zone.

Zweifel believes that the anomalies must be related either to primary sedimentation or to metamorphic-metasomatic processes. He does think that the low CaO content of the ore zone could have resulted from a primary difference in the sediments deposited in the ore area compared with that of those outside it. The high K_2O content definitely, he says, is related to the ore zones (and probably to the processes that produced them). The concentration of Fe and Mg in veinlets and schlieren, Zweifel says (obviously), must be related to metamorphic or pneumatolytic processes with potassium entering the ore zone and Mg and Fe being moved from west to east. He does believe, however, that tourmalinization and scapolitization indicate that penumatolytic processes were active in the mineralized zone.

In 1917, Geijer suggested that the mineralization at Nautanen, which is similar to Aitik and is located about 13 kilometers north-northwest of it, was formed by hydrothermal processes that took place under high-temperature (hypothermal) conditions and was related to the intrusion of the Lina granite. This granite surrounds, and is younger than, the metasediments that contain the Aitik ores. If the Aitik ores were formed syngenetically with the sediments that now contain them and developed their present characteristics through metamorphism, their genesis cannot be related to any activity of the Lina granite. If the ores, on the contrary, are epigenetic in the sediments, they may yet be old enough that they may have been formed by fluids given off by that granite before the pegmatites associated in space and genesis with the Lina granite were introduced. To some extent it seems certain that the ores were affected by the metamorphic events that took place in the Aitik area, but it is by no means as certain that the ores were not introduced at an appreciable time after the sediments were first laid down.

Zweifel (1972) considers that the ores are pre-pegmatite because the grain size of the ore minerals is greater the nearer the ores are to these magmatic bodies and to concentrations of tourmaline. From this he concludes that the ore minerals were introduced into their present surroundings by metamorphic remobilization. He does, however, state that the primary introduction of the copper must have been either: (1) by original sedimentary concentration or (2) by hydrothermal activity related to the earlier phases of the Lina granite. He adds, however, that a "later remobilization relation to the intrusion of the pegmatites and tourmalinization and scapolitization must be considered". He concludes that the latter model must be correct because: (1) the mineralization of the Aitik-Nautanen zone is essentially confined to a sepcific geological horizon and (2) the Lina granite resulted from granitization which covers large areas, and does not seem to have produced hydrothermal solutions anywhere else in Norrbotten.

Zweifel, therefore, puts farward the following scenario for the formation of the Aitik ores: (1) copper was reconcentrated (from sea water?) into the rocks of the Aitik area, particularly those that are now part of the Aitik ore formation (he does point out, however, that the mineralization at Liikavaara (3.5 kilometers east of Aitik) is in a different stratigraphic horizon although the two mineralizations are much the same); (2) folding and the first stages of granitization that developed the Lina granite, plus regional metamorphism and possibly the first remobilization of the sulfides; (3) sericitization and (further) mobilization of the sulfides in connection with early pneumatolytic activity and the removal of Mg, Fe, Ca, and Na from the sericite zone (in the eastern part of the present mineralized zone); (4) intrusion of pegmatites from the molten (portion of?) the Lina granite plus tourmalinization, scapolitization, and mobilization of mafic minerals and sulfides; (5) introduction of skarn schlieren and veinlets, both of which contain sulfides.

Although this scheme of genesis seems to fit most of the known facts, it seems to me not to be the only one that could be suggested. Let the ores have been introduced during the middle stages of the Lina granite formation, whether it had been developed by granitization or by synkinematic melting, during either of which processes water-rich fluids might have separated from the magma mush or melt. These fluids, having entered the region of the present mineralization would have deposited essentially the minerals that the ores now contain. Later during the second stage of folding (which saw also the intrusion of the pegmatites and the tourmalinization and scapolitization) some modification of the relations of the sulfides to their host rocks might have occurred, such as the increase in ore grain size and perhaps, though less probably, the development of quartz-sulfide veinlets, and the possible introduction of additional tourmaline and scapolite.

If the second model for the formation of the ores is acceptable, then they must be classified as Hypothermal-1 to Mesothermal on the assumption that the deposition of chalcopyrite, so far as it overlapped sericite formation, probably took place over both temperature and pressure ranges. Some effect of metamorphism should be noted in the classification (Metamorphic-C). If Zweifel's mode 1 is correct, the ores should be categorized as Sedimentary-A3, plus Metamorphic-C.

LAISVALL, NORRBOTTEN

Middle Paleozoic Lead, Zinc Telethermal

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Notes

The Laisvall mine (66°08'N, 17°10E) is on the border between Swedish Lapland and Norway. Although ores of the Laisvall type are common in the Caledonian Mountains, Laisvall is the only one that is even moderately well represented in the literature. Even Laisvall was only sparsely described until the publication of the detailed paper by Rickard and his colleagues (1979).

The Laisvall deposit is contained in a zone of late Precambrian and Cambro-Ordovician sediments that form the eastern border of the Caledonian Mountain chain over a length of some 2000 kilometers. These sedimentary rocks lie unconformably on the Precambrian basement and are overlain by nappes of Caledonian age thrust eastward from Norway. These ores generally are confined to psammitic sediments that are located both under and above the nappes. At Laisvall, the ores are in autochthonous sediments in the nappe belt, but the ores are not confined to the thrust front and occur well within the Caledonian chain. Some veins containing the same minerals as at Laisvall also are known in the area; they may penetrate the basement rocks and have been mined for hundreds of years for their silver content. Its annual production of four million tons of ore makes Laisvall one of the largest lead-zinc mines in Europe.

The ore-containing rocks of the Laisvall area lie unconformably on the weathered surface of the Precambrian basement and are combined into the Laisvall group. This group, from bottom to top, consists of the Ackerslet formation, one to nine meters of polymict conglomerate; plus seven to nine meters of arkose (this arkose is irregularly mineralized near Maiva, six kilometers east of Laisvall); then follows the Sävvovare formation, the base of which is shale with randomly distributed pebbles (its thickness is not stated); conformably above this shale is a 40-meter-thick quartzitic sandstone; the sandstones are divided into lower, middle, and upper segments, of which the lower (about 25 meters thick) is a well-sorted, medium-grained rock that includes thin shale bands and that contains one of the two principal Laisvall ore layers, the middle segment is similar to the lower but has a clay matrix, is about seven meters thick, and is poorly mineralized; the upper segment is a coarse-grained sandstone with well-sorted layers intercalated with fine-grained layers; the upper sandstone is up to 11 meters thick and contains the other (and larger of the two) Laisvall ore layers. To the south and east of Laisvall, the lower and middle segments grade laterally into shales and siltstones. Above the upper sandstone is the Grammajukku formation that is some 45 meters of siltstone with shale and sandstone; a bed of limestone is near the top of the Grammajukku. The youngest formation in the mine area is an alum shale. The age of the rocks of the Laisvall group ranges from lower to middle Cambrian. The conglomerate at the base of the Laisvall group, however, may be a tillite from the Varangian glaciation; perhaps all the rocks through to the top of the Sävvovare formation may be Precambrian.

The basement of the ore-bearing autochthonous sediments of the Laisvall area dips 3°W. During the Caledonian orogeny, these sediments were overthrust by a series of complex nappes. These Rickard and his colleagues divide into a lower parautochthonous complex, a middle Kaskejaur complex, and an upper Yraf complex. The first of these nappes (the parautochthonous complex) is made up of a number of slices of autochthonous sediments that have been shifted only a short distance from the west. The Kaskejaur complex is tectonically mixed with the parautochthonous complex, but it is composed of very different rock types. The Kaskejaur complex included principally cataclastic to mylonitized basement and sediments that were moved in from a bit farther west than the nappes of the lowest (parautochthonous) nappe complex.

These two lower nappe complexes are largely parautochthonous rocks, but the Yraf complex above them is part of the huge Seve-Köli nappe complex that is one of the three major complexes in the Caledonides. The Yraf complex is composed of mica quartzites, mica schists, gneisses, and amphibolites. The basal portion of the complex is strongly sheared sericite-chlorite schists. These rocks probably came from no nearer than the west coast of Norway and probably off that coast. Laisvall is about at the eastern most point of advance of the Yraf nappe. The overthrusting of the nappes apparently took place in the order in which they have been discussed here.

The effect of these large-scale overthrusts on the sediments contained in the nappes has been (Rickard and colleagues, 1979) surprisingly mild. The most evident result was the strong shearing and disruption of the alum shale, a formation above the ore zone. The rest of the sequence is essentially unmetamorphosed.

The ores have been introduced into this overthrust sequence at some time before the arrival of the two latest Caledonian nappes, the rocks of which are not mineralized, and after the lithification of the sediments of the Laisvall group. For all practical purposes, the mineralization is confined to the parautochthonous nappe (although a little extends down as veins into the basement). Thus, the mineralization almost certainly is older than the introduction of the Kaskejaure nappe into the Laisvall area.

It is, therefore, possible that the ore was introduced as late as after the movement of the parautochthonous nappe into its present position. The ores, however, appear to have been introduced before the faulting that was earlier than the overthrusting. On the contrary, joints in the ore probably were opened during the earliest stages of the Caledonian revolution, and these joints are older than the ore. All this evidence justifies only tentative conclusions, but the best guess that can be made now places the mineralization in the late Silurian and, therefore, as middle Paleozoic.

At Laisvall, the ore mineralization is enclosed in two very thin sheets, one in the lower and one in the upper quartzitic Sandstone; these sandstones dip gently to the west-northwest. The long axis of the thin ore body in the lower Sandstone is oriented N40°E; the major axis of the ore body in the upper Sandstone is elongated in the same direction. In the upper ore layer, there is a minor elongation at essentially right angles to the major ones. The lower ore body is 5000 meters long and 200 to 600 meters wide, it is two to 24 meters thick. The dimensions of the upper Sandstone are 2000 to 1000 meters long, and 300 to two meters wide; the thickness of this layer is from 2 to 8 m. About 80 million tons of four per cent lead-zinc combined ore were originally present; the ratio of Pb:Zn is 8:1. Although the ore can be considered to be both strata-bound and stratiform, Rickard and his colleagues point out that strata-controlled is a much better term, as the ore was emplaced long after the sedimentary structures were solidly determined. Each deposit consists of very thin layers of ore-bearing sandstone in which the lead-zinc contents range within wide limits. The variation of ore grade in the horizontal dimension is far less than that in the vertical. The Pb-Zn content in sandstone layers may be quite constant over horizontal distances of several hundred meters.

The ore certainly is not confined to a specific layer or series of layers, the ore rising through the Sävvovare formation from southeast to northwest. The middle Sandstone, though normally unmineralized, will contain minor mineralization where ore-grade ore in the lower and upper sandstones is in contact with the middle segment. A northwest-southeast traverse shows that the ore drops 100 meters in 2000 meters whereas the base of the formation drops at a steady three degrees.

Weak mineralization (less than 0.2% Zn and 0.5% Pb) can be found widespread outside the ore volumes and is present as far beneath the ore volumes as into the basal arkose of the Akkerselet formation. The ore is stopped to the northeast by a hill of basement rock; in the northwest, the ore in the upper Sandstone is cut off by overlying siltstones and shales. Laterally, the contacts are gradational, with two per cent Pb-Zn requiring up to 200 meters to drop to practically zero. To the southeast, however, sharp lateral contacts do exist where ore with 10 to 20 per cent lead is bounded by essentially unmineralized sandstone. Where ore crosscuts the beds, contacts are quite sharp, a fact (Rickard, and others, 1979) that emphasizes the incongruent aspects of the deposit.

The ore has been little affected by tectonic movements. The weakly wavy ore beds are concordant with the structure of the lower Laisvall group sediments and with the basement. The wave lengths are up to 1000 meters and the amplitudes less than 10 meters; the fold axes trend generally north-south. A small thrust fault in the central part of the ore body produced folds with wave lengths of 100 meters and amplitudes of 10 meters with no change in fold-axis directions. In addition to this small thrust, the deposit has been cut by several steeply dipping faults and smaller thrusts. The Nadok and Kautsky disturbances are two major fault zones through the ore body. The latter forms the southwest boundary of the ore; it is a steeply dipping reverse fault zone with a 30-meter displacement - it was later cut by thrust faults of low dip; these successively displace the vertical fault to the northeast. Rickard and his colleagues indicate that the spatial relationships of faults and ore show that the ore zone was mineralized before the thrusting took place, and thus that the mineralization was present in the ore layers before the Caledonian orogeny began.

The Nadok disturbance is a small, nearly vertical northeast-southwest normal fault with a displacement of 20 meters. Its displacement of the upper Sandstone mineralization suggests that the ore was introduced before the faulting; this conclusion is confirmed by the presence of mineralized sandstone fragments in the fault zone.

Two major joint sets are parallel and transverse to, respectively, the trend of the Caledonides. The former (northeast-striking) set is mineralized where it members cut the sandstones, the latter (southeast-striking) contains only calcite as open-space fillings. This indicates that the northeast set is older than the southeast set, but that both were involved in the mineralization process or were filled by minerals remobilized later. I incline to the former opinion but Rickard and his colleagues say that "the spatial relationships do not exclude either interpretation".

Where the sedimentary structures are prominent in the ores, the ore follows them; as the structures become less apparent and the sandstones more massive, the ore, galena particularly, occurs in sub-spherical spots. These spots range in diameter from one to several centimeters. Randomly arranged spots grade into linear spot groups as the bedding planes become more obvious; these linear spots then grade into laminae, and then into elongated patches. The random spots also grade into irregular patches so that, if the mineralization process continues long enough, the final result is the same in stratified or massive sandstone. Cross-bedding is followed by the ore minerals as readily as the more horizontal stratification. Small cracks become small veinlets where they are filled with ore minerals.

The ore at Laisvall averages 4.3 to 4.7 per cent lead and 0.6 to 0.7 per cent zinc, but considerable volumes of 10 or even 20 per cent lead ore have been mined; silver is quite low, too low apparently to be recovered, and averages about 160 ppm. The minor amount of sphalerite contains 0.3 per cent cadmium. The iron content of the sphalerite ranges from zero to 3.4 wt. per cent. The ore-bearing rock averages 80 vol. per cent of quartz, 9.0 vol. per cent of phyllosilicates, and 0.9 vol. per cent of detrital feldspar. The remaining 10 vol. per cent is nearly half galena; the rest is either sulfides (sphalerite and pyrite) or non-metallic gangue. A considerable portion of the quartz is quartz cement that filled the bulk of the pore spaces, leaving no more than 10 per cent to be filled by the other introduced minerals. Some small part of the galena actually is replaced by quartz cement. The importance of these relationships will be considered under the section on the genesis of the ores.

Rickard and his colleagues (1979) doubt if a simple paragenetic sequence can be established for the ore (and gangue minerals). Galena and sphalerite seldom occur together, being separate (even where closely associated in space) even under macroscopic examination. Normally, galena and quartz cement are surrounded by calcite and/or fluorite, which relationship appears to establish the relative time of deposition of the former pair to the latter. But this arrangement does not hold for all cases. These authors consider that sphalerite generally is older than galena. The pyrite (bravoitic in places) is very early. Late-stage replacement of the sulfides shows, at least, that the latest solutions could dissolve the two major sulfides.

Lead occurs in about the same percentages in both upper and lower Sandstone, but zinc principally is in the upper Sandstone. Where the two sulfides are present in the same general rock volume, zinc is to the northwest of the lead. Zinc content in the lower Sandstone is about 0.1 per cent but is as much as 1.3 per cent in the upper Sandstone volumes that are low in lead. F and Ba are concentrated in the lower part of the lower Sandstone, whereas Ca (in calcite) is most abundant in the upper part of the upper Sandstone. The distribution of these three elements is not related apparently to that of Pb and Zn; the association of these two groups of elements, therefore, depends on different factors for one and the other; it is essentially a geologic coincidence that they are together.

In an effort to determine at what stage in the geologic history of the deposit its present low porosity was attained, Rickard and his colleagues produced estimates as to the loading pressure on the Laisvall rock sequence. They concluded that the maximum pressure was developed directly after the Caledonian overthrusting, this developing a loading pressure of more than 1000 meters of rock. Before this loading, the depth of burial of the Laisvall sequence was about 100 meters. Thus, the porosity of the original sand sediment was reduced to about 30 per cent. As the ores probably were introduced before the advent of the nappes into the area, much of this 30 per cent porosity was destroyed by the deposition of quartz cement and the ore and ore-associated minerals before the nappes arrived. The quartz cement did not deposit, apparently, as overgrowths on the original quartz grains but as easily distinguished cementing materials. Probably, most, if not all, of the cement was deposited as a preliminary to the ore-mineralizing process, leaving the sandstones that were mineralized later with a considerably reduced porosity from the 30 per cent left after the pre-nappe compaction. How much of the reduction from the initial 30 per cent porosity was due to quartz cementation and subsequent ore mineralization and how much to the pressure of the overlying nappes (after their arrival in the area) is not easily estimated, but the authors apparently think that the pre-thrusting mechanical compaction reduced the porosity to about 30 per cent with the present-day extremely low porosity resulting from the somewhat more than 1000 m of rock that overlay the ore layers after the Caledonian thrusting was finished.

Another attempt to extimate the degree of metamorphism of the Laisvall sequence was made by the Weaver method of illite crystallinity measurement. The crystallinity of the illite is determined by two factors, the source of the sediments and the physical-chemical conditions that prevailed during burial. The Laisvall illites do not show any variation in source, and the differences in crystallinity are, therefore, location-dependent. Within the thin sequence, hydrostatic and lithostatic pressures were much the same from one place to another, and physico-chemical conditions, so far as this can be judged, were essentially equal throughout. This means that temperature was the only remaining factor that could have determined the crystallinity of most of the Laisvall illites. Since Rickard and his colleagues say that, because nothing is known of the kinetics and mechanism of the illite-muscovite transformation, further comments on this possible temperature control cannot be justified. It can, however, probably be said that the metamorphic temperature at Laisvall was less than the temperature at which the ores were formed.

From this it follows that primary fluid-inclusion studies should give a strong indication of true deposition temperatures. Primary two-phase inclusions in upper-Sandstone sphalerite show modal homogenization temperatures of 150° to 155°C; as yet unpublished work shows essentially the same results for lower Sandstone sphalerite. The ore fluids were highly concentrated brine (23 to 26 equv. wt. per cent NaCl), and this suggests that the ore fluids must have boiled before they reached the area of Laisvall ore deposition. The low freezing-point temperature (less than -50°C) suggest to Lindblom and Rickard (1978) that the calcium concentration was "significant". The presence of inclusions showing similar homogenization temperatures in joint and in sandstone pores indicates that the ores in joints were products of the same mineralizing solutions at the same times of deposition. Thus, the joint mineralization could not have been caused by remobilization.

Measurements of delta^{13}C values of ore calcites and part of the sedimentary calcites in the Laisvall ore volumes average about -12°/$_{oo}$ (per mil), whereas the other sedimentary calcites average about -1°/$_{oo}$. Rickard and his colleagues argue from these data that the original carbonates from the Laisvall sandstones reacted with a solution that reached these sandstones at a later time and achieved equilibrium in ^{13}C between themselves and a portion of the sedimentary calcites. Work with ^{18}O ratios showed that almost complete equilibrium was attained between the ^{18}O of the solutions involved and that of the sedimentary carbonate. In addition, of course, ore carbonate was precipitated. Since the ore calcites have delta^{13}C values the same (essentially) as those of the affected sedimentary calcites, the solution effecting this re-equlibration must have been the ore solution itself. Similarly, the agreement in delta^{18}O values between ore calcites and re-equilibrated sedimentary calcites also seems to demonstrate that the solutions that flushed through the Laisvall sequence were the ore fluids and not a later, low-temperature groundwater. Although the data on the hydrothermal and re-equilibrated calcites differ somewhat from one example to another, considering the hydrologic complexity of the system, the data appear, to Rickard and his colleagues, to be remarkably consistent.

Although the Laisvall deposit is unusual in many respects, the strangest aspect of the deposit is that the delta^{34}S of the sulfur in the galena has a mean value of +23.4 per mil whereas that of the sulfur in the barite has a mean delta^{34}S value of +14.9 per mil. If the sulfide ions in the galena were produced by the reduction of sulfur in sulfate ions in the ore fluid (sulfate ions essentially the same in delta^{34}S values as those that slightly later entered into the formation of barite), then the sulfide ions in galena would have to have a less positive delta^{34}S value than that of the sulfate ions from which they were derived. This follows because, in the reduction of S^{+6} sulfur ^{32}S is preferentially reduced over ^{34}S. This means that, in any sulfide containing sulfide ions so produced, the ratio ^{32}S/^{34}S would be larger than the ratio in the sulfate ion from which it came. This further means that the delta^{34}S of the sulfide so formed would be less positive than that of the sulfate sulfur from which it came. This conclusion is definitely stated by Rickard and his colleagues when they say that "the only conclusion to be reached is that the Laisvall sulfide was not derived from the barite-sulfate". They offer two possible explanations for this condition: (1) that sulfide precipitation took place in a single fluid in which there was no sulfide-sulfate equilibrium and (2) that the metal sulfides and the barite were deposited from different solutions.

The mean delta^{34}S value of the sulfur in the barite (+14.9 per mil) is far from that that would be expected in barite deposited from a mid-Paleozoic ocean in which the delta^{34}S of the sulfate sulfur therein contained was between +20 per mil and +30 per mil. The barite sulfur, therefore, must have come from some other source than the ocean in the Laisvall area in pre-Devonian times.

A second suggestion as to the source of the sulfur in the barite (Rickard, and others, 1979) is that it came from the sulfur in the pyrite in the Alum Shale above the Grammajukku formation. The delta^{34}S of the sulfur in this pyrite averages about +15.0 per mil. If this S^{-2} is oxidized to S^{+6}, however, I cannot agree with the statement by these authors that the "low-temperature oxidation of pyrite does not result in significant sulfur isotope fractionation". It is my understanding that the lower the temperature at which fractionation takes place, the greater will be the degree of fractionation. Rickard and his colleagues do not cite a reference for the correctness of the statement just quoted, whereas my concept is derived from the work of Sakai, 1968, Geochem. Jour., v. 2, p. 24-29 and Ohmoto, 1972, Econ. Geol., v. 67, p. 551-578. These authors make clear that minerals formed at low temperature (and almost certainly lower pH than obtained at higher temperatures) are enriched in ^{32}S over those produced at higher temperatures.

Bonds formed by a light isotope in comparison with its heavier counterpart are more readily broken than those of the heavier one as long as an oxidation-reduction reaction is not involved. Once the forces that that type of reaction are created in the system, the effect on ^{34}S ions is greater than that provided by the difference in bond strengths between ^{32}S and ^{34}S. But with or without such oxidation reduction, the effects are greater the lower the temperature at which the reaction in question takes place. Thus, whereas Rickard and his colleagues consider that no appreciable change in the delta^{34}S of the pyrite in the Alum shales would result if that sulfur were oxidized to the S^{+6} of sulfate, I think the oxidation with make the delta^{34}S of this sulfur considerably more positive than that of the source pyrite. If this is true, then the essential identity between the delta^{34}S of the pyrite in the Alum Shale and that of the Laisvall barite is due to coincidence and not to any direct genetic relationship.

To me, the high salinity of the fluid inclusions indicates that the ore fluid during its travels prior to deposition the Laisvall ore underwent an appreciable period of boiling. This boiling was not continued long enough to cause any more than minor deposition of sulfides, as there is no evidence in the Laisvall ores of sudden dumping of major amounts of sulfides. Neither, however, should the degree of boiling that occurred have done anything to produce the peculiar relationships of delta^{34}S values in the sulfides and in barite. No reactions of which I can conceive would have made less positive the delta^{34}S of sulfate ions in the ore fluid while not having had any effect (or a much lesser effect) on the delta^{34}S of the sulfide ions.

To summarize, therefore, Rickard and his colleagues present no acceptable explanation for the delta^{34}S relationships between barite in the ore and the slightly earlier sulfides, except for the concept that a second solution deposited the barite after the original sulfide-depositing fluid had passed out of the system. Where the second solution may have originated or how it was developed remains a mystery unless it is possible that the oxidation of Alum Shale pyrite did not change the delta^{34}S from that obtaining in the pyrite. I still find this suggestion difficult to accept. Rickard and his colleagues favor the introduction into the Laisvall system of two solutions, the first an essentially homogeneous sulfide solution and the second a metal-bearing sulfate solution that was quite heterogeneous. This second solution they envision as having been initially zinc-rich and later as lead-rich, with the zinc-rich portion traveling farther and stratigraphically higher than that containing lead before it began to deposit its sulfide load. Despite the higher position, the ZnS appears to have deposited almost entirely before the PbS began to do so. These authors believe that the H_2S solution occupied the pores in the formation (with H_2S having a partial pressure of about 14 atmospheres) and that it was gradually displaced by, and mixed with, the metal-rich solution; this mixing caused the sulfide precipitation. Despite their reference to work by Hitchon and Friedman to support their belief that the first solution may have been H_2S-rich, it appears less than likely that the mechanism proposed would have been effective. To have the metal-rich solution pass through the now ore-bearing formation in such a manner as to use up all the H_2S this first solution contained would be most difficult to accomplish. Almost certainly a considerable fraction of the H_2S-rich solution would have been driven before the entering metal-rich solution in such a way as to prevent the H_2S in the first solution from being able to react with the metals of the second sufficiently to use up more than a minor fraction of the available H_2S. Despite the point made by Rickard and his colleagues, originally made by Haas, that molecular diffusion becomes more important if there is a significant density difference between the two interacting solutions, did such a density difference exist and were there actually two different solutions? Since all the density information available is that provided by the fluid in the inclusions, it is impossible to tell if there was the significant density difference between the two fluids postulated by Rickard and his colleagues. Even if there were such a difference, the flushing action of the metal-rich solution entering the rock volumes that now contain the Laisvall ores undoubtedly would prevent diffusion for playing a significant role in determining the concentration of the ore-forming fluid.

Although the amount of salt in the ore fluid indicates that the total ore fluid must have contained 10^9 to 10^{10} tons of salt, this does not say where the salt came from. It does say, however, that it must have obtained from some location available to the solution, either where it was generated or on its way to the site of deposition.

For a consideration of the chemical thermodynamic constraints on the ore-forming system, Rickard and his colleagues assumed a three-molar concentration of NaCl and a deposition temperature of about 150°C. They estimated solution pH from the stability of calcite and the P_{CO_2} and a_{F^-} and decided that the lower limit of the pH was about five and the upper limit six (or even as much as eight). Their final decision is that the pH ranged around the pH of neutrality at 150°C (5.8) and probably was not more than one unit of pH from this 5.8 at any time. They emphasize that this was the pH of the ore-forming environment and not of the ore-forming solution. With the available data, and a belief in the two-fluid concept, it is possible only to assign pH limits to the solution produced by the interaction of the two postulated to have existed in the ore-containing volumes. The authors quote Helgeson's argument that, in a solution in which the sulfide was in equilibrium with lead, the pH would have had to have been less than 4.1. This does not, apparently, allow for the possibility that the lead and sulfur were combined in a single complex.

In most of the Laisvall deposit, sphalerite has been deposited at higher stratigraphic elevations than galena, and galena at lower. In the large majority of pore spaces containing sphalerite, sphalerite is the only mineral they contain. The same relationship appears to obtain for galena though not as commonly or obviously as it does for sphalerite. The authors believe that the sphalerite was deposited first of the sulfides and that galena is largely later. They consider that such a relationship can have been produced only by a series of homogeneous ore fluids, the first of which was cabable of depositing essentially only sphalerite but did so at higher stratigraphic levels than the second that mainly deposited galena at lower stratigraphic levels. Still later ore fluids that could deposit only calcite, fluorite, and barite were introduced. To some extent, this arrangement did not always hold for some galena is found being replaced by sphalerite and sphalerite by galena, and galena appears to replace all of the three gangue minerals in places. To explain these relationships, the authors postulate that the first (and homogeneous) solution deposited sphalerite, the second galena, and the third the gangue minerals. Obviously for the various departures from this sequence to have taken place, there must have been some local reversals of the order in which these individual fluids were introduced.

It seems to me, however, that the situations that obtain can be explained by a continuous supply of heterogeneous ore-fluid. The first ore fluid to enter the ore volume deposited galena in the lower reaches of the ore zone. After the galena-depleted solution had reached the upper sandstone, it began to deposit sphalerite. Locally, and late in the cycle,, the temperature of the ore fluid may have been lowered sufficiently to precipitate late galena. Unfortunately, fluid inclusion studies cannot be made on these (or any) galenas, so it is impossible to say at what temperature, relative to sphalerite, this late galena in the upper Sandstone was deposited. Finally, the temperatures required for the deposition of the three gangue minerals were reached and they were precipitated. Local variations in this paragenetic sequence, perhaps several times repeated, may have resulted from differences in solution composition and temperatures, situations highly possible in an ore volume as large as that at Laisvall.

If the situation outlined in the preceding paragraph actually obtained at Laisvall during the time of ore deposition, the need for two solutions postulated by Rickard and his colleagues probably disappears. In the scenario just given, it would appear much more likely that the metal ions were brought into a water-saturated rock volume that contributed little if anything to the ore-forming process. The metal ions probably were bound to sulfur ions, precipitation of galena from such lead-sulfur or lead-sulfur-hydrogen complexes resulted from the stripping of excess sulfurs from such complexes reaction with hydrogen ions with that of galena and sulfur being brought to neutrality (and precipitatibility) at an earlier time (and a lower stratigraphic position) than sphalerite. What part, then, did the high concentration of salt, mainly NaCl, play in the formation of the Laisvall ores? Probably no more than the part played by passengers in an aircraft in making the aircraft fly.

The Laisvall sphalerites contain a wide variety of organic compounds in amounts up to 100 ppm. The authors suggested that these organic materials might be syngenetic with the sphalerite although the presence of squalene indicated modern contamination. Although all the amino acids were introduced by contamination, the much larger portion of the organic material, the n-alkanes and bituments are original in the Laisvall sphalerites. They think that these compounds were not directly derived from biologic materials but through secondary reactions involving original organic compounds.

The Laisvall minerals appear to have been deposited at temperatures of around 150°C. The depth of burial of the Laisvall rocks was not great enough by far to have raised the temperature of water-rich fluids to this temperature. Certainly, the thermal effect of overthrusting cannot have been responsible for so heating the ore fluids, since they had done their work before the nappes entered the area, if not for other reasons. Nor did the Laisvall nappes cause much reduction in porosity through the increase in quartz and carbonate cement, although certainly some cement was formed through the effects of the pressure of burial. At any event, they believe that the effects of pore-closing processes on the Laisvall rocks was to guarantee that the main direction of fluid flow in these rocks would be horizontal.

Where, then, did the Laisvall ore-forming fluids come from? Probably from a magma produced by the partial or nearly complete melting of a portion of the earth's crust that contained sufficient water, lead, zinc, sulfide, fluoride, sulfate, and carbonate ions to produce the minerals of the Laisvall ore when they reached a rock volume where conditions permitted them to precipitate. Of course, the ore-forming ions may have been scavenged from the rocks through which the water-rich fluid passed, but, if so, where did the heat come from to bring the ore fluids to temperatures above the 150°C that they had at the time of ore deposition? At any event, ores deposited from solutions at temperatures of about 150°C and with the simple mineral suite present at Laisvall must be classed as telethermal no matter how they acquired their burden of ore and gangue materials.

KIRUNA-GÄLLIVARE, NORRBOTTEN

Middle Precambrian	Iron as Magnetite, Hematite Apatite	Magmatic-3b

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Notes

The huge iron mine of Kiirunavaara (vaara = mountain) almost certainly is the largest single deposit of iron ore found anywhere in the earth's crust. [In this discussion Kiirunavaara will be referred to as Kiruna for the sake of brevity.] Kiruna has been known for the almost 250 years since its discovery in 1736, but it was not the first iron-ore body to be found in Norrbotten nor the first to be mined. The first-known ore body was that of Junosuanda, found in 1644, but it never amounted to much. The ore bodies at Gällivare (or Malmberget = ore mountain) were first mentioned in the literature in the early 18th century, and, in the middle of that 100-year period, mining and production of iron were conducted on a minor scale with the ore being transported by reindeer-power to small blast furnaces in the neighborhood. Because of the high percentage of phosphorus in the ore (over 1 per cent), no appreciable production was attained until after the commercial success of the Thomas process for smelting high-phosphorus ore made profitable the construction of a railroad from Luleå to the Gällivare mines. Between 1898 and 1924, nearly 46 million tons of ore were produced from the various ore bodies at Gällivare.

A few years after the railroad reached Gällivare, it was extended to Kiruna and, from there, to the year-round, ice-free port of Narvik in Norway. This rail connection allowed ore shipments to be made throughout the year, through Luleå while the navigation was possible in the Gulf of Bothnia and through Narvik in the winter. During the period of railroad construction, the Kiruna ore body was being readied for open-pit exploitation, and mining of the ore was begun in 1901.

No further mines were opened until about 1925 when Tuolluvaara (about 7 km northeast of the center of the Kiruna pit) and Luossavaara (a continuation of the Kiruna mineralization that begins about 1.8 km northeast of the point at which the Kiruna ore ends on the south shore of Luossajärvi) both were put into operation. Production at Luossavaara ceased near the end of the 1960s but that at Tuolluvaara appears still to be continuing. Beginning in about 1955 and ending in 1959, production was taken from the Mertainen ore body about 10 km northwest of the old Svappavaara copper mining region. Svappavaara, in turn, is slightly less than 40 km southeast of the Kiruna mine. Also in this region is the Leveäniemi mine which began operation between 1960 and 1964; also in this area are the Gruvberget and Tansari ore bodies, 2 and 4 km generally west of Leveäniemi.

In the same general time span, about 1964, work was commenced on the Nukutusvaara mine, one of the Per Geijer ore bodies (of which more later), which has its center about 2 km NNE of Luossavaara (but in a different geologic horizon). The Haukivaara mine was opened in the 1965-1969 period and is located a little over 1 km east of the southern end of the Luossavaara deposit; it is in a higher geologic horizon even than the Haukivaara mine.

The mines mentioned above are all of those that Grip (1978) listed in his table of producing mines. One of the most interesting mines not in that table is the Rektor mine from which some production obviously has been taken but which certainly never was more than a small operation; it will be discussed later.

In the period from 1898 through 1974, production from Kiruna was nearly 500 million tons; this means that the total to present time must be somewhere in the neighborhood of 600 million. The grade of ore recovered from Kiruna was 61.3 per cent Fe; 0.97 per cent P; 0.03 per cent S; and 0.07 per cent Mn. From Gällivare, about 230 million tons were mined from 1898 through 1974 with a grade of 54.3 per cent Fe; 0.67 per cent P; and 0.03 per cent S. From 1925 through 1974, the mine at Luossavaara produced nearly 200 million tons of ore with a grade of 63 per cent Fe; 0.3 per cent P; 0.02 per cent S; and 0.06 per cent Mn; in the same period, Tuolluvaara provided about 200 million tons with a grade of 61.8 per cent Fe; 0.25 per cent P; 0.01 per cent S and 0.04 per cent Mn. The grades from Leveäniemi, Mertainen, Nukutusvaara, and Haukivaara all are appreciably lower than those just cited, averaging about 50 per cent Fe with the last two running well over 3.0 per cent P.

Grip and Frietsch (1973) estimated the reserves of iron ore in the general Kiruna-Gällivare region as being over 3.4 billion tons over 2 billion at least must be in the Kiruna mine proper; the Gällivare ore bodies probably contain as much as 0.6 billion tons. The remaining ore bodies may include as much as 0.9 billion tons or less. Of this, probably 0.6 billion tons are in the mines of the Svappavaara group (Leveäniemi and Mertainen are the main ores). About .16 billion tons are in the Ekströmsberg-Pattok group (not previously mentioned) that are some 30 to 40 km southwest of Kiruna. The outcrop or sub-outcrop areas of these ore groups are: (1) Kiruna, 700,000 m^2;

(2) Svappavaara, 400,000 m^2; (3) Ekströmsberg-Pattok, 150,000 m^2; and (4) Gällivare, 185,000 m^2.

In northern Norrbotten, the supracrustal rocks are about equally divided between those of sedimentary origin (now mainly gneisses) and igneous bodies; they are arranged in belts that are oriented north-south and NNE-SSW with an appreciably less distinct east-west orientation also present (Frietsch, 1980). Recent regional geologic mapping has shown that the Archean fold belt is more than 2700 to 2800 m.y. old; these cover only a small area in the northern part of the district. The remainder of the Precambrian is composed of rocks involved in the Svecokarelian fold belt. In areas where radioactive ages have been determined, the Svecokarelian folding took place about 1800 m.y. ago; in Norrbotten, however, some igneous and volcanic rocks are post-orogenic in relation to this folding. These younger rocks, however, also are folded and metamorphosed, indicating a post-Svecokarelian orogeny to which the formation of the 1500 m.y. old granites may be related.

The metamorphism seems to have reached the amphibolite facies in the entire rock assemblage; these rocks in many places contain scapolite resulting from a quite late-stage regional matamorphism.

In the area north of Kiruna, Archean rocks are exposed; these are biotite-bearing, weakly schistose, microcline, porphyritic granite with a radiometrically dated age of 2759 to 2899 m.y. Farther to the north and east, next to the Finnish border, huge areas are covered by gneiss that probably belong to the basement.

Within the Svecokarelian portion of the stratigraphy, four supracrustal groups have been identified, but no major unconformities have been noted between any two of them. Oldest of these is the Greenstone group, which, in the Kiruna area, is called the Kiruna Greenstone group. Under the main part of this greenstone are narrow horizons of quartz-bearing conglomerate and quartzite that locally contain siltstone and limestone-dolomite. This Tjärro quartzite is the lowest member of the Greenstone group and was formed as an epicontinental facies during the Svecokarelian. The Greenstone group, however, is principally composed of splitic lavas that, in many places, are basaltic, pillow-bearing extrusives. Some andesites and peridotites also are parts of this group. All of these rocks are subalkaline to alkaline, and locally, hypabyssal rocks of this character are present. Mainly in the stratigraphically higher part of these mafic volcanics are intercalations of tuff, tuffite, phyllite, graphite-bearing schist, limestone-marl, and chert. Frietsch considers that these formations were laid down in a closed-basin environment during the initial stages of the Svecokarelian orogeny. Frietsch reports that similar rocks in Finland have been dated as being 2000 to 2200 years old.

Overlying the Greenstone are minor thicknesses of mica schists and conglomerates known as the Kurravaara conglomerate in the Kiruna region; this schist-conglomerate group is of restricted extent but forms a distinct marker horizon where it does occur. The next unit in the rock sequence is the Porphyry group it is mainly sodium or sodium-potassium intermediate, subalkaline to alkaline rhyolitic to trachytic rocks. Although this group is limited in area in Norrbotten, it is present throught the Kiruna region. Mostly the original porphyritic textures have been removed by metamorphism which normally has reached no farther than the greenstone facies. The metamorphosed variety is fine-grained and equigranular; in a few places it is coarser and gneissic. Among the intermediate volcanics, in the main, are narrow intercalations of metasediments.

Not considering the ores in their position in the stratigraphy for the moment, overlying the porphyries are (oldest) the Lower Hauki rocks and, above them, the Vakko sediments. The lower Hauki rocks are mainly flows, agglomerates, and tuffs. These volcanics later were covered by the Vakko sediments that contain fragments of both porphyries. It appears certain (to Geijer, Ödman, and Frietsch) that the porphyries and the volcanics of the lower Hauki rocks were emplaced before the introduction of the ores. Parak, Forsell, and Godin, on the contrary, believe that the ores were produced as volcanic-exhalative sediments on the (somewhat) eroded surface of the syenite porphyry; later the ores were covered by the quartz porphyry.

The Per Geijer ores are now structurally above the quartz porphyry and, as Geijer, Ödman, and Frietsch believe, were intruded between the quartz porphyry and the overlying lower Hauki rocks. This is denied by Parak, Forsell, and Godin who hold that the Per Geijer ores were formed at the same time as the main Kiruna and Luossavaara ores by volcanic-exhalative processes that, elsewhere, produced these latter ores. Parak, Forsell, and Godin think that the present position of the Per Geijer ores above the quartz porphyry resulted from folding followed by overthrusting.

Thus, to Geijer, Ödman, and Frietsch, the structure of the rocks of the Kiruna region is that of a monocline with the dip generally to the east, and the rocks are in a stratigraphically normal position. To the other three, however, the Kurravarra and the Hauki conglomerates are of the same stratigraphic age; that is, they are younger than the syenite porphyries and older than the ores. The last three authors, in the cross-sections they present (1980), show the Luossavaara ore as being continued over an anticlinal crest in the Rektor ore and the lower Hauki rocks. They have not gone so far, however, as to show how they connect the Tuollavaara ores with those of Luossavaara and Rektor. To Geijer and his colleagues, the Tuollavaara ores are a separate intrusion that was of the same age as the main Kiruna ore but followed a different path to reach its final resting place in syenite porphyry to the east of the Vakko sediments from which the prophyry there is separated from the Vakko by a fault contact.

Forsell and Godin show a definite anticlinal structure with the Kiruna ore on the east limb but no continuation of the Kiruna ore on the west flank. This simply suggests that the volcanic-exahalative process that produced the Kiruna ore [in the opinion of these two authors (and Parak)] was confined to a rather narrow, more or less north-south extending sea trough.

Obviously, it is more favorable to the volcanic-exhalative hypothesis if the anticlinal structure is proved correct and more so for the magmatic if the monoclinal-faulted structure is more accurate. But it appears that the decision as to the favored manner of introduction of the ores into the Kiruna area strongly influences the choice of a structure for the region.

Another major difference between the two interpretations of the structure and genesis of the rocks and ores of the district depends on the genesis and character of the lower Hauki rocks. Locally the lowermost of the Hauki rocks was designated by Parak as "detritus of porphyry". The clastic nature of this material is admitted by Geijer and Ödman (1974), although Geijer (1950) had called it a phase of his Rektor porphyry which carried feldspar spherulites or spherulite-fringed phenocrysts in a ground mass made up entirely of secondary quartz. If this material was of clastic origin, as Geijer seems now to admit, it is difficult to reconcile the two descriptions. It appears probable, however, that the basal portion of the Rektor porphyry was developed by weathering and some sedimentary reorganization on the upper surface of the quartz-bearing porphyry and then was covered by the true Rektor porphyry that was extruded as a silicic lava in which recrystallization and metasomatic alteration have destroyed the primary textures. The main minerals of this reconstituted lava are quartz and K-feldspar, plus sericite, calcite, and iron oxides. Locally some of this lava was devitrified and later converted to silicified glass in which spherulites of K-feldspare are, in places, aligned in bands to give the rock a sedimentary appearance. If the lower Hauki rocks (the Rektor porphyry in Geijer's sense) were formed as a continum with the underlying quartz porphyry, the iron ores would have been introduced after probably both the Hauki and Vakko rocks had been laid down. If the Per Geijer ores between the quartz porphyry and the lower Hauki rocks were formed by a volcano-exhalative process, it is difficult to see how the "detritus of porphyry" could have been formed on top of the ore. The balance of probabilities is that the lower Hauki rocks are a melange of lavas and weathered and eroded rocks, the debris of which accumulated in sufficient quantities in certain places to be identifiable as a clastic sediments but which probably only make up a small part of the largely volcanic sequence of the Hauki rocks. If this is true, then the Per Geijer ores must have been a later intrusion.

The Vakko rocks were developed with a minor unconformity on the lower Hauki rocks and contain fragments of both syenite-syenite porphyry and quartz porphyry. Thus, it is certain that the Vakko rocks were deposited after the porphyries were emplaced; it still is possible that the porphyries were under a relatively thin cover of lower Hauki rocks. Geijer, however, thinks this unlikely as the ground-mass textures of the two porphyries are too fine-grained to have been anything but flows.

In his earliest work on the district, Geijer thought that the magnetite-apatite ores were flows, but further work convinced him that the ore was intruded between the two porphyries. The most convincing evidence for this belief is that ore-cemented breccias occur along both foot- and hanging wall contacts, structures that could not have developed unless both walls were in place when the ore magma arrived in the area. The angle to the horizontal at which the ore was intruded still is subject to question. Geijer thinks that the formations had not been folded when the ores were emplaced. His best evidence for this is that igneous activity producing porphyries both in the Kiruna-Luossavaara line and the Per Geijer line was in progress both before and after ore deposition so that there may not have been time for folding between any two stages of magmatic activity. The fractures into which the ore entered were due to the force of intrusion.

Vogt, on the contrary, believed that the ores were introduced after the rocks had attained their present attitude, apparently basing his idea on the need for earth movements to have developed quite cleanly broken fissures through which the ore magma could have moved.

The magnetite-apatite intrusion (granted there was one and that the ores were not volcano-exhalatively produced) was not the last event in the igneous activity in the Kiruna region. Two generations of quartz-bearing porphyry dikes cut both the hanging- and footwall porphyries, and the later of the two dikes cuts the ore as well. At some time after the introduction of these dikes, dikes of granophyre (aplite) also were intruded; these may be appreciably later than the porphyry dikes because they do not contain the common to abundant magnetite of the other igneous rocks.

In addition to the magnetite and apatite, the ore contains a little diopsidic augite (now largely converted to uralitic hornblende). A small part of the magnetite has been altered to martite (hematite pseudomorphic after magnetite). The age of the apatite relative to magnetite is not completely clear, but it appears that normally magnetite crystallized before apatite. This can be seen in the common occurrence of irregular streaks or veins of apatite mixed with magnetite cutting fractures in magnetite. Ore volumes unusually rich in apatite will be surrounded by apatite-poor rock volumes, indicating an appreciable degree of apatite segregation prior to any large degree of magnetite crystallization. Very locally, a "stratified" ore is found in which apatite crystallized before the bulk of the magnetite, and slabs of such ore ordinarily are enclosed in much purer magnetite. The usual apatite grains in relatively pure apatite masses are thick prisms arranged at random but where smaller and thinner forms occur, they generally show a fluidic texture.

To some extent, the wall-rock porphyries adjacent to the Kiruna ore body have been altered, although the alteration is far less than would be expected around a hydrothermal ore deposit of comparable size. The iron deposit at Cornwall in Pennsylvania is certainly not one-tenth of the size of the Kiruna ore body, yet the alteration at Cornwall is far more obvious and extensive. What alteration there is at Kiruna has affected the syenite to a greater degree than the quartz-bearing porphyry and is much more impressive in volumes of ore breccia than where the contact between ore and wall rock is essentially unbroken. In the syenite porphyry, where ore breccia is more strongly developed than in the hanging wall and in which there are more silicates (as opposed to quartz) than in the quartz-bearing porphyry, the alteration is more highly developed. Alteration, however, is more apparent in the normal pink quartz-bearing porphyry than in the greenish syenite because the principal alteration minerals are actinolite and white albite feldspar.

If, as Geijer suggests, the ore was injected into the volumes of broken wall rock as a molten and quite fluid solution of apatite in magnetite, the ore breccia texture is exactly what would be expected. Into the broken debris along the margins of the fissure, the ore magma would have penetrated and would have had far more surface area with which to react than would have had the main portions of the magma against the solid, unbroken parts of the porphyry walls. Between the silicate-rich masses of broken rock of the breccia and the iron-rich melt, reactions would have occurred, and it is reasonable that the products should have been iron-rich amphibole and albite. In fact, in some of the narrower fractures, the reactions have gone so far as to convert all of the injected magnetite into actinolite.

Although the alteration reactions are dominantly those of the melt with solid, there is some evidence that the magnetite-apatite magma contained some water, the most striking of which are the uralitization of the primary pyroxene and the occasional, small, late-stage veinlets of quartz and primary hematite that are found in the ore and in the adjacent wall rock.

Another peculiar feature of the apparently deuteric activity is the amygdaloidal character of the upper part of the syenite porphyry. The amygdules are filled with hornblende, magnetite, titanite, and apatite and are surrounded by rock volumes that are composed almost entirely of feldspar and are free of the minerals in the amygdules. Geijer's thought that these amygdules were formed under the influence of water in the late stages of the crystallization of the syenite seems reasonable, particularly if the syenite was extruded as a flow as Geijer also believed.

In recent years, Geijer's theory has been questioned by Parak who suggests that the ores are the result of volcanic-exhalative activity. Parak (1975) looks at Kiruna through far different eyes than Geijer or Frietsch. He believes (Parak, 1975) that, after the lavas from which the syenite-syenite porphyry complex developed on crystallization, fumarolic activity began. In topographically suitable water-filled basins, exhalatively introduced iron oxides and apatite were simultaneously deposited. Parak says, however, that it is difficult to determine which factors caused the rhythmic formation of the ores; he considers these factors to include: temperature, composition of gases produced, solutions, and rocks traversed by the solutions, the pH and Eh values, and the amount of oxygen and electrolytes. He also believes that he has proved that sodium/potassium ratios change rapidly and that this may have played an important part in determining the minerals that precipitate in this postulated marine environment. He even considers it possible that the original iron oxide was hematite, later altered to magnetite when the macro- and microevidence seems to point in the opposite direction. He would produce the necessary conditions for such a reduction in changing the water depth, that is, probably increasing it.

Parak has decided that, after the solidification and some erosion of the iron-apatite ore, came extrusion of the quartz porphyry. These silicic lavas covered the entire ore surface and the erosionally developed conglomerate above it. Pauses in the eruption cycle resulted in erosion products and chemical sediments intercalated into the porphyry sequence. Before these silicic lava extrusions stopped, fumarolic-solfataric activity began. This resulted in deposition of the Per-Geijer type ores that consistently are located at this stratigraphic horizon. Before the long period of erosion that followed the completion of the Vakko rocks, the basins in which this type of ore is contained probably were appreciably more continuous than they are now.

The Vakko beds, according to Parak began with quartz-banded ore, followed successively by Hauki-type syenite porphyry, Hauki hematite, quartz-bearing porphyry, graywacke, and phyllite. The rather narrow thicknesses of these rocks were followed by the considerable accumulation of Vakko sandstone with conglomerate.

The southern end of the Kiruna ore proper is twice faulted; these faults strike about NE-SW and probably dip southeast. The ore north of these faults has a dip of about 75°NE. Within this (Sofia) ore, breccias with usually sharply bounded fragments of both syenite porphyry and quartz-bearing porphyry occur within the ore. The larger fragments of this ore are normally quite fresh but do show a red-colored margin against the surrounding ore; the smaller fragments, however, generally are altered to a fine-grained red rock without any porphyritic texture - these fragments are veined by ore and grade gradually into it. This process has completely destroyed the outline of the fragments, thereby giving them the appearance of being younger than the ore and cutting it. These "schlieren" of red rock also have been found in ore outside the breccia areas; this relationship gives the impression that the iron-oxide melt assimilated a large amount of wall rock fragments that the force of the intrusion broke off from the walls. Geijer and Ödman believe that the occurrence of ore breccia this near to the cross faults indicates that the Sofia ore is not continued to the south of the faults but that the ore there derived from separate intrusions.

Under the lake (Luossajärvi) there is a zone of ore dikes and veins within the quartz-bearing porphyry that parallels the ore in its hanging wall. It was not, and probably is not now, economic.

The deposit of Luossavaara is located in the same level in the volcanic sequence of the Kiruna area as does the Kiruna deposit proper. Luossavaara apparently does not have an actual connection with the Kiruna ore. Between the work of Giejer in 1910 and Parak in 1973, a fairly comprehensive understanding of the relations of the Luossavaara ore to its wall rocks has been obtained. There probably are two sets of ore veins in the syenite porphyry, one being segregations within that rock and the other, larger but far less numerous, are apophyses from the ore. Parak found that the segregations actually are much more abundant than the surface indications would show, but it is now certain that the apophyses do cut the segregations and are definitely younger than the segregated ore.

Parak thought, however, that the true ore veins at Luossavaara were not connected with the ore body because the ore body is bounded along the footwall by a seam of chlorite (the sköl of Swedish miners). Locally, the sköl changed to a quartz vein with some patches of chlorite; this vein was, at a maximum, more than 2 m wide and branched out into the ore which it brecciated. There must, therefore, have been movement along the footwall contact of the ore after it had been emplaced. Thus, the later dikes and narrower veins of ore in the footwall really are offshoots from the main ore body. In Luossavaara, the hanging wall of the ore does not contain any distinct ore veins in the quartz porphyry nor is there any amphibole skarn. Only inconspicuous, microscopically distinguishable veins of magnetite can be recognized in the porphyry where it is finely brecciated. Between this breccia and the ore is a mass, some dm wide, of vuggy quartz, chlorite, and tourmaline, analogous to the material found along the footwall.

The presence of ore inclusions in the quartz porphyry has suggested to some that the ore body could not have been derived by intrusion. Geijer, instead, argues that the ore cannot have been emplaced differently than the nearby and larger Kiruna ore body proper and that the inclusions of ore in the quartz porphyry must have come from an ore body, now removed by erosion, which he thinks lay above the present surface to the north or northeast of the Luossavaara summit.

The quartz-bearing porphyry at Luossavaara developed as pyroclastic flows; the porphyry, therefore, shows its normal development. Inclusions are scattered in some zones in the porphyry, but they mostly are a kind of agglomerate, with the inclusions more or less rounded that derived from the underlying syenite porphyry or, to a lesser extent, from the quartz porphyry itself. Inclusions of the standard ores of Kiruna- or Luossavaara types, always angular, also are present. This conglomeratic material probably was the result of a series of pyroclastic flows, interrupted by a period of erosion which accounts for the normal conglomerate containing boulders of both types of porphyry and of rounded masses of iron ore. The extent of this conglomerate, however, was limited, and it probably was no more than the filling of a stream bed rather than part of a broad beach. If this is so, then the conglomerate proves no more than that, as Geijer and Ödman (1974) thought, the pyroclastic porphyry was formed by a series of eruptions and the conglomerate developed during one of them and cannot be used as proof against the intrusive nature of the Luossavaara ore. Geijer and Ödman, however, say that more thought must be given to the problem and further discoveries of material of this sort must be sought.

The Tuolluvaara ore deposit, located as previously stated above, is separated from the Kiruna ore proper by at least one fault of great throw. Geijer and Ödman believe that this shows, better than any other deposit, the intrusive nature of the ordinary type of apatite-bearing iron ore. As was true at Grängesberg, before mining the Tuolluvaara suboutcrop was swept clean of overlying glacial debris, an ideal condition for surface observation. The ore is completely enclosed in quartz-bearing porphyry that is exactly like that at Kiruna, except that the feldspar phenocrysts are smaller at Tuolluvaara. The rather irregular placement of the ore bodies here seems to have been produced by a fissure system present when the ores were introduced. Parak thought that these ore bodies were once continuous as, of course, would be required by his concept to the ore bodies as having been formed sedimentarily by material volcanically exhaled along a narrow trench. The faults at Tuolluvaara are clean-cut and straight and never grade into a system of extensive ore breccias. Geijer (1967) demonstrated that the intrusion of ore was in two distinct episodes using the same fissures. On the surface, it is possible to see directly the transition from an ore body of economic extent, through branching ore dikes of gradually lessening thickness, to typical ore breccia. An expansion of the total volume of rock required for the production of such structures must have been due to forceful injection. Thus, the fractures through which the ore initially was intruded were of different origin and different character than those caused by the fact of injection under high pressure.

The final category of ores in the Kiruna area to be considered here the so-called "Per Geijer" ores. The bodies (neglecting for the moment the deep-seated Lappmalmen ore body) are located at the contact between the quartz-porphyry of the Kiruna type (below) and the Lower Hauki rocks above. One exception to this is northwest-trending apophysis of Henry (Rektor-type) ore body that leaves this structure far enough that its tip penetrates into the underlying syenite porphyry and ends in a series of ore breccias in the syenite porphyry. The other exception is the Haukivaara ores that lie at the contact between Lower Hauki rocks (below) and the Vakko sedimentary rocks (above). The more southerly of the true "Per Geijer" ores is the Rektor deposit that lies on the contact of the quartz-bearing porphyry (below) and the Lower Hauki rocks (above). The more northerly is the Nukutusvaara ore body, the southern extension of which is known as the Henry deposit. A small body still further north of the Nukutusvaara ores is apparently known (or was known 30 years ago) as the Simeon deposit.

At depth beneath the Rektor ore body lies the Lappmalmen body that was discovered by magnetic measurements. Part of this ore body is a direct continuation of the Per Geijer (Rektor) ore horizon, whereas the lower part of it is certainly displaced by faults, having been moved upward along a series of faults so that its highest position is about 400 m below the surface some 1000 m west of the Rektor ore. The Lappmalmen ore body is about 2500 m long at least and has an area of about 200,000 m^2 on the 700-m level; this area probably is larger with depth, and the northern end has not yet been followed to its end. The ore minerals at Lappmalmen are magnetite (upward from the footwall of the ore) and hematite above that. The ore is higher in phosporus and lower in silica than the remainder of the Per Geijer ores. The wall rock always is quartz-bearing porphyry on the footwall; the hanging wall is the strange rock already mentioned that is designated by Parak as "detritus of porphyry" but is described by Geijer as Rektor porphyry which is considered by him to be obsidian, the glassy ground mass of which has been completely silicified.

The Lappmalmen ore body (Geijer and Ödman, 1974) exhibits a transition as it is followed upward from apatite-banded iron ore to quartz-banded ore. These authors consider Parak's description of this transition to be as clear and decisive as drill-hole evidence can make it. In surface exposures over the Lappmalmen zone in the Rektor-Henry-Nukutus ore areas, quartz-banded, iron-bearing rock is present but forms the base of the Lower Hauki beds with Per Geiger ore interposed between quartz porphyry on the footwall and Lower Hauki beds on the hanging. The basal Lower Hauki, Parak says, is his detritus of porphyry (d-of-p), but Geiger called it Rektor porphyry; this rock extends as far north as 1 km from Rektor. No ore is known for 1.8 km north of Rektor, where Per Geiger ore begins again in the Henry deposit and continues on north through the Nukutus body to end 2.2 km farther north, against a NNE Fault beyond Syväjärvi.

The only explanation that credits both men with sound observation requires that what Geiger saw years ago had been removed by mining prior to Parak's work in the area in question. It is apparent that this problem can be solved only when means can be found to reconcile the obvious differences between what each man saw.

The ores of Rektor, Nukutusvaara, and Henry, however, are not of the banded type of ore in Lappmalmen. Instead, so far as I can tell, despite Parak, definitely, and Geijer somewhat less so seeing sedimentary structures in the ore, it would appear that the first major difference between Kiruna-Luossavaara ore and Rektor ore is that the apatite content in the latter type of deposits is appreciably higher than in the former. Also, the content of hematite is higher in the Rektor ore than in those bodies along the contacts between the two porphyries. In some places along the strike from Rektor, the material along the contact is almost entirely apatite. Apparently, the higher the apatite content, the later in the differentiation of the parent magma of the iron-phosphorus ores' immisibility occurred or the later the two immiscible fractions parted from each other, the higher the phosporus content in the iron rich fraction.

A second, and very real, problem to solve is why there was so much conversion of magnetite to hematite in the Rektor ore, and a third is why the wall rocks of the Rektor ore body, both foot- and hanging wall, are so badly altered?

First, the magnetite to hematite change. This, of course, requires oxidation of iron from the ferrous to ferric state. This is not normal behavior for magmatic melts or hydrothermal solutions. Usually, the trend is exactly the opposite, ferric iron is converted to ferrous as witness the bleached rocks around many hydrothermal deposits or the consistent dominance of magnetite over hematite in most bodies of magmatic origin (e.g. Kiruna or Iron Mountain Missouri). If hematite is to occur in amounts greater than magnetite or if hematite is to be a sizeable proportion of the iron oxides in the ore, some process must have been operative that would have provided nascent oxygen to the system. The ways in which this might have been done are not numerous.

One of the suggestions for providing oxygen in the amounts needed is the dissociation of water. The dissociation constants of water at the temperatures and pressure that must have obtained in the formation of the Rektor (or Gällivare ores) are so small that the amount of free oxygen available would not have been sufficient to initiate the reaction, much less bring it to an appreciable stage

$$(1)\quad 2H_2O + 4Fe^{+2} + O_2 = 4Fe^{+3} + 4OH^{-1}$$

of completion [see equation (1)]. Other methods of providing the necessary oxygen also can be shown to be impractical.

As Kennedy (1948) has pointed out, fresh basalts are universally less highly oxidized than gabbros. This means that the partial pressure of volatiles (and therefore of oxygen) in a flow or near-surface intrusion, mafic or silicic, is appreciably less than in a deep-seated intrusion unless the rock cover overlying the magma is remarkably pressure resistant. In any magma, the following relationship holds [equation (2)]:

$$(2)\quad KO_2 = \frac{(H_2O)^2}{(H)^2\ (O_2)^2}$$

If the total content of H_2O is less in a system near the surface (as it normally would be), the O_2 and H_2 contents would be lessened (even though O_2 comes from other sources than water) and the equilibrium in reactions involving O_2 as an oxidizing agent would be shifted in the direction of less oxidation.

Since the proportion of hematite to magnetite in the Rektor deposit is higher than in any of the other ore bodies of the Kiruna region, the degree of loss of volatiles, even allowing for the some escape to the immediately adjacent portions of the Rektor porphyry probably was appreciably less than from those ore bodies in which the hematite-magnetite proportion was much less.

From the considerable wall-rock alteration that surrounds the Rektor ore (both on the foot- and the hanging wall), much greater than around Kiruna, it seems that it is most probable that the Rektor melt was much higher in water than in the stratigraphically lower melts that formed Kiruna and Luossavaara. This well may be because less volatile material was lost from the Rektor melt than from the others, but it also is highly probable that the Rektor melt was separated from (became immiscible in) the silicate melt at a later stage than did the Kiruna iron-rich melt. The same water-rich melt-type probably also obtained in such similar deposits in which the amount of hematite is high as Cerro de Mercado in Mexico, Iron Mountain, Missouri, and Gällivare (of which last more below). Those iron-rich melts that produced such deposits as Kiruna, however, could not have been even near to being boiled dry or FeO or even native iron would have been developed, and these minerals never have been reported from magmatic magnetite deposits. In the group with less water to begin with or with a greater near-surface loss would fall Kiruna and Luossavaara, Grängesberg, several of those in the Urals, and El Tofo and Algarroba in Chile.

It remains to explain a possible mechanism for the production of nascent oxygen from the relatively high content of volatiles retained in such melts as that at Rektor. It has been indicated above that the small partial pressure of oxygen gas that could have been developed by the dissociation of water could not have been sufficient to have reduced the large fraction of the original Rektor ore from magnetite to hematite. There must, therefore, have been an additional source of oxygen available from the dissociation of oxygen-bearing gases far less stable than water or carbon dioxide. The most likely gases to meet these requirements are the various oxides of nitrogen. One of the possible series of reactions that may have obtained is:

$$(3)\quad 4HNO_3 = 2N_2O_5 + H_2O$$

$$2N_2O_5 = 4NO + 3O_2$$

$$4NO = 2N_2 + 2O_2$$

These are reactions favored by high temperature and relatively low confining pressure, although the last one is the least likely to go forward unaided.

In the early stages of the formation of the earth, nitrogen was most likely to have been combined with hydrogen as a nitride, but, because of the endothermic nature of this reaction, the amounts formed probably were not large. As the temperature of the cooling earth dropped, oxides of nitrogen would have been produced by reactions between nitrides and dissolved oxygen on the one hand and oxygen and oxygen ion on the other. These reactions would have been favored by high temperatures and pressures. These nitrogen compounds would have reacted with water to form a variety of acids, the principal product of which would have been NO_3^{-1}. As there are no natural nitrates (probably because of their high solubility), nitrates would have been exploded into the atmosphere in volcanic eruptions and have been concentrated in surface deposits such as those of the Atacama desert of northern Chile. If, however, fluids or melts high in nitrates had been trapped near the surface, such reactions as those in equation (3) could have taken place and free nitrogen and oxygen formed, the latter of which could have reacted with magnetite to convert it to hematite as equation (4) shows:

$$(4)\quad 4FeFe_2^3O_4 = O_2 = 6Fe_2O_3$$

As long as melts containing nitrogen-oxygen compounds or radicals are held well below the surface, the various nitrogen-oxygen anions would be quite stable, but, once brought near the surface, the lowered confining pressures and still high temperature would have caused dissociation of these anions into nitrogen and oxygen or oxygen and lower nitrogen oxides.

These reactions would insure a sufficient abundance of nascent oxygen for appreciable oxidation of the magnetite in such a deposit as Rektor or Gällivare the exact proportion of hematite to magnetite depending on the amount of free oxygen supplied by the break-down of the nitrogen-oxygen material available.

It seems probable, if not certain, that the conversion of magnetite to hematite at Rektor was due to oxygen that came in with the molten iron-oxygen material but reacted with the magnetite after the bulk (if not all) of it has solidified. I prefer this concept that the hematite was produced by later hydrothermal solutions. The degree of alteration in the wall rocks at Rektor (especially in the hanging wall) is large enough to raise a serious question as to the capacity of the melt to contain enough water (containing the proper ingredients) to effect the huge amount of wall-rock alteration around that ore body. The whole problem of Rektor alteration of the ore from magnetite-dominant to hematite-dominant and of the wall rocks to a mixture of sericite, ankerite, and, in places, a little biotite, apatite, magnetite, and even less tourmaline needs further study.

Parak claims that fragments of ore and rock types identical to those in the lower Hauki series are found in the upper sections of the quartz-bearing porphyry. This suggests to him that the deposition of the lower Hauki must be older than, or contemporaneous with, the extrusion of the quartz-bearing porphyry.

Parak's position is challenged by both Geijer and Ödman (1974) and Frietsch (1978). Frietsch points out that Parak considered that the Kiruna ores were deposited in a volcanic-sedimentary environment. Because the rocks Parak has interpreted as sediments are but poorly represented in Kiruna area and can equally well be considered to be of magmatic origin, Frietsch thinks that Parak's position is a misleading one. Unfortunately, the criteria put forward by these radically opposed authors do not definitevely settle the problem. When it is remembered how much field and laboratory work has been done on the Kiruna rocks and ores and how different are the opinions of these two authors, it is remarkable that they are so far apart in their opinions.

Frietsch points to Tuolluvaara as a place in which ore in veins and dikes cuts the host rocks in branching networks; the host rocks have been broken into angular fragments, it appears to Frietsch, by the ore forcing its way into the rocks along, and out from, joint planes. This situation convinces Frietsch that the ore did not enter the Tuolluvaara area through fissures that already were present. Parak, on the contrary believes that the fillings of the veins were provided by remobilized sedimentary ore from the main ore body under the influence of metamorphism. The appearance of these ores certainly agrees better with Frietsch's concept than that of Parak. These veins contain magnetite, hematite, apatite, and sparse amphibole (exactly the mineralization of the main Kiruna ore), and it is difficult to see how the remobilized ore, if it was such, could have retained the same mineral character as its parent.

It appears from what already has been said about the ore in the Kiruna region that it is more likely that the Kiruna ores were deposited by the crystallization of a molten, iron-rich magma that probably developed from the parent magma of the quartz porphyry as the quantity of iron oxides (and the other materials of the ore magma) became too large to remain dissolved in the quartz-porphyry melt and gradually separated out from the silicate melt. This immiscible iron-rich portion of the original magma gradually changed composition with the further crystallization of the quartz porphyry so that the early immiscible fractions were lower in apatite than the later ones. Further, the apatite-rich portion also contains greater quantities of water and other volatiles than did the initial immiscible fraction. If this outline is correct, then the Kiruna ores proper (including Luossavaara and Tuolluvaara) were intruded before the more apatite-rich ores of the Per Geijer type, but both were products of the same differentiation cycle. Further work may find fatal flaws in this concept, but, up to now, it appears to provide a more satisfactory explanation that does Parak's volcano-exhalative theory. Granted that the magmatic concept is correct, then the ore of the Kiruna region should be classified as magmatic-3b. Were Parak correct in his explanation, they would be categorized as Sedimentary-3A, plus Metamorphic-C.

The age of the Kiruna deposits, both Kiruna proper and Per Geijer probably is about that of the volcanics in the region for which Rb/Sr determinations indicate an age-range of 1605-1635 m.y. This places the ores in the latest middle Precambrian or the earliest late Precambrian with the former being preferred.

The ore bodies of the Gällivare ore field center around the town of Malmberget, about 80 km south southeast of Kiruna and about 4.0 km northeast of Gällivare (town), which last is on the main rail line from Kiruna to southern Sweden. The main ore body, known as Stora Malmlagret (great ore layer) is composed of a number of nearly separate, but connected, ore areas that extend for 4000 m in an irregularly northeast-southwest direction. Southeast of the northeast end of Malmlagret is the Koskullskule ore body and the Dennewitz ore body lies between the two just mentioned. South-southwest of Koskullskule about 2 km is Kaptenslagret. Mining began in the area in 1888, and, up through 1974 180,000,000 tons had been mined. The total area of the ore bodies is 185,000 m^2 with most of the production now coming from Kaptenslagret. Folding has been much more prominent in Gällivare with the ore bodies being elongated parallel to fold axes that run 45°S-SSW. In the Dennewitz-Koskullskule area, the ores have been so folded as to concentrate them in a shorter horizontal distance than is true of an equal ore area of Storamalmberget. The Kaptenslagret is either a separate ore zone parallel to Storamalmberget, or it may be a southern limb of that ore. It appears that the ore of Stora Malmberget is higher in phosporus than that of the smaller ore bodies. The horizontal ore width has a maximum of about 100 m, but, in detail this ranges widely; the dip normally is southeast, averaging 60° and ranging between 40° and 80°. The ore is known to extend for at least 700 m in depth; reserves probably are about 600 million tons, and the grade, over nearly 90 years of mining has averaged 54.3 per cent Fe; 0.67 per cent P; and 0.03 per cent sulfur.

For many years, Malmberget has been thought to be a highly dynamometamorphodes Kiruna, and most Swedish geologists still consider it as such. There are, however, certain differences that probably existed before metamorphism and others that were developed by that process and by the intrusion of pegmatites into and near the ore. The only important initial difference between Kiruna and Gällivare ores was the presence of important amount of hematite in the latter.

The host rock at Malmberget is now leptite or leptite gneiss; originally this rock was composed of silicic volcanics that have been recrystallized to a groundmass size of 0.03 to 0.05 mm. As sediments may be intercalated into any volcanic sequence, the possibility exists that some leptites originally were sediments. Many Swedish leptites possibly may have been of that character, even including sedimentary iron ores. At Malmberget, at least the majority of the leptites probably were of igneous origin. The Gällivare leptites are strongly stratified, in contrast to the generally massive rocks at Kiruna. This stratification at Gällivare (or rather ease of ingress along planes of stratification) to a considerable extent controlled the emplacement of the ores.

Two divisions in the leptites and leptite gneisses easily can be seen at Malmberget - into red leptites and gray ones. The red leptites are mainly alkaline feldspar (microcline and albite) with or without quartz, whereas the gray ones contain essentially no quartz but have a high percentage of dark minerals, mainly magnetite and biotite. The plagioclase of the gray leptite ranges between sodicoligoclase and albite; microcline is quite subordinate or entirely absent. These gray rocks also contain not only magnetite and biotite but also some hornblende and pyroxene.

Geijer (1930) points out that an appreciable percentage of both leptites have relict phenocrysts of feldspar (microcline and albite) in coarse perthitic intergrowths in the red rock and plagioclase in the gray. Much of the leptite has amygdaloidal textures, probably indicating near-surface, igneous origin for the bulk of the leptite at least.

Before any ore was injected into the leptites, however, considerable tectonic movement apparently occurred. The ore is found not only in lit-par-lit injections in the country rock but also as cement for large volumes of leptite breccia. In some instances, the force of ore injection was so great as to push leptite beds as much as a few tens of meters apart; by this process, massive beds of ore, comparable to, but much smaller than, the

Kiruna ores were developed. If this were all that had happened to the Malmberget ores, their structure would differ from that at Kiruna mainly in the greater amount of ore breccias and the impressive lit-par-lit banding made possible by the stratification of the leptites as opposed to the massive character of the Kiruna porphyries. Thus, there was no major, all-prevasive fissure at Gällivare as was present at Kiruna; instead, the solid masses of Gällivare ore are separated along dip and strike from each other by volumes of ore-cemented breccia and by interbedded magnetite and leptite too low in grade to mine.

The impressive amount of breccia at Gällivare also distinguishes that ore from that at Kiruna; this Gällivare brecciation occurred before the ore was emplaced and accounts for the much greater irregularity of the ore outline that obtains at Gällivare. The bulk of the folding probably was concomitant with the brecciation. An argument against the brecciation and folding having been post ore is that (Parak to the contrary not withstanding) magnetite is not a mineral that flows readily (as do such sulfides as chalcopyrite). The ore would not have filled the fractures so completely if it had had to be moved after it was solidly emplaced before brecciation.

Probably closely following the emplacement of the ore was the intrusion of a considerable amount of granitic material that generally is pegmatite, although locally aplites are prominent. These pegmatites are mainly masses of red albite and quartz, but they also contain, in places, hematite and apatite that may make up as much as half the volume of a given pegmatite. Geijer thinks it probable that these pegmatites incorporated ore already present in the rocks rather than these minerals having been part of the pegmatitic magma. Amphibole (hornblende) also is present in the pegmatites; it may be indigenous to them or it may have come from ingested skarn or amphibole-impregnated leptite.

Where the pegmatite directly adjoins the ore, impressive reaction rims developed in which hornblende, diopside, and probably garnet are major minerals. In the vicinity of most pegmatites, hydrothermal veins or veinlets containing quartz and hematite are present; these probably were late differentiation products of the pegmatitic melt. The veinlets show in miniature the same types of reaction rims that the pegmatites exhibit adjacent to ore. On the contrary, reactions rims are almost or entirely lacking between ore and leptite except in the breccia zones where the surface area presented to the intruding material is much greater than in the lit-par-lit injections.

In the ore breccia zones, the first step in the reaction between ore and leptite was the development of a halo of disseminated hornblende in the leptite. If the ore in the breccia fractures was in sufficient volume, further production of hornblende converted the volume nearest the vein into massive skarn composed entirely of hornblende or of hornblende and recrystallized albite; disseminated hornblende formed halos out beyond the skarn. Included in the skarn were minor amounts of magnetite, pyroxene, apatite, biotite, and lesser amounts of other minerals, of which scapolite is of considerable interest as it indicates the presence of an appreciable quantity of volatile materials in the ore magma. If recrystallized albite is not intimately intergrown with the hornblende, it formed massive bands of red albite between the hornblende and the original fracture. If the amount of available magnetite was small (i.e. the fracture was narrow), the reaction process usually resulted in completely using up the magnetite so that only skarn remained. In the larger fissures, however, where more magnetite was introduced than could be consumed in reactions with the wall rocks, the result was an ore breccia separated from leptite (from the vein outward) by either mixed albite-hornblende skarn and disseminated hornblende or by successive bands of red albite, massive hornblende, and disseminated hornblende.

The skarn breccias or ore breccias with skarn borders are more prominent in red leptite than in gray but also are to be found, showing the same stages of development in the gray leptite as well. In either type, further, largely mechanical, movement has confused these relationships to various degrees, but there are sufficient places in the mine where the pattern just described can be seen and clearly interpreted.

In most of the ore at Malmberget, magnetite is the principal mineral, but about 20 per cent of the ore volume is hematite. This hematite is located mainly on the footwall (northwest side) of the body and occurs as a matrix for porphyritic crystals of magnetite. Otherwise, there is very little mixing of the two varieties of iron oxide. The next mineral after the iron oxides is apatite that averages 4.5 per cent of the ore, but there are volumes where apatite is essentially lacking (as is true at Kiruna). Hornblende is rather widespread in the ore, but it never is common; some diopside and less biotite also are present.

From this description of the Malmberget ores, they appear to be intermediate between those of Kiruna proper and Rektor (and the other Per Geijer ores); the hematite present is a smaller percentage of the total ore than at Rektor, the wall-rock alteration is more extensive than at Kiruna but less so and of a higher temperature than at Rektor. The hematite at Gällivare apparently is a primary mineral and not a result of the alteration of already crystallized magnetite (as at Rektor); replacement textures are not exhibited by hematite against magnetite. This last relationship, when considered with the magnetite phenocrysts found in some of the hematite, suggests that the oxidation of ferrous to ferric iron occurred after the ore melt had left the magma chamber in which it was immiscibly segragated and, of course, before hematite crystallization began.

The mechanism for the oxidation of ferrous to ferric iron probably was the same as that postulated previously in this discussion for producing the Rektor hematite, but, at Gällivare, the water content of the iron-rich melt was appreciably lower than at Rektor or more was lost or a combination of both.

Even when I visited Sweden for the first time in 1954, the opinion was current among a considerable segment of Swedish geologists that the Gällivare and Kiruna ores were metamorphosed sediments even though Oftedahl would not publish is ideas on volcano-exhalative ore genesis for four more years. In 1956, the thought was that Gällivare was much more likely to be of a sedimentary origin than Kiruna because Malmberget wall rock could more easily be seen as having been formed by sedimentary processes. Now the possibility of Kiruna having had a volcanic-exhalative manner of genesis is accepted among a considerable group of Swedish geologists; I have not seen, however, any recent work that attempts to apply this concept to Gällivare despite the very appreciable similarity between the two deposits.

The progressive change in ore character and relations to host rocks from Kiruna to Gällivare to Rektor strongly suggests to me, at least, that such changes and relationships would have been produced only if these three deposits had been formed by much the same process, one in which the degree of availability of nascent oxygen determined how much of the magnetite in the parent melt would have been changed to hematite either during or after the consolidation of the various iron-rich melts. The differences in water content among the three melts would largely have determined the intensity of the wall-rock alteration.

On this basis, it would seem that the idea of Gällivare having been produced by the dynamo-metamorphism of a Kiruna-like parent body seems unreasonable. The host rocks well may have been strongly metamorphosed, much more so than those at Kiruna, but the amount of metamorphism after the ore had been introduced appears to have been minor. On this basis, the deposit at Gällivare should have the same classification as that at Kiruna, that is Magmatic-3; there is no justification, in my opinion to add "Metamorphic-C" to the Magmatic-3.

ALNÖ ISLAND, NORRLAND

Late Precambrian	Barite, Columbium Apatite, Lime	Mesothermal (Ba) Magmatic-Fla (Cb, Ap)

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Notes

Alnö Island, the location of the bulk of the above-sea-level portion of the Alnö syenitic-carbonatitic complex, is immediately across the narrow Alnö Sund from the port of Sundsvall. This sound follows the trace of one of the major faults that border the Gulf of Bothnia, and the island largely blocks the entrance to the bay of Klingerfjärden at its north end. The portion of the syenitic-carbonatitic complex above sea level is located on the northeast corner of Alnö Island with the outcrop on the island about 10 km northeast of Sundsvall. The center of the complex proper lies under the eastern entrance of Klingerfjärden, about 1.25 km north of the island-gulf contact of the complex; a few off-shoot dikes of the complex are known on the mainland, immediately east of the island. Only the extreme southeastern tip of the sövite core occurs on the northeast margin of Alnö Island; the bulk of the sövite core lies off shore with most of it being beneath the waters of Klingerfjärden Bay but with small amounts being known on the small islands immediately off shore from Alnö Island. The central part of the complex, containing the coarse-grained sövites, covers an area of about 4 sq km and is surrounded by 21 sq km of syenites and fenites. Approximately 65 per cent of the total area of the complex is under the waters of the gulf. The location of the town of Alnö is 62°22'N,17°30'E.

During World War II, a shortage of strontium for military flares prompted the exploration of Alnö Island, but barium was found rather than strontium. Barium was needed by many Swedish industries and the importation of it had been stopped by the war. Von Eckermann discovered two barite dikes [Pottäng and Hartung with production (6333 tons) only from the former] that are between 2 to 3 m wide with marginal zones of fluorite and calcite, and mining on them continued at least until 1951. Apatite-bearing sövite has been mined, finely ground, and used as a fertilizer; its ready break-down to release phosphorus appears to be due mainly to the presence of large numbers of carbon-dioxide-containing inclusions in the sövitic carbonate. The sövite also, in places, contains appreciable quantities of a variant of pyrochlore $[Na,Ca)_2Cb_2O_6(OH,F)]$, but it has not been found in high enough grade to justify mining.

The country rock surrounding the syenitic-carbonatitic complex is migmatitic Archean granite-gneiss that is cut by pegmatites and porphyry and dolerite dikes of Jotnian age. Relics of graywacke are enclosed in these rocks and have been converted largely to mica schists. For as much as 12 km from the center of the main sövite mass, cone sheet dikes occur; these range in composition from pure carbonatites to kimberlite and melilite basalts.

Radial dikes of much the same variety in composition are known as far as 25 km from the sövite center.

Over the years since the first serious geological report on the district (Högbom, 1895) was published, ideas concerning the genesis and manner of emplacement of the carbonatites and the alkaline rocks, especially of the former, have undergone appreciable change. Högbom considered the carbonate rocks in the area to have been segragations of limestone in nepheline-syenite magma; he did not think that they could be metamorphosed sedimentary limestones. Daly (1914) and Shand considered the areas to be a primary example of the syntexis of limestone with the carbonate bodies being relics from the assimilation process. Högbom later accepted Daly's idea to the extent that anatexis of limestone at some unknown depth may have added calcium carbonate to the nepheline-syenite magma; in the late stages of the crystallization of the $CaCO_3$-enriched magma, an immiscible $CaCO_3$ phase developed that crystallized as the sövite masses in the Alnö district. In 1922, however, Geijer pointed out that the limestones could not have been assimilated by the rising magma for, if that had happened, the gas necessary to produce the diatremes that are characteristic of this (and other) carbonatite districts would not have accumulated in sufficient quantity. Geijer also stressed the difficulty of a diatreme funnel having been blown out through kilometers of overlying silicic rocks and only limestone blocks falling back into the volcanic pipe. Nor, in this region that consists almost entirely of granitic rocks, could the rising magma have been expected to have found enough limestones to produce the sövites by any process. The problems in interpreting the genesis of the sövite was, of course, complicated at Alnö by the sövite showing through the glacial overburden in only a minor number of isolated patches. Until the rock surface was exposed to a major degree during road building between the two world wars, the fact that the sövite was present in a big central core and in the brecciated rocks of the cone sheets and the radial dikes could not have been appreciated. These relationships were understood only after the thorough and intensive field work carried out by von Eckermann.

The Alnö complex is a crudely lenticular diatreme (a breccia-filled pipe formed by gaseous explosions) about 7.5 km long in the east-of-north trending direction and 3.5 to 4.5 km wide at 90° to that direction. Numerous fractures, of both concentric and radial types were developed more or less symmetrically in relation to the complex center. Into these were introduced a wide variety of rocks, including carbonates (ranging from almost pure calcite to dolomitic and ankeritic varieties and containing many mineral species in trace to major proportions), kimberlites, carbonatic kimberlites, foyaites, phonolites, alnöites, tinguaites, and many other kinds of silica-low rocks.

Outside the core, much sövite is present in anastomosing, mostly irregularly concentric dikes that differ widely in shape, length and size; some small number of the sövite dikes appear to be radial. The structures that are filled by the dikes appear to have been developed as the first step in the emplacement of the carbonatite-syenite complex as a thermodynamic shattering of the surrounding country rocks. These are known as the breccia sövite dikes, not because the sövite is brecciated but because it cements the fragments of country rock produced in the thermodynamic fracturing of the migmatite.

The carbonate of the core of the complex appears to be quite pure sövites (1.63 to 1.8 per cent of $MgCO_3$ in solid solution in the 70 to 90 per cent $CaCO_3$); also included are 1 to 13 per cent apatite, 1 to 8 per cent biotite or manganophyllite, plus pervoskite (with 0.04 per cent U_3O_8, 3.00 per cent ThO_2), dysanalyte (a Cb-perovskite with 0.01 per cent U_3O_8 and 0.08 per cent ThO_2), and pyrochlore with minor U_3O_8 and up to 1.3 per cent ThO_2.

Where fragments of the country rock dropped into the carbonatic magma that filled the dikes, these commonly were changed to pyroxenitic fenites.

The core sövites have cataclastic strain features, showing that stress was continued into the fracture-filling stage. Marginally to these fenitized fragments, spherulitic serpentine is typically present. Locally, masses of sövite pegmatite with calcite crystals as long as 1 to 2 dm occur in the core. Although all of the five types of pegmatite recognized by von Eckermann are calcite-rich, they are quite varied in composition and are: (1) calcitic-juvite - cancrinitic juvite, (2) sövite - melteigite-sövite - malignite sövite, (3) sövite - pyritic-sövite, (4) sövite - baritic-sövite; and (5) fluoritic-sövite - sövitic fluorite.

Cone-sheet sövite dikes (more about which later) also occur outside the alkalic complex and are in two sets: (1) one set converging at a depth of 2000 km below the present surface (mainly alvikites - fine-grained sövites) and (2) the second converging at a depth of about 7000 m below the surface (mainly beforsites, i.e. rauhaugites). Also, von Eckermann recognizes (1966) a minor zone of sövite formation, about 3500 m below the present surface; what type of sövite was produced here is uncertain from von Eckermann's text.

In the cone sheet dikes within the complex, von Eckermann finds the sövites to be more varied in composition and divides them into four classes: (1) sövites with a maximum of 4 per cent apatite and 4 per cent biotite; (2) biotite sövites with 4 to 15 per cent biotite and 0 to 10 per cent apatite; (3) apatite sövites with 4 to 15 per cent apatite and 0 to 10 per cent biotite; and (4) pyroxene sövites with 4 to 20 per cent pyroxene and 0 to 12 per cent apatite; usually no biotite. In these sövites, the calcite content ranges between 40 and 96 per cent. These dikes also differ along strike; biotite-sövite dikes grade into pyroxene sövite or, rarely, into apatite-pyroxenite. In the wall rocks of the larger dikes, appreciable amounts of apatite have been developed as far as 1 dm from the dike wall. Accessory minerals are quite varied in species and content. Excess CO_2, over that required by calcite, probably is in that gas enclosed in fluid inclusions in the calcite.

In the purest sövites, 51.6 per cent run 80 to 96 per cent calcite; 24.1 per cent of the sövites have 70 to 80 per cent calcite, 14.8 of the sövites contain 60 to 70 per cent calcite, 4.9 per cent have 50 to 60 per cent calcite, and 4.6 per cent have 40 to 50 per cent calcite.

The narrower sövite-breccia dikes have the most distinct and well-defined contacts; in wider dikes, the replacement effect becomes more noticable and the contacts become most diffuse. In the widest dikes and those nearest the complex center, widths of the diffuse contacts seldom are above 0.5 m and usually are from 0.2 to 0.3 m. In these contact zones, fluorite normally is quite abundant and is vividly violet or purple. The strike of these dikes in brecciated rocks always is concentric, although the pitch differs with the distance from the complex center and with the cone-sheet or ring-dike character of the fracture. In the southern part of the district, cone-sheet fractures in the outer parts of the complex are well separated from the ring dikes; these fractures seem to merge together on the eastern and western sides; the northern side is, of course, hidden under the sea.

In the fine-grained sövites outside the core (sövites that came from about 4000 m below the surface at the time of sövite emplacement) almost universally are in cone-sheet fractures; only exceptionally are they in radical breaks. The calcite content ranges from 21 to 67 per cent, and they contain from 2 to 16 per cent $MgCO_3$ in solid solution in the $CaCO_3$. Some of these fine-grained sövites actually grade into the various silicate dike rocks, and, in these, the $CaCO_3$ content is less than 20 per cent. Such transitional types are produced as calcitic orthosite, calcitic foyaite, calcitic analcite tinguáite, calcitic alnöite, calcitic kimberlite, and calcitic ouachitite. Also in these fine-grained sövites occur such minerals as biotite, orthoclase, quartz, melanite, chlorite, olivine (largely serpentinized), pyroxene, melilite, apatite, nepheline, barkevikite, riebeckite, wollastonite, with accessory magnetite, ilmenite, pyrite, dahllite, corundum, anatase, fluorite, and perovskite. Some of these dike assemblages contain excess CO_2, probably in fluid inclusions. These dikes may be bordered by zones of fenitization.

In the rauhaugite dikes outside the complex, the $CaCO_3$-rich minerals also contain dolomite, ankerite, and/or siderite in solid solution in the calcite. The farther these rauhaugite dikes are from the sövite center, the higher they are in $MgCO_3$. Carbonate ranges in these dikes from 18 to 80 per cent, and they contain a wide variety of other minerals such as microcline, orthoclase, quartz, biotite, melilite, olivine, riebeckite, plus a considerable list of accessory minerals. Here also free CO_2 occurs in the fluid inclusions.

Three bodies of baritic carbonate are known in the district, two within the complex and one outside it. The cores of these bodies are quite pure barite, though some fluorite may be present in them. The wall zones (1 dm thick) contain 40 per cent barite, 40 per cent calcite, and 20 per cent highly visible fluorite.

The discussion thus far might give the impression that the Alnö complex contained nothing but masses of sövite and dikes containing various varieties of sövite. This, of course, is far from the truth. Actually, the area into which the carbonatites (and their associated low-silica rocks) were forcefully intruded originally consisted of Archean migmatites and granite gneisses. Between major masses of migmatite and the complex lie bands of thermodynamically shocked rocks that have not been appreciably altered by the various processes connected with the emplacement of the sövites and their associated syenitic fennites, rheomorphic fennites and truly magmatic rocks (in that they appear to have crystallized from molten silicate magma, even though the silica content of these magmas was low).

Thus, von Eckermann (1966) divides the complex and its immediate environs into 10 zones, number from 0 to 9, the zone given the number (9) being the sövite core. The zones as defined by von Eckermann are as follows:

(0) unaltered wall rock - Archean granite gneiss and migmatite

(1) the thermal shock zone in which the wall rock is well jointed and contains much granulated and optically strained quartz

(2) quartz-syenite fenite

(3) syenitic fenite

(4) Nepheline-bearing syenitic fenite (alkaline fenite); this zone contains the outermost nepheline-bearing rock

In zones (1) through (4) the rocks still show the original strikes and dips of the rock before it was "fenitized" so that the change in its composition appears to have been metasomatically induced by fluids given off from the various masses of molten material (mainly kimberlitic-carbonatitic magma or magmas) that invaded the district; these preserved strikes and dips appear to rule out more than the most minimal rheomorphism, if that.

(5) nepheline syenitic fenite of which both leuco- and mela-types are recognized; in this zone, the contrasting layers of the migmatite are preserved, but the strike of these rocks has changed toward, or to, being concentric to the invading sövite; the dip is converted to, or toward, a cone-shaped relationship to the source of the sövite magma; zone (5) also is thought by von Eckermann to have formed metamorphically in situ, but must have been plastic enough to have moved in relation to its previous dip and strike.

(6) mela- and leuco-nepheline, highly sodic rocks, rich in melanite garnet - $Ca_3(Fe^3,Ti^3)_2(SiO_4)_3$ - and locally containing appreciable wollastonite; because the original attitude has disappeared with the strike now being concentric to the core and the dip cone-sheet-related; these rocks von Eckermann considers to be rheomorphic fenites

Zones (5) and (6) may be absent; thus zone (4) in such cases will be in a direct contact with zone (7).

(7) melanocratic igneous rocks that are rich in pyroxene and melanite

(8) leuococratic igneous rocks that are rich in nepheline and anorthoclase

(9) carbonatites - sövites, fine-grained sövites (alvikites), and rauhaugites (beforsites)

Von Eckermann refers to the rocks of zones (7) through (9) as "homogenous alkaline rocks" of the "central alkaline area".

Von Eckermann shows the relationships of the major bodies of the rocks of these nine zones on his detailed map (1948, Plate 60). Except for the sövites which occur as breccia dike fillings as well as in the main sövite core, the remaining rocks of the various zones are more or less banded around the core. Certainly the thermodynamic shock zone surrounds the syenitic-carbonatite complex completely, separating it from the migmatite outside it. Quartz-syenite fenite also forms a complete band between the thermodynamic shock zone and the syenitic fenite. After that, however, the relationships are more complex. Within the syenitic fenite, which covers at least half of the area of the complex on von Eckermann's, are isolated, but small, bodies of migmatite that, in turn, are separated from the syenitic fenite by bands of quartz syenitic fenite. Other bodies of syenitic fenite within the quartz syenitic fenite do not have migmatite cores. On his map, von Eckermann combines zones (4), (5), and (6) into nepheline-syenitic fenite, and then divides the combined zones into melanocratic and leucocratic. This group of zones forms the second largest outcrop area on von Eckermann's map with the leucocratic variety being more abundant than the melanocratic. The sövite dikes are much more common in the rocks of zones (2) and (3) than they are in the remaining outcrop area, although one area of melteigite-malignite-jacupirangite near the southern boundary of the complex has nearly half its outcrop area as sövite dikes.

Von Eckermann divides the rocks of zones (7) and (8) into two varieties: (1) those that contain the rocks of the series alkorthosite-juvite-malignite-jacupirangite and (2) those that contain urtite-ijolite-melteigite-jacupirangite. Although von Eckermann says that the rocks of variety (1) are much more abundant in the complex than those of variety (2), their outcrop areas shown on his map do not seem greatly different in total size.

Except for the small number of areas where breccia dikes coalesce, no bodies of sövite [zone (9)] exist outside the core.

Just outside the western edge of the complex, near the village of Båränge, is a small body of rocks of zones (3) and (4) and dikes of zone (9) rocks; at least three of these dikes extend into the migmatite. Nearly as much sövite in coalesced dikes outcrops in this area as does the syenitic fenite of zone (3); zone (2) rock is minor in amount. No rocks from zones (4) through (8) are shown on the enlarged map of the Båräng area.

Von Eckermann (1966) describes the processes that resulted in the development of these various and widely different rock types as follows:

(1) thermodynamic shattering of the microfabric of the surrounding country rock, presumably on the (initial?) development of the diatreme

(2) introduction of CO_2, F, and H_2O to produce carbonate and fluorite pseudomorphs after the feldspars, hydrating the sillimanite (in the migmatite) and the alkali feldspar components and raising the oxidation ratio (whatever that may mean)

(3) introduction of Ca and some P and Ti in exchange for Si and some Na, replacement of biotite and hydrated minerals by aegirine augite and soda-orthoclase

(4) dehydration and further loss of Si resulting in the crystallization of potash-rich nepheline

(5) release of leucocratic components of the fenite leaving a pyroxenitic rest-fenite rich in melanite formed by the break-up of the aegirine molecule

(6) accumulation of rheomorphic fenite-magmas subjected to agpaitic crystal differentiation [agpaitic = crystallization in the presence of an excess of alkali (especially sodium) so that the amount of aluminum oxide is insufficient for the formation of aluminum silicates leading to the formation of juvites, foyaites, urtites, ijolites, melteigites, malignites, and jacupirangites].

Through these steps, von Eckermann expects that the migmatitic-gneissic rocks would have been converted to the rather complex series already outlined. Of the 9 zones into which von Eckermann divides the fenitized rocks, using "fenitized" in the broadest sense possible, only the rocks of zone (9) which includes all of the various types of sövites, were not, in his opinion, developed by the fenitization process. But, the materials needed to convert the country rocks to the rocks of zones (1) to (8) were added from the sövite magma and those that had to have been removed in the various aspects of fenitization (in the broad sense) were carried away in the fluids produced during the crystallization of the sövite magma, either in its core mass or in the fractures into which it was intruded. The process of fenitization further was complicated by von Eckermann's belief [based largely on the geometry of the cone-sheet and radial fractures produced in zone (0)] that four sövite magma chambers were formed during the fenitization process. The highest, and most important in quantity of these was the one centered some 2500 to 3000 m below the surface as it existed at the time of sövite production, next below this was the minor one at 4000 m below that surface, the principal carbonatite product of which was fine-grained sövite (alvikite), next below was an even smaller volume of molten sövite, so small that its effect in the fenitization process must have been neglible, and finally a moderately large volume of dolomitic-ankeritic-sideritic carbonatite (rauhaugite in the sense used in the Fen area of Norway or beforsite in the non-preferred term of von Eckermann). Presumably, the three zones of fenitization below the first and major one are only to a minor extent, if at all, responsible for fenitization effects now visible on the surface. The molten sövite probably existed in that state for a considerable period of time, a period long enough for the internal gas pressure of the sövite magma to build up sufficiently to exceed the lithostatic pressure provided by at least 2000 m of migmatite-gneiss and Jotnian sediments. Appreciable time, therefore, existed for the accumulation of a carbonate-rich liquid phase (not a gas phase because of the high confining pressure exerted by the overlying rocks) that could carry materials through the fractures produced during the thermo-dynamic shock phase to produce the various steps in the fenitization process. The upward journey of these liquids was followed shortly, if not concurrently, by the intrusion of actual sövite magma. This magma was quite different in composition, depending on at what stage of the crystallization of the carbonatite magma it was expelled from its particular magma chamber. Its composition also depended on its relationship to the parent kimberlite magma that filled the bulk of the volcanic channelway, at various elevations along which the four varieties of sövite magma were immiscibly separated from the dominant kimberlite magma.

Certainly the different centers, to which the cone sheets containing the rauhaugite carbonatite are related, must have exploded at different times than those that produced the various varieties of sövite because the pressure needed for the explosion at 9000 m must have been far greater than that needed at 2500 to 3000 m. This is true even granted that the upper part of the pipe (the initial diatreme) had been fractured severely by the explosion centered at the higher level.

As calcium and the other carbonates in the sövite began to come out of solution (in the molten state) in the kimberlite magma, the amount of the kimberlitic material that any particular increment of sövite contained would depend on how complete was the immiscibility produced before the carbonatite magma was driven out into the cone-sheet and radial fractures produced by the various explosions generated by the concentration of CO_2 (under high pressure) that came out of solutions in the carbonatite magma. The earlier the sövite was expelled from its kimberlite parent, the higher would be its concentration in the silicate minerals that the kimberlite would contribute in the earlier stages of immiscibility. Only if the two phases (kimberlite magma and carbonatite magma) co-existed in contact with each other over a long period (geologically) would the immiscibility be sufficient to provide a relatively clean separation between the carbonate and silicate phases. Thus, it is not surprising that essentially all gradation between pure sövite and highly contaminated carbonate magma would have been driven out of the carbonatite magma chamber and introduced into, and solidified in, the cone-sheet and radial dikes.

At essentially the same time (or times) that carbonatite-rich material was being extruded from the volcanic chimney, kimberlite magma also was driven out of the kimberlite magma column. Of the 274 dikes categorized by von Eckermann (1958) 184 are of various degrees and types of carbonate composition, whereas 54 (30 per cent of 184) are kimberlite. This suggests that the gases needed to generate explosive extrusion were more abundant in the carbonatite portions of the magma column in the volcanic chimney. The probably greater viscosity of the kimberlite magma in relation to that of the carbonatitic material also is likely to have reduced the amount of kimberlite driven out of the magma chamber. Incidentally, 35 per cent of the carbonatite dikes came from the lower of the explosion centers whereas the kimberlite dikes were divided essentially evenly between the two levels of explosion. Of the 274 total dikes examined by von Eckermann, 35 per cent were radial and, of course, 65 per cent cone-sheet type. Interestingly, despite the frequent references to alnöite dikes in the Alnö complex area, only 3 of the 274 studied dikes are composed of that rock.

Although the number of sövite filled dikes is much higher than that of those containing kimberlite and its minor number of alnöitic variants, these kimberlite magmas driven out of the volcanic magma chamber are of considerable interest in attempting to elucidate the manner of genesis of the Alnö carbonatite. Actually what von Eckermann describes as "kimberlites" are better designated as olivine-melilite peridotites, but there is no question but that they grade into alnöites in a few places. But his kimberlites are not in the original definition of that term, that is, porphyrytic peridotites in which olivine and, less commonly, phlogopite are present as phenocrysts; the olivines normally are serpentinized or carbonatized, and the phlogopites generally as chloritized; the ground mass is fine-grained calcite and second-generation olivine; geikielite ($MgTiO_3$) and Cr-pyrope may be present as phenocrysts and ilmenite, serpentinite, chlorite, perovskite, and magnetite as accessories. With phlogopite, kimberlites are called "lamprophyric kimberlites", without phlogopite they are termed "basaltic kimberlites", a poor term because they do not contain feldspar (Heinrich, 1966). At Alnö, the kimberlites are later than the core carbonatite and than the younger carbonatite contained in ring and cone dikes. Von Eckermann suggests that the Alnö kimberlites are melilite basalts from the basaltic portion of the crust; in this he probably is wrong - they are much more likely true peridotites from the mantle. Wylie has shown that an alkalic peridotitic magma can yield a lime-rich carbonatitic residue by fractional crystallization, but it appears that it is more probable that the separation is caused by liquid immiscibility.

Von Eckermann is convinced that both of the two low-silica rock series (the juvite and the ijolite) end in jacupirangite, the only difference being that the jacupirangite associated with the juvite series is biotite-jacupirangite and the connectionwith the ijolite series is amphibole jacupirangite. A major difference between the two is that the most leucocratic member of the juvite series is orthosite (a syenite made up almost entirely of orthoclase) whereas the most leucocratic member of the ijolite series is urtite (a nepheline-rich rock in which the mafic minerals amount to no more than 30 per cent and usually are much less). The difference between juvite and ijolite is that the former is a nepheline syenite in which all or most of the feldspar is orthoclase and the latter is a nepheline-rich rock (40 to 70 per cent nepheline with 30 to 60 per cent mafic minerals, mainly clinopyroxene). In malignite, the nepheline content is normally appreciably more than 5 per cent and equal amounts of pyroxene and potash feldspar are present whereas in melteigite, nepheline is present as from 10 to 40 per cent of the rock and the remainder is mafic, mainly pyroxene. Finally, the two series end in jacupirangite in which feldspar is lacking - the rock is essentially a nepheline-bearing pyroxenite. Thus, the difference between the two series is that the juvite series (through malignite) contains potash feldspar in appreciable amounts whereas the ijolite series has abundant to moderate amounts of nepheline but no feldspar. Thus, if these rocks are crystallized from magmas anatectically derived from various types of fenite produced from migmatite, the juvite magma parent still contained enough silica to produce, on crystallization, feldspar and the ijolite magma parent did not. Which magma would be developed by anatexis would depend on how far the feldspar content of the parent migmatite had been reduced during fenitization, and not a great deal of difference in this content would send the magma along one line of descent or the other. Certainly, von Eckermann does not believe that either series was produced by fractional crystallization of kimberlite. Of the ten major areas of the two types of anatectic rocks shown on von Eckermann's major map in his 1948 report, three appear to contain rocks of the juvite series only, three appear to include nothing but rocks of the ijolite series, and four contain both. Of the four in which there are rocks of both series [including one bordering directly on the sövite core on Långörsholmen (island)], it appears that the rocks of the more silicic types [urtite, juvite, ijolite, and foyaite (another nepheline syenite with dominant potash feldspar)] are enclosed by the more mafic rocks (melteigite, malignite, and jacupirangite). So mixed up are these rocks in any given outcrop that no more definite separation can be made in mapping. This enclosure of the more silicic rocks by the more mafic ones, however, is exactly what would be expected to happen in the fractional crystallization of these antectic magmas, particularly when it appears probable that no one of these magmas ever had time to be well homogenized. Certainly, the composition of the anatectically melted fenite was such, however, that the first mineral to crystallize would have been either potash feldspar or nepheline (depending on the silica and potassium and sodium contents of the melt in question).

Von Eckermann also believes that much of the mafic minerals in the jacupirangites came from fenite residues without ever having been completely dissolved in the anatectic magma from which they eventually precipitated, at least where they are directly associated with sövite. He suggests that these jacupirangites are simply the femic remains of a fenite where all the leucocratic material had been "sucked out" by (dissolved in) the engulfing fenite. Perhaps this explains some of the jacupirangites, but it seems more probable that those directly connected with rocks of the potash-rich juvite series or the soda-rich ijolite series were initially as molten as the rest of the anatectic magma and were formed from the last minerals to be crystallized from it.

That the juvite and ijolite series of rocks were, in large part at least, crystallized before the last of the sövite had been introduced into the syenitic-carbonatitic complex is shown, for example, by the numerous sövite dikes that cut through rocks of these two series, such as the long sövite dike that cuts cross Långörsholmen from almost north to south or the huge swarm of sövite dikes that cut through the juvite-ijolite complex that centers around Stavätt on the south side of the complex. Alteration of the rocks through with the sövite dikes pass is impressive with the wall rocks of the dikes being considerably to greatly enriched in apatite. Locally, the dikes may gradually change from sövite to biotite-sövite and finally to pyroxene-sövite or from sövite to apatite-sövite to apatite-pyroxentite if enough biotite originally was present. I am sure most of the "contaminants" in the sövite were contained in it when it left its magma explosion chamber and were not picked up to nearly as great an extent by reaction with the wall rocks. The principal "contaminant" minerals are such species as would much more logically have been transferred from the kimberlite magma to the carbonate immiscible fraction when the two separated from each other rather than minerals that could have been extracted from the wall rock. Of course, for each grain of apatite transferred from a molten sövite dike-filling material an equal volume of some mineral (or minerals) from the fenite must have been taken into the sövite, but their effect on the sövite appears to have been far less drastic than the sövite-generated alterations were on the wall rocks (fenites or members of the juvite or ijolite series) through which they passed.

Such columbium as is present in the sövite should be classed as Magmatic-Fla, that is, early crystallization, early separation (from the sövite magma). Obviously, the sövite is not now an ore of columbium, but it may be in the future. The barite at Alnö is in late dikes that range in width between 1 and 3 m. In the dike centers, the barite is very pure, 96.6 to 99.9 $BaSO_4$; the small remainder is fluorite. Both sides of the dikes are bordered by 1 dm-wide margins that contain 40 per cent barite, 40 per cent calcite, and 20 per cent fluorite. In different dikes the fluorite is of different colors, brownish yellow and pale green at Pottäng and honey yellow or strongly violet at Hartung. Von Eckermann says that the barite bodies should be designated as veins and not dikes which would seem to indicate that he thought them to have been deposited from solutions and not from a variant of the sövite magma. Accepting this, the barite veins probably should be classified as Mesothermal, but the problem cannot be considered as being closed.

The alkaline complex of Alnö Island almost certainly was introduced into the area during Jotnian (probably late Jotnian) time so should be categorized as Late Precambrian or even Eo-Cambrian.

SMÅLANDS TABERG, JÖNKÖPING

Late Precambrian	Iron as Magnetite Titanium as Ilmenite, Ulvite Vanadium in Magnetite	Magmatic-3b

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Notes

The vanadium-bearing magnetite-ilmenite deposit of Smålands Taberg is located at 57°42'N, 14°05'E and is the most southerly of the major metallic deposits of Sweden. Because of the close association of magnetite and ilmenite in the same mineral (designated as titano-magnetite) at Smålands Taberg, it was found impossible to separate them mechanically. Despite this condition, it is reported (Hjelmqvist, 1950) that mining was begun at Smålands Taberg in the Middle Ages. By the end of the 16th century mining had ceased, but a visit by the then King, Charles IX, in 1604 resulted in reopening the mine. With the help of the visit of another King, Gustavus II Adolphus, by 1639 five smelters and iron works were in operation there, and, by the beginning of the 18th century, 14 blast furnaces were working. In the last half of the 18th century, the amount of ore was about 3000 tons per year. By the middle of the 19th century production was about 9000 tons annually, but operations slowed down after 1877 and stopped altogether by 1895 with the last blast furnace being drawn down in 1888. Hjelmsqvist suggests that this was because of the low-grade of the ore about 31 per cent iron, part of which was present in silicates and not, therefore, available for recovery in smelting.

With all these handicaps, made even worse by the lack of locally available limestone, the Taberg iron was considered to be of good quality. The iron produced from the Taberg ore was believed to be particularly useful in the making of iron wire because the iron was reported to be most homogeneous and lacking in sulfide impurities.

In 1939, with the beginning of World War II and the Germans wanting iron ore of almost any type for their war industries, mining began again at Taberg. Output increased to far above that of the older times, reaching more than 200,000 metric tons in 1943. This large increase was due more to the vanadium that could be leached from the ore (it is reported by Hjelmqvist to run 0.86 wt per cent V_2O_3 from titanomagnetite in a coarse grained dike) than from the value of this iron ore in which titanium formed such a large percentage of the ore. The only possible explanation for the continued use of the ore (although in only small quantities up to 1939) is that the titanium was removed in the slag. Certainly, metallic iron with the titanium in the ore retained in the metal would be essentially useless industrially, I also can testify from personal experience that the titanomagnetite-olivine rock is the hardest rock I ever have attempted to break; I easily broke an Estwing hammer on it, and I can see why an effort to mine underground was quickly given up (after sinking the deeper of two shafts only 7 m below the surface).

Incidentally, the discovery of vanadium as an element was made by a Swedish geologist-chemist N. G. Sefström in 1830 from rod-iron made from Taberg ore.

The first geologic description of any importance was published by D. Tilas in 1757; he distinguished three types of ore: (1) rich iron-ore veins, normally less than 15 cm wide; (2) ordinary ore; and (3) spotted ore, the spots being white feldspar crystals. In 1789 a pupil of Werner's examined the ore body; he called ore ferriferous trap (quite correct) but believed, true to what he had been taught, that it was a sedimentary bed intercalated in the gneiss (by which the ore largely is surrounded). The first modern work was by Sjögren in 1876; in addition to aereal geology, he also examined the rock and ore microscopically and designated the ore as magnetite-olivinite. He thought the Taberg ore was an eruptive rock intruded into the surrounding gabbro (hyperite) which had been introduced into the gneiss before the ore. In 1881, Törnebohm differed with Sjögren; he considered that the magnetite-olivinite was intercalated in the hyperite and what he calls hyperite-diorite and was genetically directly connected with the other two and formed successive transitions into them. He said that the ore was a variety of hyperite in which the heaviest minerals (magnetite and olivine) had accumulated by gravity, displacing the lighter ones. Törnebohm considered that the Taberg ore bore a great similarity to that of Iron Mine Hill in Rhode Island, quoting Wadsworth's studies of 1881. Sjögren replied that he still thought the magnetite-olivinite showed no real transition with the hyperite. Hjelmqvist (1948) thinks that both were right, Sjögren in holding that the ore actually cuts the hyperite and Törnebohm in that the two rocks are genetically related.

The hyperite body at Taberg, in which the ore mass is contained, is bounded by gneiss and gneiss-granite and has its long dimension of about 4 km in a north-northwest-south-southeast direction. It is one of a group of similar bodies, which usually are dike-like in form and cut through the Archean rocks that are in the south-southwest part of the county of Jönköping. The northern portion of the hyperite mass at Taberg is composed almost entirely of magnetite-olivinite; this body is cut off to the southeast by a fault that runs NNE-SSW and has a pronounced topographic expression developed on a major zone of crushed rock produced by the fault process. Toward the west, outcrops are quite common and show that the magnetite-olivinite is, in part, cut off against hyperite that, in turn, passes gradually into amphibolite. These two rocks form a belt of varied width lying between the magnetite-olivinite ore and the adjacent gneiss-granite. Locally, the magnetite-olivinite abuts directly against the granite gneiss; this magnetite-olivinite is here somewhat altered.

The gneiss contains intercalations of layers or sheets of an older amphibolite that is similar to the amphibolite into which some of the hyperite has been converted. This older amphibolite is difficult to distinguish from that formed from the hyperite; not even determinations of minor element contents serve to tell one amphibolite from the other.

In both the hyperite and the gneiss-granite that surrounds it are dikes and lenses of (granite) pegmatite; no pegmatite has been found in the magnetite-olivinite.

Beyond the east border of the hyperite-ore body is a covering of thick glacial debris; along the southeast side of the hyperite-ore mass, the mafic rock is bordered by the NNE-SSW fault; on the southeast side of the fault is a considerable area of hyperite amphibolite that, in turn, stops against gneiss-granite to the northeast and, presumably, in other directions that are covered by glacial debris. On the east side, under the debris, may be hyperite or hyperite-amphibolite between the ore and the gneiss-granite; Hjelmqvist thinks that this is probable.

On the west side of the ore-mountain, the strike of the gneiss-granite is normally about NW-SE or NNW-SSE, and the dip is toward the southwest. The dip is quite steep on the west side of the mountain, normally being between 60° and 80°; it may be as low as 40°.

Hjelmqvist mentions that, on the western slope of the Taberg Hill, both the gneiss and the older amphibolite are folded; these folds turn off to the NNE, and the folded rocks dip ESE. Here the hyperite cuts the gneiss; in all other places, the gneiss-hyperite contact is conformable.

To the east of the iron-rich hill, Hjelmqvist has found only one outcrop on which strike can be measure, and this is at some distance from the ore. This place is not located on the map, but he reports the dip to be about 30°WSW; from this it appears that the strike direction should be NNW. Törnebohm reports, however, that, farther west, he measured a dip of 60°W. On the northeastern slope of the Taberg Hill, the strike bends to the northwest and west, and still farther west, the gneiss-granite strikes N60°E with an almost vertical dip.

The large tabular plagioclase crystals in the magnetite-olivinite in many places have a parallel arrangement, suggesting a flow structure in the ore; other areas of ore, however, exhibit far less parallism of the feldspars. Because of the high magnetic attraction of the ore for the compass needle, the strike of these tabular crystals Hjelmqvist could not determine accurately. It appears, however, that the tabular crystals (Hjelmqvist says "tablets") dip from the margins toward the center of the ore mass at angles between 40° and 50°, but he believes that he has too few determinations on which to base any definite conclusions as to in flow directions of the ore magma.

The relations of magnetite-olivinite to the hyperite, most importantly a strong bulge in the western margin of the iron ore into the hyperite that nearly reaches the gneiss-granite, indicate that the ore magma was intruded into the hyperite after that rock was essentially, if not entirely, solidified. Hjelmqvist could find no outcrop that showed any gradual transition between ore and hyperite; the contact between the two is sharp, and it can be followed easily with the naked eye. Definitely sharp fragments of hyperite can be found in the ore; the fragments range from a few cm in maximum dimension to several meters, and no chilled margins are provided in the ore enclosing the hyperite fragments. This would seem to inidicate that the intrusion of the ore occurred while the solidified hyperite was still warm if not hot. The close association in time between the two rock types is confirmed by the distinct conformity of the mineralogy of the one with the other, only the proportions are different.

Toward the western boundary of the hyperite with the gneiss-granite, the hyperite is much finer-grained near the contact than farther away from it; the relationship still obtains when the hyperite has been converted to amphibolite. Some secondary schistosity has been produced in the hyperite in its contact zone with the gneiss-granite, and there the hyperite has been, in part, incorporated into the older rocks. No such schistosity was developed in the magnetite-olivinite.

The ore-rock, themagnetite-olivinite, typically is a fine-grained, massive, quite homogeneous rock that is almost black in color; the grain give the appearance of being small metallic dots in a brown-black ground mass. The essential minerals are only titano-magnetite and olivine with accessory plagioclase (labradorite) and some secondary amphibole. The better ore is, of course, that lower in plagioclase, and it is most common on the southern part of the intrusive body, with much smaller patches of it being located mainly in a belt of isolated bodies, the long dimensions of which are about north-south and which are arranged in a parallel line that runs from west to east. The ore-rock is essentially a peridotite, although its content of titano-magnetite is far higher than is typical of that rock type. The feldspar-rich variety of the ore-rock is quite similar to that of the cumberlandite (Wadsworth, 1881) of the Iron Mine Hill in Rhode Island.

The greater bulk of the peridotite, however, is irregularly sprinkled with tabular feldspar crystals that are macroscopically brown; all transitions exist between feldspar-free ore-rock and those rich in the plagioclase. The feldspars in many places are oriented in parallel and range in length from a small fraction of a millimeter to as much as 30 mm or more. The feldspars weather to a white color and stand out prominently in the weathered rock. Locally, the feldspars are arranged in star-shaped groups or less regular patches.

Apatite is present in locally isolated prisms but much of the ore-rock is completely lacking in this mineral.

The titanomagnetite is, under the microscope, seen to be made up of magnetite, ilmenite, and a spinel. Sulfides, pyrrhotite, pentlandite, chalcopyrite, and pyrite, are present but quantitatively are insignificant. Hjelmqvist gives three analyses of the ore-rock as follows; the per cents are based on volume, not weight:

Olivine	58.5	45.7	43.0
Titanomagnetite	34.9	28.7	25.7
Labradorite	4.5	9.9	25.2
Amphibole	2.1	15.7	6.0
Apatite	-	-	0.1

The ratio of olivine to titanomagnetite can be seen to be quite constant, from left to right, these ratios are 1.67, 1.59, and 1.67.

In general, the olivine is prismatic in form with rounded corners. The grains normally are in closely packed aggregates scattered among the grains of titano-magnetite; the grain size is between 0.2 and 0.5 mm, but some grains may be from 1 to 2 mm in length. Unaltered olivine is brown but serpentinization develops initially along irregular fissures with the entire crystal turning colorless even where the grain is only partially serpentinized. This colorless olivine is not a common constituent in the ore-rock. The analysis of typical brown olivine in weight per cent is as follows:

SiO_2	36.20
TiO_2	0.49
Al_2O_3	0.91
Fe_2O_3	1.89
FeO	29.21
MnO	0.28
MgO	29.60
CaO	0.61
H_2O (above 105°C)	0.57
H_2O (below 105°C)	0.04
Total	99.80

The titanomagnetite is in anhedral crystals in spaces among olivine crystals; where the olivine crystals are closely packed, the Ti-magnetite may be in narrow veinlets that follow the contacts of olivine grains. Where olivine crystals are sparse, idiomorphic olivine grains may be completely surround by Ti-magnetite. A few isolated octahedral crystals of magnetite with rounded corners have been observed. Certainly the Ti-magnetite is later than the olivine as its position as fillings around olivine crystals shows. The magnetite grains are about the same size as those of olivine.

In polished section, the titanomagnetite grains are seen to be composed of three minerals; the bulk of the mineral is magnetite, clear green spinels (color seen only under high magnification) in fine lamellar intergrowths are normally oriented parallel to the cubic faces of the magnetite. The spinel, as demonstrated by X-ray studies is pleonaste [$Mg,FeAl_2O_4$]. In addition to the spinel lamellae, some of that mineral occurs as drop-shaped or irregular intergrowths that are scattered through the magnetite. In places, the spinel lamillae in magnetite-grain margins are larger and more closely spaced than those toward the center. Grain lengths range from 0.1 to 0.2 mm; widths are measured in microns. Spinels within a single magnetite grain may differ in size by 20 times. In most magnetite, however, spinel is uniformly distributed throughout the grain.

The magnetite also contains small ilmenite lamellae that are arranged in an octahedral pattern; they are visible only at high magnification. Locally, the ilmenite may have a cubic pattern, parallel to the cubic faces of the magnetite. The length of the ilmenite lamellae ranges from 0.01 to 0.05 mm, being only 1 to 2 microns wide. Only very small and isolated grains of pyrrhotite and pyrite are seen in the magnetite. In a few places, the ore minerals form narrow veins that cut through the main mass of magnetite-olivine rock; such veins contain a few grains of olivine and are cut by fissures that include serpentine, epidote, and carbonate. The grains in these small veins are larger than those in the enclosing ore-rock; such grains are high in spinel, but they contain the usual amounts of ilmenite lamellae. Some small fissures that contain only ilmenite occur in vein-magnetite and some ilmenite-free magnetite is present in similar fissures, both of these probably are the result of exsolution processes in the titanomagnetite.

Some of the plagioclase in the magnetite-olivine rock anhedrally fills spaces between olivine and titanomagnetite; these plagioclases usually are free of inclusions. The distribution of plagioclases is erractic, and, in the ore rich in iron, they are essentially absent. The tabular feldspar crystals normally are 5 to 20 mm long and 0.2 to 1 mm wide, but much smaller ones are known. The composition of these labradorites ranges between 50 and 57 per cent anorthite with the average composition being $Ab_{47}An_{53}$. Some slight alteration of the plagioclases to micas has been noted. The dark color of the feldspars megascopically is probably due to their excessive thinness which allows the brown color of the magnetite-olivinite to show through. Between the plagioclases and the olivine and magnetite grains are amphibole reaction rims of varied width; larger plagioclases may be completely replaced by amphibole. The amphibole reaction rim is made up of needles of pale-green or colorless amphibole, but some reaction rims may be composed of layers of variously colored amphiboles. Around some magnetite, a thin rim of garnet may be found between the iron mineral and the amphibole. In rare instances, a thin rim of biotite is present around titanomagnetite.

Such apatite as is present may be prismatic, but occasionally it is anhedral to the neighboring olivines; thus, at least part of the apatite is contemporaneous with the olivine. The hyperite and the less closely associated anorthosite contain appreciably more apatite than does the magnetite-olivine rock.

The sequence of mineralization, where plagioclase phenocrysts are lacking, is: olivine, titanomagnetite, and finally plagioclase to fill the interstices among the other two minerals; at least part of the minor apatite crystallized with the olivine. Where labradorite phenocrysts are present, they are contained in an olivine-titanomagnetite ground mass that fills the wedge-shaped spaces among the plagioclases; in some places, this filling is solely titanomagnetite. Not only are the titanomagnetite crystals the youngest in the ore-rock, but they also are present as compact veins in the ore-rock.

Hjelmqvist gives several chemical analyses of magnetite-olivine of which the following three are given here:

SiO_2	25.30	20.99		21.84
TiO_2	6.89	8.03		5.04
Al_2O_3	7.15	4.92		7.65
Fe_2O_3	15.91	19.46		15.78
FeO	22.63	24.10		26.39
MnO	0.31	0.35		0.34
MgO	16.68	17.36		18.93
CaO	2.12	1.89		0.73
Na_2O	---	---		0.28
K_2O	---	---		0.18
V_2O_5	0.26	0.31	V_2O_3	0.24
P_2O_5	0.092	0.110		0.12
S	0.023	0.024		0.06
Cu	0.002	0.002		---
F	---	---		0.02
H_2O (loss on ign.)	2.633	2.454		2.30
Totals	100.00	100.00		99.90

An analysis given by Hjelmqvist for cumberlandite from Iron Mine Hill is not appreciably different from those for Smålands Taberg. A weight-per cent analysis of a magnetite-olivinite from Taberg gave the following result:

Labradorite ($Ab_{47}An_{53}$)	3.5
Olivine ($Fo_{67}Fa_{33}$)	53.0
Amphibole	3.0
Titanomagnetite ($Mt_{46}Il_{29}Sp_{25}$)	40.5

About 100 m north of the northern summit of Taberg a quite thick dike (about 0.5 dm) is composed of an unusually coarse-grained magnetite-olivinite rock; it intersects a normal magnetite-olivinite. The dike follows an irregular course with its contacts with the country rock being quite sharp. Hjelmqvist believes that this dike derived from a late fraction of the magnetite-olivinite magma that had a higher than average content of what he calls "mineralizers", probably mainly water; its mineral composition is much the same as that of the ore-rock that it cuts. In this dike, the titanomagnetite shows, when highly magnified, that the ore mineral between the spinel and ilmenite lamellae is made up of two minerals. Considerable detailed work has shown that the second of the two ground-mass minerals is not ülvite [Fe_2TiO_4]. The two ground-mass minerals, one light gray and the other dark, are closely associated together in a micropegmatite type pattern. A similar pattern is shown by Ramdohr (2d English translation of the 4th German edition, p. 919), but he identifies the dark mineral as ülvite which, in the Taberg specimen, Hjelmqvist concludes cannot be ülvite. At the Cornwall mine in Pennsylvania, two types of magnetite have been reported from the same specimen, when these are seen together (they cannot be identified if they are not together) one is bluish and the other brownish. Hickock [Econ. Geol., 1933, v. 28, p. 219-220], the last author to report on this phenomenon, could not explain why they were developed, and the texture they show in the Cornwall specimens is quite different from that of the Taberg ones. This only suggests that the two minerals at Taberg are both varieties of magnetite; much additional work would be needed really to discover what these two Taberg minerals are, but, from the chemical composition of the magnetite, it seems unlikely that they could be anything but two varieties of magnetite. Hjelmqvist reports that X-ray examination shows that both are spinels with one having an appreciably larger unit-cell dimension than the other [8.09kX as compared with 8.37kX, the latter certainly being magnetite]. The cell lengths make certain that neither of the two minerals is ilmenite.

The plagioclase in this coarse-grained titanomagnetite may be as small tabular crystals enclosed in the Ti-magnetite or as clusters of larger anhedral tabular crystals that fill in spaces among other minerals. No reaction rims are present between the plagioclase and the dark minerals. The plagioclase is labradorite with An between 56 and 59 per cent. Apatite is uncommon but resembles that in the ordinary titanomagnetite.

Locally, the magnetite-olivinite has been altered both in structure and in mineralogy with altered ore-rock forming quite narrow zones in the unaltered material. Hjelmqvist considers that, in many places, the altered zones are narrow fissure fillings; the stronger the tectonic effects, the wider these altered zones. These fissures of altered ore-rock may intersect each other and may be separated from the unaltered rock by serpentine and perhaps other minerals. These fissure fillings are confined to narrow crushed or foliated zones. Where the tectonic effects are strongest, foliated types of Ti-magnetite are broken into shreds.

Where olivine has been altered in these zones, it has been converted to serpentine, carbonate, and light green amphibole, plus chlorite where alteration has progressed farthest.

The hyperite, according to Rosenbusch, is a rock of gabbro- like texture, intermediate in composition between norite and gabbro. In the strict sense, Hjelmqvist suggests, the hyperite has the texture of a diabase (dolerite in the English usage). After examining about 100 thin sections of hyperite, Hjelmqvist found hypersthene (actually orthorhombic pyroxene) in only 43 per cent of them so hypersthene is not essential to a rock's being designated as hyperite.

Thus, Hjelmqvist summarizes his remarks by saying that hyperite has an ophitic or subophitic texture and that its main components are labradorite, clinopyroxene, frequently olivine, hypersthene may be present but more usually is lacking. Accessories are ilmenite, apatite, and magnetite. Typical Taberg hyperite is a medium-grained, blackish-brown rock with dark brown tabular crystals of labradorite (the brown color being partly due to tiny dark inclusions and partly to the thin nature of the plagioclases that permits the brown rock-color to show through. The mineralogical differences between magnetite-olivinite and hyperite is that the latter has an appreciably higher content of plagioclase, less olivine and titanomagnetite, and some pyroxene (lacking in the magnetite-olivinite); also the apatite content in the hyperite is higher than in the magnetite-olivinite. The hyperite is much coarser-grained than the magnetite-olivinite. In the hyperite, the volume percentage of feldspar averages about 62.6, 13.6 olivine, 5.3 clinopyroxene, and 7.4 titanomagnetite. The plagioclase has An between 50 and 60 per cent. The hyperite is altered in various degrees toward amphibolite, and some rather small fragments of anorthosite are found in the magnetite-olivinite, although these fragments and the rare anorthosite outcrop are too high in dark silicates to be a true anorthosite.

Hjelmqvist says that all the relations between the two major rock types (hyperite and magnetite-olivinite) show that the former is older than the latter. The hyperite magma was the first to be intruded and had crystallized before the magnetite-olivinite was introduced. The magnetite-olivinite contains fragments of hyperite. The hyperite has a common composition with the huge number of dikes and intrusive sheets of that rock that occur within a 500 km-long belt in southwestern Sweden; the magnetite-olivinite and the anorthosite are far less common and owe their formation to conditions peculiar to small portions of the hyperite belt. The hyperite is nearer in composition to the anorthosite than it is to the magnetite-olivinite. The magnetite-olivinite is definitely enriched in iron and magnesium over the hyperite and the anorthosite enriched in calcium, aluminum, and the alkalies over the hyperite. Of the very minor constitutents in the three rocks, vanadium, cobalt, and nickel are most abundant (though minor) in the magnetite-olivinite; chromium is most abundant in the anorthosite. Vanadium is directly proportional to the quantity of magnetite in the ore-rock. Cobalt and nickel are concentrated in both magnetite and olivine, thus explaining their concentration in the ore-rock.

It seems probable that, when the parent magma of the hyperite, magnetite-olivinite, anorthosite series was a magma, it was closest to the hyperite composition but differed from it sufficiently to make it possible for differentiation to produce all three rocks types in the proportions that exist in the Taberg area. Since the outcrops of anorthosite are much less abundant than those of the other two rock types, it is probably impossible to make anything approaching an accurate approximation of the composition of the parent magma, but this does not prevent tracing the steps through which the differentiation process proceeded. Almost certainly the first step in the differentiation scheme was the immiscible separation of the magnetite-olivinite melt from the parent magma. Being much heavier than the remaining silicate-rich fraction, the Fe-Ti-rich fraction sank to the bottom of the magma chamber. The first tectonic forces exerted on the two-magma chamber appear to have resulted in the expulsion of part (probably a large part) of the hyperite magma. This intrusive mass crystallized to form the present hyperite body. Further tectonism caused the intrusion of much of the Fe-Ti rich residuum that appears to have shouldered its way into more or less the center of the hyperite body. There seems to have been little reaction between the hyperite body and the Fe-Ti-rich magma, but an appreciable number of hyperite fragments were broken off during the intrusion and carried along with the Fe-Ti-rich magma to be incorporated in the solid magnetite-olivinite as xenoliths. The Fe-Ti-rich magma was not completely uniform in composition as is indicated by the better ore coming from the plagioclase-poor portion of the magnetite-olivinite body with the portion richer in plagioclase making up a larger proportion of the magnetite-olivinite than the platioclase poor one.

The relationship of the anorthosite to the hyperite is not clear (mainly because of the paucity of anorthosite outcrops, but the lack of anorthosite xenoliths in the ore-rock and their presence in the hyperite indicate that the anorthosite was a product of further differentiation of the hyperite magma after it had separated from the magnetite-olivinite portion of the parent magma rather than having derived more closely from the Fe-Ti-rich magma.

Thus, the magnetite-olivinite magma must be classified as having formed by early separation and late solidification or magmatic-3b. Nevertheless, the separation of the oxide-rich melt appears to have occurred earlier than is normally the case for most ore deposits of this type such as Allard Lake, Cerro de Mercado, the Iron Mountains in Missouri and Wyoming, St. Urbain, Sanford Lake, the Bushveld Magnetite, Mustavaara, or Grängesberg.

Although the hyperite, anorthosite, and the ore intruded into them are definitely intrusive into the ancient (Archean or early middle Precambrian) granite and granite gneiss, when this intrusion took place is not clearly dated. From the character of the ultramafic association of the magnetite-ilmenite ore with the hyperite and the anorthosite is so characteristic of late Precambrian deposits all over the world, it seems necessary to date the Smålands Taberg ore as late Precambrian as is done here.

FINLAND

FINLAND (GENERAL)

Middle Precambrian	Copper, Nickel, Cobalt, Chromite, Iron, Titanium, Lead, Zinc, Pyrites, Silver, Gold	Magmatic, Hydrothermal, Volcanic Exhalations

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Notes

Except for a tiny area of Cambro-Silurian rocks in the northwest corner of Finland, the bedrock of the country is Precambrian in age and makes up part of the Baltic Shield. The oldest rock, the basement that underlies the entire country, is early Precambrian and consists of highly metamorphosed rock in great variety. These rocks, however, are not known either to contain or to have been the parents of any important ore deposits, although they are thought to have been to some extent, remobilized during the orogeny that occurred at, or near the end of, the middle Precambrian sedimentary cycle. The rocks of the upper and middle Precambrian are now known as Svecokarelian; at one time, this rock type was separated into two major divisions, Svecofennian and Karelian, that were obviously of different metamorphic grade, the former lying to the southwest of a line running somewhat irregularly southeast from the head of the Gulf of Bothnia to Lake Ladoga in the former Finnish province of Karelia that is now part of the U.S.S.R. Because the Svecofennian rocks are appreciably more metamorphosed than those of the Karelides, the former originally were considered to be much older than the latter. Later studies, mainly those of isotope ratios, indicate that the two rock groups are of essentially the same age, with the Svecofennides being the eugeosynclinal rocks and the Karelides the miogeosynclinal of a single cycle. This relationship between the two rock groups cannot be considered to be completely established, but it seems to be the most reasonable explanation that thus far has been brought forward (Rutten, 1969). In the orogeny that ended the Svecofennian cycle, two different schist zones were developed that are, not unexpectedly, named the Karelian (northeast) and Svecofennian (southwest). The metamorphism in the Svecofennides is much stronger than that in the Karelides, but both seem to have occurred at the same time. The Svecokarelian rocks appear to have been formed rather late in the middle Precambrian, having isotopic ages that range from 1870 to 1640 m.y. The Prekarelian rocks, the basement of the Finnish portion of the Baltic Shield, probably were formed over a considerable period of time that centers around 2600 m.y. A more detailed division of these Prekarelian rocks (Polkanov and Gerling, 1960) divides them into (from bottom to top): (1) the Kataarchean that consists of a lower granite (3500 to 3200 m.y.) and an upper granite (3100 to 3060 m.y.), (2) the Saamides that consists of a lower group (2870 to 2500 m.y.) and an upper group (2490 to 2100 m.y.), and (3) the Belomorides (2100 to 1900 m.y.) that are confined to the Kola Peninsula and the Skellefte area of Sweden. The Belomorides are enclosed in a big intermontane massif within the wide folded belt of the Karelides.

The lower Karelides are older than the Postkarelian synkinematic granites and the synkinematic Svecofennian granites; the younger Karelides are not intruded by these synkinematic granites but they are older than the postkinematic rapakivi granites that have an age of 1640 to 1610 m.y. The rapakivi granites are much the same age as the alkaline intrusions of northeastern belts of the Karelian rocks of the Kola Peninsula and Karelia with the rapakivi granites being located, in contrast to the alkaline rocks, in the southern belt of the Svecofennides and in part of the southern Karelides of Karelia.

All of the ores of both the Svecofennides and the Karelides are older than the younger Karelides and the rapakivi granites (except for the former Finnish deposits at Pitkäranta (now Pitkyaranta) and younger than the entire suite of Prekarelian rocks; this apparently applies even to the Kemi and Mustavaara deposits as is discussed under those two headings The oldest of the Finnish deposits probably are the Outokumpu ores. Nevertheless, all the Finnish metallic deposits are middle Precambrian, and most of them are late middle Precambrian.

According to Kahma (1973), over 90 per cent of the Finnish sulfide deposits that have been mined or are known to exist as reserves are contained within a belt (the Main Sulfide Belt) that is 40 to 150 km wide and at least 400 km long and extends from Lake Ladoga (southeast) to the Gulf of Bothnia northwest of the Vihanti mine. This belt does not include the geologically and/or economically important deposits such as Ylöjärvi (copper-tungsten), Haveri (gold-copper) Kylmäkoski (nickel-copper), Vammala (Stormi) (nickel-copper), Orijärvi (zinc-lead-copper), Aijala-Metsämonttu (copper-zinc), Kemi (chrome), and Mustavaara (vanadium). Within the 90 per cent belt are the magmatic sulfide deposits of Kotalahti, Hitura-Makola, and Laukunkangas and the volcanic-exhalative or hydrothermal sulfide deposits of Vihanti, Pyhäsalmi, Outokumpu, Vuonos, Luikonlahti, Hammaslahti, and Hallinmäki. Outside the 90 per cent belt are the Orijärvi-Metsämonttu-Aijala belt in the Tampere area and the Ahlainen-Kylmäkoski nickel belt (Vammala and Kylmäkoski) farther south of Tampere. It is possible that the two nickel belts actually form a rude circle partly around the granite core of southwestern Finland and are not separate and distinct entities. Otanmäki, Kemi, and Mustavaara are not, of course, sulfide deposits and do not bear any genetic relationship, the one to the others.

The deposits of the Main Sulfide Belt are in Karelian schist. The Vihanti-Phyäsalmi sub-belt and the Kotalahti nickel belt lie essentially along the demarcation line between the Svecofennian rock belt (southwest) and the Karelian belt (northeast). Simonen (1980) points out that the Karelian belt contains an abundance of quartzites that represent an evolutionary phase on a peneplane, the peneplane having been developed on the Prekarelian granite-gneiss basement. The Svecofennian schist belt is made up of thick accumulations of graywackes in geosynclinal troughs. Simonen, following Väyrynen for eastern Finland, subdivides the Karelian schists into the Jatulian sedimentary group (older) and the Kalevian sedimentary group (younger). The Jatulian group is further broken down from top to bottom into: (1) Sarolian arkoses and conglomerates, (2) Kainnuan quartzite, and (3) marine Jatulian phyllites and dolomites. The Kalevian is not separated into individual formations but is composed of phyllites and mica schists. Simonen thinks that the Jatulian sedimentation occurred on a stable platform and the Kavelian in a subsiding geosyncline. The Karelian rocks in north Karelia do not contain mafic volcanics but do include sills of uralite diabase in the Jatulian; these diabases do not penetrate the Kalevian. The Karelian rocks in other parts of Finland can more or less readily be correlated with the sequence just given for eastern Finland.

The Svecofennian rocks of Finland are not clearly separated from the Karelian because no definite evidence of an unconformity between the two groups has been found. The Svecofennian supracrustal rocks originally were graywackes and impure arkoses that have been metamorphosed into slates, mica schists, mice gneisses, and quartz-feldspar schists, not much pure quartzite or limestone is present. In some parts of the Svecofennian belt volcanics, mainly mafic lavas and tuffs were abundant; these now are metamorphosed into amphibolites. The rocks are considered to constitute a geosynclinal association with the mica schists being similar to the graywackes of orogenic belts and the quartz-feldspar schists (leptites) are thought to have been impure arkoses or, in some of the more leptitic phases, silicic lavas and pyroclastics.

Because the Svecofennian zone itself is quite irregular in direction, its general strike is quite varied, although dips of schistosity and bedding generally are parallel and vertical to steep and usually parallel to the bedding planes. In contrast to the Karelian sedimentation, no ancient basement on which Svecofennian sedimentation took place has been found. Delineation of a stratigraphic sequence in the Svecofennian beds largely is impossible, although, in some places, sedimentary structures are so well preserved as to allow layered rock sequences to be determined. Graded bedding is the structure most commonly usable for this purpose, but others also can be used to some extent. One example of such a sedimentary sequence is in the Tampere schist belt where the series, from top to bottom is: graywacke slates - greater than 3000 m thick; quartz-feldspar rocks (arkoses, graywackes, and pyroclastics) - 1500 to 2000 m thick; mafic and intermediate volcanics - 800 to 1500 m thick; conglomerates and graywacke slates and arkoses - 700 to 800 m thick; and mafic volcanics - greater than 1000 m thick. Similar stratigraphic sections have been found in other parts of the Svecofennian belt. Simonen believes (1980) that the Svecofennian can be divided into lower, middle, and upper sections with the lower two groups listed immediately above being lower, the mafic and intermediate volcanics the middle, and the upper two groups, the upper. The total thickness of the Svecofennian rocks in the Tampere area is at least 8 km, and the sequence has strong affinities to the graywacke-basalt associations of younger geosynclinal deposits.

The absolute age of the Svecofennian and Karelian rocks still is uncertain, but Rutten (1969) considers them to have been formed between 1870 and 1640 m.y. ago, but the age of detrital zircons in the metagraywacke near Tampere is about 2400 m.y. On the other hand, igneous pebbles in intraformational conglomerates in the Tampere field contain zircons that are only 1900 m.y. old; this is in close agreement with the about 1800 m.y. age of the granitoids that penetrate the schists. It is Simonen's tentative opinion that the Svecofennian sedimentation took place from 2400 to 1900 m.y. ago with most of the deposition of such sediments having taken place immediately before the climax of the orogenic movements. The dates given by zircons in the Svecofennian geosynclinal volcanism range between 1920 and 1880 m.y. ago.

In the Karelian beds, the magmatism, dated by the U-Pb method on zircons, occurred between 2200 and 2000 m.y. The iron formation and the dolomite (upper Jatulian) give Pb-Pb ages of about 2080 to 2050 m.y. The Kalevian sediments are younger than 2000 m.y. but are older than the synorogenic plutonic rocks of the Svecokarelides that penetrate the Karelian schists. Thus, it is not unreasonable to assume that the two types of sedimentation (Svecofennian and Karelian) took place during about the same time interval.

The types of sedimentary rocks in both the Svecofennian and Karelian beds and the percentages provided by each are given in the following table:

Rock Types	Kareldic %	Svecofennian %	
Micaceous schists and gneisses	45.2*	79.9	
Quartz-feldspar schists	2.8	6.5	
Quartzites	26.4	0.3	
Limestones	0.3	0.3	
Metabasalts and amphibolites	25.3	13.0	(Simonen, 1980)

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*no Kareldic micaceous rocks are classified as gneisses; all are schists.

Plutonic rocks are prominent in both the Svecofennian and Karelian sections of Finland with the plutonics occupying 40 per cent of the Kareldic folded area and 80 per cent of the Svecofennian. The types of igneous rocks and their abundances in the folded areas are given in the following table:

Rock Types	Kareldic %	Svecofennian %	
Ultramafic and gabbro	3.3	5.9	
Quartz diorite and granodiorite	23.6	56.5	
Granite	73.1	37.6	(Simonen, 1980)

The ultramafic Kareldic rocks mainly are in the Kalevian beds, and they have been altered both metasomatically and metamorphically to serpentinites, talc-magnesite rocks, and chlorite-, anthophyllite-, and asbestos-bearing rocks. They form discontinuous lenticular bodies interbedded with mica schists; originally they were mainly, if not entirely, dunites.

The Svecofennian ultramafics are small bodies of peridotites or hornblendites, the former now being mainly serpentinized olivines, diopsides, and hornblende, and are associated with gabbros that pass gradually into even more silicic rocks. The gabbros normally are quite light and contain 10 to 15 per cent of quartz.

The majority of Finnish igneous rocks in both the Kareldic and Svecofennian belts are granitoids with a color index of less than 30 and range from quartz diorites to granites. The bulk of these lighter rocks in the Svecofennian belt are quartz diorites; in the Kareldic belt, they principally are granites.

The Svecofennian formations are strongly folded, and the main fold axes are subhorizontal with their axial planes being vertical or steeply dipping and usually being parallel to the bedding planes.

The main part of the Finnish Precambrian is contained in the Svecokarelidic orogenic belt with the folding in this belt normally having taken place in the same time-range as the Svecokarelian regional metamorphism and igneous activity, that is, from 1900 to 1800 m.y. ago.

The Presvecokarelian, highly folded basement was penetrated by layered mafic intrusions, the age of which still is not certain. The original opinion was that these mafic bodies were older than the Svecokarelian rocks. For example, Kouvo gave an age of 2440 m.y. for the Porttivaara layered intrusion that includes the vanadium-bearing magnetite body of Mustavaara (see Mustavaara notes). Although this age may be correct, most, if not all, of these mafic rocks appear to have been intruded between basement rock below and Karelian beds above. Mafic bodies so introduced must be Svecokarelian in age. In fact, it is possible that all of them, of which Kemi, Penikat, Syöte, Kuusijärvi-Lipevaara, and Porttivaara are the largest, originally were part of a single connected sheet and, therefore, of the same general age. It has been suggested that the different mineralization (chromite) in Kemi and Penikat as opposed to that (vanadiferous magnetite and ilmenite) in Porttivaara may indicate a different age and/or different magmatic source, but similar intrusive relationships point to the same source and time of entry into the host rocks. If this reasoning is correct, Kouvo's dating of the Mustivaara rocks, 2440 m.y., needs reassessment.

In addition to these layered bodies, a peculiar carbonatite complex at Siilinjärvi is a vertical sheet-like body cutting through the basement gneiss. Its K/Ar dating (on an amphibole) of 2500 m.y. makes it the oldest known rock mass of this type in the world. The trace elements in which this carbonatite is enriched include P, F, Sr, Ba, the rare-earths, particularly those of the Ce group, Zr, and Nb.

The table that follows shows various important ore bodies in Finland, most of which are included in this volume, and lists the particular Precambrian rocks in which they are enclosed, the lead ages (if determined), and appropriate remarks.

Of the deposits listed in the table below, Isokangas (1978) thinks that only three (Outokumpu and its two satellites) may have been formed by syngenetically (i.e., by volcanic-exhalative processes rather than by sedimentary process in the narrow sense). In contrast, the deposits at Hammaslahti, Hallinmäki, Haveri, Orijärvi, Aijala, Metsämonttu, Ylöjärvi, Vihanti, Pyhäsalmi were formed by processes that required the help of hydrothermal fluids (in the broad sense) to achieve the end product that now can be observed. In most of these deposits, two major concepts have been put forward to explain their present condition. The first of these is that the ores were deposited as part of the sedimentary sequence, usually being derived from volcanic exhalations that entered the sea floor during the accumulation of the remainder of the rock column. Later orogenic movements modified the ores and their host rocks to give at least some suggestions of epigenetic introduction of the ores. Under this hypothesis, however, these evidences of movement of ores into fractures are considered to have been caused by remobilization (normally a process not too clearly defined) in which heated waters (hydrothermal in the broad sense) played a major role. In the second concept, the ores are considered to have been introduced by hot water solutions that have leached their mineral content from the rocks through which they passed or inherited them directly from a magma chamber in which they were produced. The time at which these solutions reached the loci of their final emplacement must have been after the host rocks had been lithified and perhaps after they had undergone some appreciable metamorphism. The major question, still not completely settled to the satisfaction of all ore geologists, is the time-relationship of the introduction of ore materials to the complete sequence of metamorphic events. For example, the Vihanti mine has been studied perhaps more thoroughly than any other Finnish sulfide deposit except Outokumpu, yet highly qualified geologists on both sides have argued in favor of one concept or the other, and no final verdict has, in my

Deposit*	Type of Rock in Which Deposit is Contained	Lead Ages	Remarks - Ages from Other Criteria
Outokumpu Vuonos Luikonlahti	Karelian	2250 m.y.	Common lead-model age High Ni and Co contents with varied Co:Ni ratios
Hammaslahti	Karelian	2250-2200 m.y.	2100 m.y. is age of enclosing schist
Hallinmäki (Virtasalmi)	Karelian	1800 m.y. ?	Ore introduced in late orogenic stage
Vihanti Pyhäsalmi	Karelian	2100-2025 m.y. or as low as 1900 m.y.	1860 m.y.-old granite penetrating ore at Vihanti
Orijärvi Aijala-Metsämonttu	Svecofennian	1820-1770 m.y.	
Ylöjärvi	Svecofennian	1840 m.y.	Pb-U from zircon in granodiorite
Haveri	Svecofennian	Probably same as Ylöjärvi	
Kotalahti	Karelian	1925 m.y.	U-Th-Pb from diorite differentiate
Hitura-Makola	Karelian	Probably same as Kotalahti	
Laukunkangas	Karelian	Probably same as Kotalahti	
Kylmäkoski	Svecofennian	1856 m.y.	Pb-Pb model age
Vammala (Stormi)	Svecofennian	Probably same as Kylmäkoski	
Kemi	Karelian (?)	2100-2200 m.y. (Kouvo proposes 2440 m.y.)	Ultramafic host rocks probably were emplaced along unconformity between basement rocks below and Karelian above; synchronous with Karelian orogeny; if so Kouvo's date is too old
Otanmäki	Karelian	2060 m.y.	Probably contemporaneous with Karelian orogeny; if so 2060 m.y. may be too old
Mustavaara	Karelian	2100-2200 m.y.	Probably host magma was intruded before the Svecokarelian orogeny but that the event was as long ago as 2400 m.y. is questionable

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*Most of the data in this Table were derived from Isokangas (1978) and Simonen (1980) opinion, been reached. On the contrary, the large majority of ore geologists who have studied Outokumpu have reached the conclusion that this deposit (and its smaller but highly similar counterparts, Luikonlahti and Vuonos) was emplaced by volcanic-exhalations, although the deposit must have been appreciably modified from its original character through the considerable amount of metamorphic action that has affected the district. Both the genesis of these two deposits and of the others in the immediately preceding table are considered in detail in the discussions appended to each of the appropriate Finnish bibliographies included in this volume.

In contrast to the essentially sulfide deposits lacking nickel in workable amounts, all of the nickel deposits, as well as the chromium deposit at Kemi, the iron-titanium deposit at Otanmäki, and the vanadium deposit at Mustavaara were formed by magmatic processes, the ore minerals crystallizing from molten material in which the water content was low, although locally containing enough water to produce appreciable autometamorphism in ultramafic rocks with which this water, when released from the magma, could react, as was the case at Kemi.

Of the non-nickel sulfide deposits considered here, most - Outokumpu, Vuonos, Luikonlahti, Hammaslahti, Aijala, Ylöjärvi (also tungsten), Haveri (also gold), and Hallanmäki (Virtasalmi) are primarily copper deposits, whereas those of Vihanti, Pyhäsalmi, Orijärvi, and Metsämonttu mainly zinc deposits, although some copper is being, or was, produced from all four; minor lead also came, or comes, from Vihanti and Orijärvi and Metsämonttu.

Outokumpu, Vuonos, and Luikonlahti are of the type that contains both chalcopyrite and sphalerite in considerable quantities. At Outokumpu, copper is 3.8 times as abundant as sphalerite; at Vuonos, copper is only 1.75 times as common as zinc; whereas at Luikonlahti, the copper grade through 1974 was 0.79 per cent with the zinc grade not given. The major distinction between the Outokumpu-type deposits is that they contain appreciable cobalt and nickel - not, of course, nearly as much as in the magmatic nickel-copper sulfide deposits but much more than is usual in chalcopyrite-sphalerite ore bodies. At Outokumpu, according to the latest figures to which I have access (Isokangas, 1978), the cobalt content is 0.20 per cent and the nickel 0.12 per cent. At Vuonos in the copper ore, the cobalt content averages 0.11 per cent and nickel 0.02 per cent. In the distinctly separate nickel ore body at Vuonos, the nickel content is 0.20 per cent, the cobalt 0.02 per cent, and copper 0.04 per cent. In the total ore mined in 1974, the copper content at Vuonos was 2.1 per cent and cobalt 0.11 per cent. Zinc was 1.0 per cent at Outokumou and 1.2 per cent at Vuonos. I estimate the zinc grade at Luikonlahti to be about 0.50 per cent, the grade of cobalt 0.15 per cent and of nickel (my estimate) 0.10 per cent. What the remarkable high absolute amounts of cobalt and nickel in these chalcopyrite-sphalerite deposits may mean as to the source and type of the ore fluids `discussed in the notes on these deposits.

The table that follows (data largely from Isokangas, 1978) shows the grades of the various ore elements in the Finnish deposits described in this volume, plus a few others. It should be referred to in studying the various Finnish deposits.

These deposits were, with the possible exception of the Outokumpu-type, emplaced in solid rocks. As is discussed in some detail in the notes in Outokumpu, it may be that the hydrothermal fluids that brought the Outokumpu-type ores to the earth's solid surface delivered them to the sea floor during a time of geosynclinal sedimentation; it also may be that the deposits were formed in solid rock by replacement and open-space filling. In either event, the initial deposits were so strongly affected by Svecokarelian metamorphism (in its various stages) that their primary character is difficult to determine. The unraveling of the changes effected by metamorphism, however, is a problem in all Finnish ore deposits discussed in this volume as is reasonable to expect for deposits, not one of which is untouched by that process. There are different ways in which metamorphism has affected the Finnish deposits. An example could be chosen from any Finnish deposit, but the autometamorphism at Kemi, the strange removal of primary sulfides from the upper ultramafic layer at Vammala, the unexplained, and perhaps unexplainable, relationships of early and late ultramafics at Otanmäki, and the apparent epigenetic relationship of the ore at Hammaslahti show how varied these metamorphic effects can or may be and how confusing they can be to anyone who attempts to unravel the genetic history of Finnish ore bodies. An attempt is made in each set of notes appended to each bibliography of Finnish deposits to explain these and other phenomena; how successful these efforts are is for the reader to judge.

	Percentages								
Deposit	Cu	Zn	Pb	Ni	Co	W	TiO_2	V_2O_5	Cr_2O_3
Outokumpu[1]	3.51	1.0		0.12	0.20				
Vuonos	2.10[2]	1.20		0.20[3]	0.11				
Luikonlahti	0.97	0.50[e]		0.10[e]	0.15				
Hammaslahti	0.76	0.10			much less than 0.10				
Orijärvi	0.93[e]	2.64	0.81						
Aijala	1.60	n/r							
Metsämonttu	n/r	3.46							
Vihanti[4]	0.62[e]	9.15	0.48[e]						
Pyhäsalmi[5]	0.79[e]	3.35							
Ylöjärvi	0.76					0.12[6]			
Haveri	0.37[7]								
Hallanmäki	0.76								
Makola	0.43			0.81					
Hitura	0.17			0.50					
Kotalahti	0.28			0.70					
Kylmäkoski	0.48			0.55^+					
Vammala	0.30^+			0.60^+					
Laukunkangas									
Otanmäki							13.0[8]	0.26	
Kemi									26.0[9]
Mustavaara							5.0[10]	0.20[11]	

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e = estimated [by me]; n/r = none reported
[1] 1913-1974; [2] Cu ore appears to contain about 0.02% Ni; [3] from a distinctly separate nickel ore body; [4] 1954-1974; [5] 1970-1974; [6] from 17% of the ore mined; [7] plus 0.28 gm Au/ton; [8] Fe content 22%; [9] Cr:Fe = 1.6:1.0; [10] Fe content 22%; [11] in magnetite concentrate.

AHLAINEN-SÄÄKSMAKI NICKEL BELT

Middle Precambrian Nickel, Copper Magmatic-2a

Ervamaa, P., 1962, The Petolahti diabase and associated nickel-copper pyrrhotite ore, Finland: Comm. Géol. Finlande Bull., no. 199, 92 p.

Gaál, G., 1972, Tectonic control of some Ni-Cu deposits in Finland: 24th Int. Geol. Cong., Sec. 4, p. 215-224

Häkli, T. A., and others, 1979, Vammala, a nickel deposit in layered ultramafite, southwestern Finland: Econ. Geol., v. 74, p. 1166-1182

Isokangas, P., 1978, Ahlainen-Kylmäkoski nickel belt in Bowie, S.H.U., and others, Editors, Mineral Deposits of Europe, Volume I, northwest Europe: Inst. Min. and Met. and Mineral. Soc., London, p. 72-73

Kahma, A., 1973, The main metallogenic features of Finland: Geol. Surv. Finland Bull. 265, 29 p.

Matisto, A., 1971, Explanation to map of Prequarternary rocks, Sheet 2121 Vammala: Geol. Surv. Finland, 1:100,000, 44 p. (Finnish, Engl. Summ.)

Papunen, H., 1977, The sulfide mineral assemblages of some Finnish Ni-Cu deposits: 4th IAGOD Symp. Pr., v. 2, p. 311-319 (Varna)

____ 1980, The Kylmäkoski nickel-copper deposit in southwestern Finland: Geol. Soc. Finland Bull., no. 52, pt. 1, p. 129-145

Papunen, H. and Mäkelä, M., 1980, Sulfur isotopes in Finnish nickel-copper occurrences: Geol. Soc. Finland Bull., no. 52, pt. 1, p. 55-66

Papunen, H., and others, Geological, geochemical and mineralogical features of sulfide-bearing ultramafites in Finland: Canadian Mineral., v. 17, p. 217-232

Notes

The material presented in this discussion of the nickel-copper deposits of the Ahlainen-Sääksmäki nickel belt is derived from two excellent articles listed in the bibliography given above, that is, those by Häkli and his colleagues in 1979 and Papunen in 1980.

The Ahlainen-Sääksmäki (A-S) nickel belt extends southeastward from Ahlainen on the Gulf of Bothnia to beyond Kylmäkoski some 150 km further in that direction. Beyond Kylmäkoski some minor deposits and mineralized areas suggest that the A-S belt continues through Telkkälä, Kitula, and Laukunkangas to Kotalahti to join the nickel belt of that name that extends further northwesterly through Hitura-Makola to end at Långbacka near Korsnäs near the Gulf of Bothnia. This almost complete circle surrounds the granitoid rocks of central Finland, but these deposits, although all associated with ultramafic-mafic rocks, are placed in four different age groups, so the mineralization, though perhaps genetically related in a general way, certainly was introduced over a wide span of time.

The A-S belt proper begins in the Ahlainen archipelago north of Pori and makes up a zone about 150 km long and 20 km wide. Isokangas (1978) suggests that this belt is related to the Svecokarelian macrostructures of southwestern Finland. The principal rocks in the A-S belt are Svecofennian migmatitic supracrustal metamorphic rocks with which are associated conformably foliated quartz-dioritic and granodioritic bodies, ultramafic massifs, and warious crosscutting dikes. The trends of the formations and of the fold axes and foliation range generally from northwest-southeast and east-west; in many places, however, the migmatites curve around the granodiorites and quartz diorites.

The most important rocks in the belt, from an economic point of view, are small intrusions of peridotite and pyroxenite that are scattered through the belt in alignment with the regional trend. Typically, the ultramafic masses have structural (shear) contacts with the rocks they intrude, and minor metamorphism has been developed along these borders. These massifs, in detail, are in discordant contact with the rocks that encompass them. The forms of these massifs differ widely, ranging from vertical pipes through plates that are parallel to the schistosity and elongated in the paths of the fold axes, to rootless bodies (rounded or flattish lenses and swarms of fragments in migmatites).

This belt has been intensively explored since the 1960's, largely by the Outokumpu Oy. An appreciable number of nickel-copper deposits have been found but only Vammala (Stormi) and Kylmäkoski are, or have been, in production.

The Finnish ultramafic and mafic rocks in which the nickel deposits might be found have been placed by Papunen and his colleagues (1979) in geologic units of the following ages: (1) postorogenic dikes that range in age from 1650 to 1250 m.y. and, are scattered throughout the country but mainly in south and southwestern Finland; (2) Svecokarelian intrusive rocks in pre- or synorogenic bodies that have radiometric ages of 1900 to 1850 m.y.; (3) serpentinites in the Kareldic belt (e.g., in the Outokumpu area) that tentatively are placed at 2200 m.y. old; (4) layered intrusions (such as those at Kemi or Porttivaara) that are intrusive along the contact between basement rocks below and Karelian rocks above with ages of 2100 to 2200 m.y. (these ages of 2200-2100 m.y. are younger than the 2400+ m.y. given by Kouvo); and (5) Archean ultramafic rocks in greenstone belts and granitoid areas - older than 2500 m.y.

Papunen and his colleagues consider that the rocks listed under (2), (3), and (4) above are those that usually show nickel mineralization or indications of it with the rocks of group (2) being the most important source of nickel. All of the nickel deposits of economic value are in the southern part of Finland, and their ultramafic host rocks were intruded under pre- or syntectonic conditions. For the circular belt of nickel mineralization that rims the granitoid rocks of central Finland no satisfactory structural explanation has been put forward. Only when the circle of nickel-bearing rocks is delimited in a general way does it provide a home for all the major and minor nickel deposits. Perhaps it is purely coincidence that places nickel deposits in such a way that this pattern can be recognized, but it seems more reasonable that there is an underlying cause for it that has not yet been worked out.

In any event, a somewhat detailed summary of the geology of the two best mineralized of the deposits in the A-S belt will throw some light on the problem. Fortunately, both Vammala and Kylmäkoski have been the subjects of recent studies (Vammala, Häkli and others, 1979 and Kylmäkoski, Papunen, 1980), and each of these will be considered in turn.

The Vammala nickel deposit is located about 5 km east of the small town of that name (61°20'N, 22°55'E) that is some 170 km northwest of Helsinki. The first evidence of ore in the area came when a local farmer (from Stormi, a village near Vammala) sent in a peridotite sample to the Outokump Oy that contained a small amount of sulfide mineralization. Later that same year, company prospectors found sulfide-rich serpentinite boulders in the bank of a small river near Stormi; these assayed 1.5 to 2.0 per cent nickel. These finds, plus the location of the site in the A-S belt, lead to further study, and several variously sized and mineralogically characterized ultramafic bodies were found. Two of these, which became the Stormi and Kovero-oja ore bodies, were discovered to contain enough nickel and copper to justify exploitation. Between 1975 and 1977, experimental mining at Kovero-ojo recovered 460,000 tons of ore running 0.43 per cent nickel and 0.32 per cent copper that was produced from both an open pit and from underground. At the same time, surface drilling indicated that about 18 million tons of similar marginal ore was contained in the Stormi ultramafic body. Full-scale mining was begun in 1978; the capacity of the mine was to be between 300,000 and 350,000 metric tons of ore annually in which the mill feed was to be between 0.5 and 0.7 per cent nickel and 0.3 to 0.4 per cent copper.

The Vammala ore is situated south of the border of the Tampere schist belt where the host sedimentary rocks are so highly metamorphosed that they have been converted to veined mica gneisses. The major minerals in these rocks are quartz, plagioclase (An_{25-35}), and biotite; garnet is present as euhedral porphyroblasts that are especially common at the contacts with the ultramafic rocks. The gneisses contain calcium-rich orbicules, and tuffaceous amphibolite; uralite, and plagioclase porphyrites are present as intercalations; the last are cataclatically brecciated. Some occasional marls have been converted to diopside gneiss.

In places, the mica gneisses grade into a coarser-grained garnet-cordierite gneiss that locally includes sillimanite and, rarely, graphite; Häkli and his colleagues refer to them as kinzigites. Both mica gneiss and kinzigites have been migmatized in many places into veined gneisses. In some places the graphite gneisses (usually associated with kinzigites) contain skarn, amphibolite, and rarely limestone. The graphite-rich rocks may run as much as 15 to 20 per cent of that mineral and also contain quartz, plagioclase (An_{30}), biotite, and local cummingtonite.

The major plutonic rocks are granodiorites and quartz diorites; both rocks contain plagioclase, An_{30-35} in the q-d and An_{30-45} in the granodiorite and both are normally not foliated. Quartz, biotite, hornblende, and some pyroxene are major minerals; some minor microcline is in the granodiorite, and some garnetiferous q-d is present. Granite is rare, and it mainly is present as veins of pegmatite and aplite.

The bulk of the ultramafics is peridotite, and it is in isolated and rather small bodies; gabbros and diorites are rare; they may be separate bodies or grade into periodotite or quartz diorite, respectively. The ultramafic bodies are mainly hornblende, usually massive but also present are porphyrytic hornblendites or amphibolitic olivine- and serpentine-bearing cortlandite (hornblende peridotite). Some pyroxene peridotites are present that include olivine and serpentine as well as amphibole.

In the eastern portion of the area, the schistosity normally trends ENE; in the center it gradually converts to ESE, and in the west to SE; dips range between 70° and 90°. On close examination, it is found that the schistosity is most varied in strike as is the elongation of mineral grains and fluid inclusions. Were it not for the extreme metamorphism of the Vammala area, the development of this area probably was much like that of the Tampere schist belt. The crosscutting and the character of the contacts indicate that the putonic rocks are younger than the highly metamorphosed sediments.

The Stormi ultramafic body is vaguely pear-shaped; its narrow eastern neck (Korvalampi) is about 150 m wide and of the same thickness; in its wider western portion, is 500 to 600 m in diameter and reaches a depth of 300 m. The total length of the intrusion is 1.5 km, and it is enclosed in high-grade garnet, cordierite, and sillimanite gneisses. The intrusion is somewhat disconformable with the wall rocks but locally is definitely so. To the west, the ultramafic tongues into the gneiss, and at its western end, it grades out as conformable wedges into the gneiss. The intrusion is made up of at least three layers that lie one on top of the other; each measures about 150 to 100 m in thickness. Thin intercalations (probably residual bands) of gneiss can be traced through the entire ultramafite body.

The lowest ultramafic layer is made up of hypidiomorphic peridotite that contains olivine, ortho- and monoclinic pyroxene, and serpentine after olivine. As accessory minerals are phlogopite, amphiboles, and chlorite. In the thickest part of the layer, the rock grades into dunite; as is true of the peridotite, the dunite exhibits a primary cumulate texture despite heavy serpentinization. On the bottom of the layer is a few meters of perknite (in the sense of being a rock composed of clinopyroxene and amphibole). Most of the sulfide-rich ore is located at the base of the lower layer, but small amounts of ore were trapped higher well within this layer. The intermediate layer is a granoblastic hornblendite that locally approaches cortlandite in composition; the grain size decreases so markedly as the rock is followed upward that it probably cooled rapidly. This layer shows some minor lineation which may mean late tectonic movement. Sulfides are minor and of small size. The uppermost layer is peridotite that in much of its volume shows cataclastic structures; the primary minerals were olivine and pyroxenes, and these have altered to serpentine and amphibole, making the rock in places a serpentinite. At the top of this layer, there are vague signs of a fourth layer similar to the hornblendite. What sulfides are present in the upper layer have been oxidized to magnetite, and chlorite is interleaved with this oxide. Sulfides in a low-grade dissemination are secondary in the upper layer.

At the base of the ultramafite, the mica gneiss shows essentially no change, whereas on the flanks the gneiss contains garnet in considerable quantity. Mica gneiss tongues penetrate into the ultramafite although this effect probably results from major penetration of the gneiss by the ultramafite. A few more siliceous dikes, ranging from gabbro to granite, cut through the ultramafic rock. Locally networks of veins form breccia in which the original composition of the fragments is difficult to determine. Fractures and fissures in the dikes are filled with sulfides, mainly chalcopyrite.

The most important ore bodies, from the economic viewpoint, are known as Sotka-Korvalampi and Syvämalmi. The Sotka body follows the northern contact of the east-trending and boat-shaped part of the intrusion and is a sub-vertical, elongated slab some 500 m long and 5 to 20 m thick. This ore body reaches the surface on the east but the upper edge of the northwestern end is about 120 m below the surface. This relationship appears to be due to a steepening of the axial plunge in the center of the ore body.

The Syvämalmi ore body is in the western part of the intrusion and is an extensive subhorizontal plate that is 5 to 30 m thick at a depth of 299 to 350 m. Both ore bodies contain pyrrhotite, chalcopyrite, and pentlandite as the principal minerals; secondary pyrite is quite common as fracture fillings.

Where they are near the contact with the mica gneiss, the sulfides are coarse disseminations or massive veins that locally cut into the gneiss. As the ore is followed from the contact inward, the sulfides change from coarse- to network-disseminations and then into fine-grained disseminations.

At the northwest end of the Sotka-Korvalammi ore body, the nickel-copper-iron sulfides reach as much as 10 m from the contact of ultramafic rock into graphite- and cordierite-bearing mica gneiss.

That, after emplacement, the ultramafic rock was tectonically affected is shown by the distorted shapes of the body, its internal and contact fractures, and its small-scale displacements. Locally, the ultramafic rock is cut by narrow carbonate-filled veins that branch or are arranged en echelon; these veins may include fine-grained serpentine and pyrite.

The Vammala ore bodies contain four sulfide assemblages that are related to the internal structure of the ultramafic body. These are: (1) monoclinic pyrrhotite-hexagonal pyrrhotite-pentlandite that is found on the contact between the lower layer of the ultramafic body and the underlying mica gneiss as well as, to some extent, penetrating into the mica gneiss below; chalcopyrite also is present and coexists with monoclinic pyrrhotite and pentlandite, and hexagonal pyrrhotite is present as inclusions in the monoclinic variety; (2) the assemblage monoclinic pyrrhotite-pentlandite-chalcopyrite ± cubanite ± mackinawite is most important within the lower ultramafic layer where the sulfides are partly oxidized and are closely intergrown with magnetite (secondary) - magnetite replaces pyrrhotite and fills fissures in pentlandite and chalcopyrite - narrow seams of violarite are between magnetite and pentlandite; intergrowths are common and complex, but pentlandite exsolution bodies do not occur in pyrrhotite, and cubanite and mackinawite coexist with monoclinic pyrrhotite. The intergrowths and exsolution bodies suggest that the original precipitate was a massive solid solution of nickel, iron, and copper sulfides that unmixed on cooling; this assemblage probably is in disequilibrium as is shown by the occurrence together of monoclinic pyrrhotite and cubanite; whether equilbrium was attained early in the exsolution process and then further changes produced the metastable assemblages or the sulfides never were in equilibrium is uncertain; (3) the assemblage pyrrhotite-chalcopyrite-pentlandite ± bornite occurs in very low-grade disseminations in the hornblendite bed; (4) the assemblage mackinawite- pentlandite ± valleriite occurs in association with the pyrrhotite-pentlandite-chalcopyrite-cubanite-mackinawite assemblage [(2)] in the lower ultramafite layer and as the sole sulfide assemblage in the upper ultramafic layer - further, this assemblage is present in serpentinite in almost the entire ultramafic body as a very fine-grained dissemination.

The ores are grouped by Himmi and his colleagues into four types on the basis of sulfide-silicate textures, as follows: (1) massive sulfide portions that appear as breccia fillings or sulfide accumulations close to the contact of the ultramafic body; (2) intergranular disseminations in the lower ultramafic layer - these sulfides fill the spaces between euhedral olivine pseudomorphs - when present in abundance, they produce an extensive net texture that encloses the olivine pseudomorphs; the texture of the intergranular disseminations is coarse, intermediate, or fine depending on the amount of sulfides present - the more sulfides, the coarser the texture - pentlandite is frequently euhedral in this texture but chalcopyrite is sub- or an-hedral; (3) sulfide disseminations that are structurally poorly defined are in mica gneiss and locally in contact rocks - these sulfides may be in large coherent fields, may fill microfractures, or are inclusions in silicates; (4) in the hornblendite bed, the sulfide disseminations are very fine-grained and low-grade; in the serpentinized rocks of the lower and upper ultramafic beds, olivine pseudomorphs contain microscopic sulfide dust, that is, inclusion disseminations; this dissemination is composed of mackinawite, pentlandite, and valleriite.

The selective replacement of sulfides in the lower ultramafic layer largely is the result of oxidation; pyrrhotite is the sulfide most affected, its margins in many places being replaced by magnetite. Some of the pyrrhotite is entirely removed by this process. Additional marginal replacement of pyrrhotite by valleriite and serpentine occurs along its boundaries. The upper ultramafite layer has intergranular textures much the same as those of the sulfide-rich lower layer; in this situation, however, the sulfides have been entirely replaced by magnetite and chlorite.

The ultramafite contains both primary and secondary oxides, the primary ones being ilmenite, chromite, and magnetite, plus Cr- and Mg-bearing hercynite ($FeAl_2O_4$), this latter in the upper part of the hornblendite layer. Secondary magnetite is common in the serpentinized rocks, and some hematite is present in some shears as thin coatings.

The content of nickel in olivine ranges between 600 and 3100 ppm; so does that content in sulfides; the result is a significant positive correlation between the two quantities that suggests that nickel is quite well equilibrated between the olivine and sulfide phases in all three layers.

The mean composition of the sulfide phase in the lower ultramafic layer is: Cu, 4.35 per cent; Ni, 6.35 per cent; Fe, 50.57 per cent; and S, 38.35 per cent. No more than traces of the platinum minerals have been found. The high nickel content of the sulfides, despite the low total-sulfur content, can be explained by equilibrium-producing reactions under subsolidus conditions. From this it follows, Himmi and his colleagues think, that the higher the abundance of the sulfide phase, the lower will be its nickel content. They also think that the Ni/Co ratio depends on whether or not the sulfides were in suspension or were exsolved from the silicate magma during crystallization. If the sulfur content is greater than 0.5 per cent, the ratio of Ni/Co increases steadily from about 10 and reaches 28 in the samples richest in sulfur.

The relative abundances of the Ni, Cu, and Co differ widely; in the basal margin of the lower layer the Ni/Co ratio is 4 and over the next 30 m up, it decreases to 0.3; but it rises to 2 at the top of the lower layer. In the lower margin of the hornblenite layer the Ni/Co ratio is 3; in the next 20 m it rises to 5 but falls by the upper margin to 0.8 - the Ni/Co ratio does not change systematically as the ultramafic layers are crossed. The situation in the upper layer as regards ratios is much different than in the lower two because, in the upper one, the sulfides were oxidized shortly after they crystallized. Apparently copper was removed by the oxidizing solutions because the copper content in the upper layer is much lower than in the lower two.

Work done by Himmi and his colleagues shows that the sulfur isotope ratios in the three layers are, for the lowest two, not greatly different the one from the other; in the third (upper) layer, however, the delta-^{34}S values average 12.5°/$_{oo}$ as compared with the average of -0.79°/$_{oo}$ in the first layer and +1.51 in the second. Little difficulty is posed to supposing that the sulfur in the first two layers came from a magmatic source, the differences in both layers from the 0.00°/$_{oo}$ of the Canyon Diablo sulfur are readily explained by minor reactions within the ore melt or even in the magma from which the ore melt separated. In the third layer, however, it is obvious that some process (or processes) must have affected the sulfur as originally contained in the third layer to change its delta-^{34}S value from the near 0 per mil of the first and second layers to the +12.5°/oo of the third. Himmi and his colleagues point out that microscopic examination indicates that the primary sulfides in the upper bed obviously came from the same source as the sulfides of the lower two beds. They believe, however, that the primary sulfides in the third bed were oxidized and removed soon after crystallization. They suggest that the secondary sulfides were produced by the reaction of iron and nickel in the silicates of the third layer with sulfide ion supplied from some source outside the magmatic system from which the third layer was solidified. They explain the lack of copper sulfides in the upper layer by assuming that copper, much less acceptable in the silicates than iron and nickel, was removed from the system at the time of sulfide oxidation whereas the nickel and iron were retained in the silicates. Their suggestion is that the secondary sulfide ion was derived from the reduction of the S^{+6} of sea water sulfate; this reduction can hardly have been by bacterial action; the sea certainly would have been too hot, if not too deep, for bacteria to perform this chemical operation.

Himmi and his colleagues obtained a radiometric age date from zircons from both a mafic pegmatite in serpentinite and from a gabbroid pegmatoid and from monazite from the gabbroid. The resulting age date was 1890 m.y., not far from the 1856 obtained (see below) from chalcopyrite fractions from the Kylmäkoski ores.

Himmi and his colleagues conclude that a tholeitic magma invaded argillaceous sediments of Svecofennian age that, at the time, were only slightly consolidated. This magma, therefore, easily penetrated the sedimentary pile conformably; this intrusion being accompanied by appreciable assimilation of clay layers. This intrusion and assimilation resulted in the development of peridotitic and hornblenditic variants at the margins of the flow. The bulk of the magma, however, consolidated as dunite, and the sulfides, by that time liquidly immiscible in the magma, sank to the bottom of this first layer where they crystallized. Some sulfides more lately expelled from the magma did not reach the magma floor but crystallized as disseminations within the body of the dunite. Some equilibration of nickel between sulfides and silicates took place at subsolidus temperatures. The second (hornblendite) flow either was a separate intrusion or was separated by crystallization differentiation from the lower layer. This layer probably initially was pyroxenitic but was converted to amphibolite metamorphically. This magma appears to have been saturated with sulfide, all (or nearly all of which) separated from the silicate portion of the magma as an immiscible melt, the bulk of which sank to the bottom of the magma chamber; a minor fraction of this massive ore, however, was trapped in the solidified silicates and itself solidified at some appreciable distance above the floor of the intruded magma. Very little, if any molten sulfides remained in the silicates to crystallize as disseminations in the silicate rock.

The authors consider that the second layer contained much less sulfide than the lower one, perhaps no more than 15 per cent of the content of that layer. It is suggested by them that, because of the fine-grain size of the silicate minerals toward the top of this layer, it must have solidified close to, if not on, the sea floor.

It seems even more certain that the third layer did crystallize on the sea floor and contained no more sulfide than did the second and perhaps even less. Himmi and his colleagues also claim that these primary sulfides were almost immediately oxidized shortly after they crystallized (under the much higher redox potential that obtained on the sea floor than in the magma prior to intrusion). [This suggests that such complete oxidation is the fate of any sulfide or sulfide-containing body that reaches the floor of the sea.] Soon after this oxidation, Himmi and his colleagues believe that the rock of the third layer, still at a high temperature, received a second generation of sulfides, with the sulfur of these minerals being much higher in ^{34}S than was true of those of the first two layers. These iron-rich sulfides equilibrated with the nickel in the igneous silicates and, therefore, have the same partition coefficient as do the sulfides in the two lower layers. The amount of this secondary sulfide material was only a small fraction of the primary sulfide in the lowest layer, probably no more than 0.5 per cent.

The discovery of the Kylmäkoski nickel-copper deposit resulted from the exploration activities in the Ahlainen-Sääksmäki nickel belt of the Outokumpu Oy in the early 1960's that also found the Vammala deposit. The first indication of the Kylmäkoski deposit was the discovery in 1962, by two school children, of a large glacial boulder rich in Ni-Cu ore in the village of Taipale. More formal geophysical exploration and diamond drilling outlined a small, ultramafic sulfide-bearing igneous body only a few 100 m from the location of the first boulder find. The ore-bearing rock does not outcrop at all, being covered by glacial debris. Later, in 1963, reserves were estimated at 515,000 tons of 0.48 per cent Cu and 0.55 per cent Ni ore. It was not until 1970 that the Outokumpu Oy decided to open the mine; the first mining was done from an open pit in March 1971 with underground mining following in October 1973. By September of 1974, the deposit was mined out; from it were recovered 690,000 tons of ore, 890,000 tons of barren rock were removed as well. The total amount of copper metal produced was 580 tons and of nickel metal 1830 tons.

The Kylmäkoski deposit is about 50 km southeast of Vammala and is in the same southwestern segment of the ring shaped belt of nickel occurrences that surrounds the granitoid rocks of central Finland. In addition to these two deposits several other areas of mineralization - Sääksjärvi, Harjunpää, Hyvelä, and Korkeakoski have been located, but these deposits apparently are not large enough or rich enough or both to justify mining at this time at least.

The Kylmäkoski deposit is located in the same migmatized belt of gneisses that contains the Vammala ore bodies and is about 40 km south of the Tampere schist belt. Papunen considers this migmatite the paleosome of the ore body, but this does not mean that it had attained its present migmatitic character at the time the ore-bearing ultramafite was introduced into the area. At present, the gneiss is mainly biotite gneiss that commonly contains bands of garnet, cordierite, hornblende, or graphite gneiss. Volcanogenic rocks are known in the same area and include amphibolites, in part uralites, and plagioclase porphyrites. The rocks of the area have been so highly metamorphosed that the type (or types) of sediments of which they were composed probably never will be known. At Sääksmäki (20 km east of Kylmäkoski) some meta-sedimentary textures are preserved that suggest that the original rocks were turbidites.

A number of synkinematic igneous rocks were intruded into the area, the differentiation series consisting of hornblendites, gabbros, diorites, and quartz diorites. To the south of the Kylmäkoski migmatite belt is a large area of heterogeneous, partly porphyritic quartz diorite in which also are contained irregular masses of diorite, gabbro, and hornblendite. To the north, a large mass of equigranular quartz diorite forms the boundary of the migmatite belt, but this also includes bodies of porphyritic granodiorite.

The ultramafic body is about 260 m long and has a maximum width of 100 m; it is 80 m in depth. The highest (surface) plan-level of the ultramafic body shows it with its long dimension oriented west-northwest, resembling an elongated tooth with a long root and a major cavity on its west side. On this same side, the surrounding rock is migmatitic gneiss. This gneiss is present as lenses along the north and northeastern sides of the body, but, to the east, farther north, and in the lower part of the northern contact, the wall rock is quartz diorite; several dikes of this rock extend into the ultramafite. Where it is in contact with the ultramafic body, the quartz diorite shows curved and locally rectilinear contacts; against the migmatitic gneiss, the contact is in smooth curves and/or in small folds that have the same axial directions as the folds in the migmatite.

A search was carried out to find if the ore body continued beyond the quartz-diorite intrusion; a hornblendite body was found about 500 m northwest of the ore-bearing rock, but it was lacking in sulfides and was low in silicate nickel.

The intrusive body contains several types of ultramafics: peridotites, pyroxenites and perknites, hornblendites, cummingtonite-bearing rocks, gabbros, and diorites. The central part of the mass is thicker than the margins and is composed of peridotites. The perknites and hornblendites are the principal rocks in the tapering peripheral portions of the igneous body; the cummingtonite rocks are found only in the eastern contact zone of the northern part. The diorites and gabbros are chemically different enough from the other types that they apparently resulted from reactions between magma and wall rock.

The peridotites are of two types: (1) equigranular and (2) peculiarly nodular and locally orbicular with large olivine nodules. Type (2) is present at the northern end of the ultramafic mass irregularly distributed through type (1) rock. Type (1) rock is composed of partly serpentinized olivine, both types of pyroxene, and brownish hornblende. Alteration has affected all minerals - olivines to serpentines, orthopyroxenes to cummingtonite, monoclinic pyroxenes to tremolite and actinolite. The richer the original rock in amphibole, the more it grades into cummingtonite or hornblendite; cortlandite is a transitional variety.

The nodular peridotite is rare; it differs from the normal type by having olivine nodules that range in diameter from 1 to 5 cm and have the external form of olivine. As these nodules have a generally common orientation, they show a wavy extinction under crossed nicols. The ground mass amount these nodules is made up of medium-grain pyroxenes and amphiboles with some olivine and sulfide. Only the outer shell of these nodules is olivine; inside the minerals are coarse pyroxene and small olivine crystals - this groundmass resembles that surrounding the nodules. The abundance of sulfides in the nodules is less than in the groundmass outside them. Papunen considers that the nodules are unusual skeletal crystals of olivine that crystallized quickly from supercooled magma. The nodular peridotite is appreciably lower in SiO_2 than the normal peridotite but higher in magnesium and slightly more so as regards iron; sulfur is almost the same; this suggests to Papunen that the nodular type was formed early in the differentiation process. Olivine-poor pyroxenites are uncommon, but perknite (a rock mainly of clinopyroxene and amphibole) is abundant at the north end of the ultramafic mass. Small amounts of plagioclase locally are present, and this rock grades into hornblende melagabbro or gabbro. The cummingtonite-bearing rock almost certainly is an alteration product of pyroxenite.

The feldspar-bearing rocks (gabbro and diorite) are uncommon and probably were formed by assimilation of wall-rock material by the ultramafic magma. Coarse-grained and dark pegmatite veins cross cut the peridotite very locally; at the contacts of these two rock types are narrow, monomineralic zones (from 1 to 10 cm wide); these zones contain, in the following order, away from the unaltered peridotite, such minerals as talc, actinolite, anthophyllite, and chlorite. This reaction is greater where the pegmatite veins cut quartz-diorite dikes. These reactions rims exist only where the host rock is peridotite; similar effects have been reported from Hitura and Kotalahti.

In contrast to many other sulfide deposits associated with ultramafic rocks, the sulfides are principally as disseminations that locally are abundant enough to form a continuous net among the silicates. In olivine- and pyroxene-rich peridotites, the disseminations are interstitial. In the hornblendite, however, the sulfides in many places are in roundish spheres. This indicates to Papunen that the spheres crystallized before the host hornblende. In the nodular peridotites, the sulfides locally occur as small, massive layers that grade upward into interstitial disseminations that are located between the nodules. Chalcopyrite in many instances is close to the top edges of the sulfide grains that rest on the olivine nodules. Apparently heavy sulfide drops settled down as a massive layer, displacing olivine crystals that previously had occupied that magma volume. Along the basal contacts of the ultramafite, sulfides have entered into fine breccia fractures that extend into the wall-rock gneiss. In the cummingtonite-bearing rocks, the sulfides are present as anhedral grains that, because of metamorphic recrystallization of the amphibole, penetrate the amphibole needles. A few veins of sulfides, up to 20 cm wide, can be followed for over 10 m in places; these veins are bounded by chlorite-rich slickensides that can continue on after the sulfide centers die out. In places, massive Ni-arsenides form the continuation of veins of massive Ni-Cu sulfides. Most Ni-arsenide veins, however, are found in the tapering ends of quartz dikes or in contact shear zones in intrusive quartz diorite.

The main minerals in dissemination and breccia ores are pyrrhotite, pentlandite, and chalcopyrite. In breccias rich in chalcopyrite, lamellae of cubanite approach the abundance of chalcopyrite; this also applies to disseminations in cummingtonite-rich rock. Pyrrhotite is principally hexagonal, but a minor amount of the monoclinic variety is as lamellae in the hexagonal phase. In the disseminations in peridotite, euhedral grains of pentlandite are common, but, in other ore types, granular and exsolution types of pentlandite are abundant. In some pentlandite, exsolved lamellae of chalcopyrite are common. In serpentinized peridotite, pentlandite crystals may contain networks of serpentine and magnetite along octahedral cracks.

The minor minerals in the ore include mackinawite, which is particularly abundant in the serpentinized portion of the ultramafic as small, oriented flakes in pentlandite or in small veinlets along former cracks in serpentinized grains of olivine, and finally as an independent dissemination in serpentinite. Graphite in places is quite common in the eastern part of the massif as large flakes or round spheres in pyrrhotite. Some molybdenite is present where graphite is abundant. In the chalcopyrite-rich ores, argentian pyrrhotite is minor but common. In the disseminated ores are rare euhedral grains of gersdorffite-cobaltite in association with sulfides; some gersdorffite is found in the Ni-arsenide veins. Niccolite and maucherite are prominent minerals in the arsenide veins. The mineral assemblage in the arsenide veins also includes chalcopyrite as well as argentian pentlandite, pentlandite, and wehrlite, galena, mechenerite, and other minerals containing the platinum-group metals.

The work of Papunen has established definitely that the ore sulfides that give the economically viable character to the ore at Kylmäkoski were brought in with the molten silicate magma from which the ultramafic-mafic rock suite that now contains the ore was developed. Most of the sulfides dissolved in this magma became immiscible in the magma before more than a small fraction of the olivine had crystallized. The evidence for this is that most of the sulfides were disseminated interstitially in both the peridotites and pyroxenites and were included to an appreciably lesser extent in the olivines. In hornblendite, however, separation of the sulfides from the parent silicate melt occurred early enough in the crystallization cycle of that type of magma that sulfides were able to collect in rounded spheres before being prevented from doing so by an increasing abundance of crystallized silicates. On occasion in the nodular peridotite, the sulfides did accumulate in small massive layers that grade upward in interstitially disseminated sulfides.

Papunen presents no evidence to show that any of the primary sulfides at Kylmäkoski were oxidized and converted to magnetite as was true at Vammala nor were secondary sulfides developed. This suggests that the Kylmäkoski ultramafic body and its ores were crystallized appreciably farther beneath the sea floor than was true of the rock-ore sequence at Vammala. Papunen, however, does believe that the olivine nodules did crystallize rapidly but that either the local temperature gradient was not steep enough to produce skeletal olivine blades or that the magma was moving during the crystallization of the olivines so large, equidimensional crystals were formed. Nevertheless, this relatively rapid crystallization of the olivine nodules does not indicate as near-surface conditions as obtained at Vammala. Certainly, at Kylmäkoski, the ultramafic body underwent appreciably more crystallization differentiation from peridotites to hornblendites than did the various layers at Vammala. If it is assumed that the cummingtonite-rich rocks and the diorites and perhaps even the gabbros were developed by reactions between the ultramafic magma and the wall rocks and not by crystallization differentiation, the differentiation at Kylmäkoski was more complete than that at Vammala.

The classification of the Vammala and Kylmäkoski deposits basically is simple; they undoubtedly both were produced by the immiscible separation of sulfides from the ultramafic silicate melts. This separation probably took place early in both deposits, and the crystallization of the sulfides occurred quite late in the cycle of solidification of the combined silicate-sulfide melt. The major difference in the environment for the solidification process in the two melts was that, for Vammala, the intrusion entered essentially unlithified sediments whereas, at Kylmäkoski, the rocks into which the molten material was introduced were quite solid. This means that the Vammala melts were solidified close to, or at, the surface whereas, the Kylmäkoski melts crystallized at an appreciable distance beneath the surface. In both melts, however, separation of sulfides from silicate melts appears to have taken place early in the solidification cycle but final crystallization of the sulfide portion of the molten material was late in the cycle with the sulfides crystallizing largely, or entirely, after the silicates. At Vammala, in the first layer of the ultramafic magma, the major fraction of molten sulfides sank through the silicate melt to the bottom of the magma chamber. Only a minor fraction was trapped by crystalline silicates and itself solidified as disseminations at some appreciable distance above the chamber floor. The behavior of the second Vammala layer was much the same as that of the first except that molten sulfides did not reach the bottom of the magma chamber to crystallize as massive body as did the bulk of the sulfide residuum did in the first Vammala layer. The peculiar oxidation of the third layer does not affect the primary classification of both the deposits as Magmatic-2a.

At Kylmäkoski, the sulfides did not have time to separate as a massive body on the floor of the magma chamber but, instead, were trapped as disseminations of minor massive bodies within the silicate mass. Thus, Kylmäkoski sulfides were accumulated as: (1) a continuous net around the silicate crystals; (2) roundish spheres, or (3) small massive layers that grade upward into interstitial disseminations located among the nodules of the nodular peridotite. Nevertheless, these various arrangements of sulfides were the result of early separation and late crystallization, so the Kylmäkoski ores also should be categorized as Magmatic-2a.

Because the Vammala and Kylmäkoski ore bodies are located in ultramafic rocks that have radiometric ages of between 1900 and 1850 m.y. and because the ores were generated within the ultramafic rocks that now contain them, their ages must be those of their host rocks. They are, therefore, Middle Precambrian.

HÄLLINMÄKI (Virtasalmi Area)

Middle Precambrian Copper Hypothermal-1

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Pääkkönen, V., 1954, Tutkimukset Juvalla: Manuscript in the Archives of the Geol. Surv. Finland (Finnish)

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Notes

The Hällinmäki copper deposit is named for the small town of Hällinmäki in the Virtasalmi district, about 90 km south of the city of Kuopio and 55 km south of the Kotalahti mine (from which mine the Hällinmäki mine is administered). Exploration in the area was begun in 1960 because of the finding, by local farmers, of copper ore-bearing boulders; exploration was carried out by the Geological Survey of Finland, but, for mining purposes, the property was turned over to the Outokumpu Oy in 1966. Initially, the ore was mined from an open pit, but, in 1972, the operation went underground. By 1974, open pit and underground operations had produced 1,570,000 tons with a grade of 0.71 per cent copper. At that time, annual production was 260,000 tons of ore; copper concentrates are produced at the mine; and estimated reserves were in 1974 15,000 tons of copper (as metal).

The Hällinmäki mine is located at about 62°04'N, 27°40'E. As is usual in glaciated terrain, outcrops are rare, averaging about 5 per square km. Geophysical measurements were made along lines 50 or 100 m apart with the distance between observations being about 20 m. These results were placed on maps with a scale of 1:2000; they were published in combination maps on a scale of 1:20,000 and 1:50,000 (Siirkarla, 1967). Measurements of the densities and susceptibility of the different rocks were made (also by Siirkarla). The general structure of the district appears on a magnetic map, but other geophysical work must be done because the same rock may have different susceptibilities in different places or different rocks may have the same susceptibility. Gravimetric methods are also confusing because different rocks may belong to the same density class. But using all these data together has resulted in the production of reliable maps of the bedrock. It was found that, by electromagnetic methods, conducting sulfide- and graphite-bearing gneisses (that have the same magnetic intensity) easily can be differentiated from non-conducting amphibolites. Some amphibolites do contain enough sulfides to be conducting, but such amphibolites can be separated from sulfide-bearing gneisses by gravimetric methods. The long limestone zone in the district can be distinguished because it is non-magnetic and is in the zones of minimum gravity. The forms of positive gravimetric anomalies provided by mafic plutonic rocks are in general more or less round, whereas weakly magnetic amphibolites are long and narrow. Diorite massifs can be told from amphibolites because the amphiboles are of appreciably lower density. It is, however, very difficult to distinguish between non-sulfide-bearing gneisses and quartz diorites by geophysical means; thus, the boundaries between these two rock types are not firmly fixed.

According to the geophysical maps, the regional axis appears to be nearly horizontal in the middle of the anomaly zones; the axial plunge, on the basis of fold axis observations, is 40°-60° SE at the southeastern end of the zones and 40° NW at the northwestern end. This arrangement suggests to Hyvärinen (1969) that the major structure in the district is an anticline (or antiform). A break in the gravimetric anomaly and a weakening of the magnetic anomalies at their northwestern end probably is caused by rise of the regional axis with the result that, underneath the diopside amphibolite, is a lighter (in mass) mica gneiss. The folding appears to Hyvärinen to be isoclinal except at the southeastern end where the folds probably are appreciably more open.

The bedrocks of the Hällinmäki mine area are supracrustal and trend in general to the northwest. In the central portion of the district, in particular, large massifs of plutonic rocks occur. In the northern section of the district, these massifs cut into the schist complex (outside the immediate vicinity of the ore), but the smaller bodies of these rocks normally are concordant with the bedding.

Hyvärinen (1969) reports that the percentages of the various rock types in the district are as follows:

Mica Gneiss	36%	Limestone	2%
Diopside Gneiss	1%	Quartz Diorite and Diorite	48%
Amphibolites	12%	Gabbro and Peridotite	1%

Thus, almost half of the bed rock of the district is made up of highly metamorphosed stratified rocks, the original character of which usually is difficult to ascertain. The foliation in most cases is parallel to the bedding or banding with the exception of fold crests and troughs where, of course, it intersects the bedding. Except for the southwestern corner of the district, the dip of the foliation is 70°-90° N-NE. Locally, the forceful intrusions of the igneous bodies have distorted the foliation, although, in general, the igneous-rock bodies are elongated parallel to the foliation.

In addition to the major folding, the axis of which strikes northwest and is essentially horizontal in dip, much minor folding in the area probably resulted from drag folding on the limbs of the major folds. The axes of these minor folds, however, normally have a steep to vertical plunge. Lineations, formed by the parallel arrangement of minerals on the foliation planes, may follow, or deviate from, the directions of the axes of the minor folds. No rocks older than the highly metamorphosed sediments are known in the district; therefore, no representatives of the pre-Svecofenninan basement are known in the area. The interrelations of the various stratified beds are highly complex, but, as far as they have been worked out, appear to agree with those determined for southwestern Finland (Härme, 1954).

Hyvärinen (1969) considers that the following stratigraphy relationships exist among the supracrustal stratified rocks (the uppermost formation is the youngest in this table):

garnet-, cordierite-, and sillimanite-bearing gneisses

amphibolites, the major part of which consists of diopside amphibolites with intercalations of hornblende gneiss, uralite-plagioclase porphyrite, and garnet skarn; the copper occurrences of the district are invariably located in, or adjacent to, the garnet skarn

diopside and quartz-feldspar gneisses with limestones

garnet-, cordierite-, and sillimanite-bearing gneisses with graphite-bearing intercalations

unknown base of sedimentation

The plutonic rocks include peridotites, gabbros, diorites, quartz diorites, trondhjemites, and pegmatites. Peridotites and gabbros are in close association with each other as are diorites and gabbros. These rocks in many places grade into each other, but the contacts between quartz diorites and more mafic igneous rocks usually are marked by zones of breccia, but, in places, the quartz diorites grade into trondhjemites. The pegmatites and aplites cut all other rock types and are post-orogenic.

The mica gneisses compose more than one-third of the rocks of the district and are largely veined gneisses; they are strongly folded and primary structural features have been largely obliterated. Quartz-feldspar gneisses, amphibolites (generally much broken), hornblende gneiss, and diopside gneiss occur as inclusions in the mica gneisses near amphibolites. Across the contact zone, in the amphibolites, narrow mica gneiss intercalations are found. The chief minerals of the mica gneiss are plagioclase, quartz, and biotite in quite varied proportions. In the Virtasalmi district, the mica gneisses change gradually into veined gneisses, the veins appearing first as small lenses that steadily increase in size and coalesce to produce veins composed mainly of biotite and quartz plus plagioclase.

The quartz-feldspar gneisses form rather narrow and vaguely reddish intercalations of varied thicknesses in mica gneiss or amphibolite. The major example of these gneisses is 100 m thick, and essentially all of them occur on the northeastern and eastern sides of the great limestone horizon. Some of these gneisses contain what apparently are bands due to bedding. The minerals in these rocks are quartz, plagioclase and K-feldspar, plus some mica. Accessory minerals in the gneisses are directly related to the rocks with which they are in contact.

The diopside gneisses only make up about 1 per cent of the rocks of the district and usually are present as intercalations in the mica gneisses near limestone and diopside amphibolite; the rock differs appreciably in coarseness. Quartz, plagioclase, and diopside are the main minerals with the coarser type having appreciable actinolite; some of the plagioclase has been converted to epidote. This rock type is thought to have been produced from carbonaceous sandstones; all transitional forms between mica gneisses through diopside gneiss to quartz-feldspar gneiss have been seen in the area.

Amphibolite intercalations are common in diopside amphibolite; many limestone beds and quartz-feldspar gneisses layers are found in the lower portions of the diopside amphibolites. In these diopside amphibolites also are huge zones composed almost completely of garnet; these may be hundreds of meters long and many meters wide. Intercalations of hypersthene and cummingtonite amphibolites are irregularly arranged throughout the diopside amphibolite. Near the plutonites, diorite or quartz diorite dikes and veins of various thickness concordantly enter the diopside amphibolites; the diopside amphibolite and its intercalations are strongly folded.

The garnet-skarn zones are concentrated in the lower part of the diopside amphibolite; during folding, the garnet layers are stretched into boudins. The biggest skarn zones are those that contain the copper minerals of the Hällinmäki ore body, these zones being 100 to 200 meters long and 10 to 20 meters wide. The main minerals are garnet and hedenbergite plus various amounts of epidote, plagioclase, scapolite, and magnetite with some accessories. Some of the garnet layers are almost monomineralic, and the crystals are idiomorphic or rounded and one to three centimeters in diameter. The garnet runs over 70 per cent andradite ($Ca_3Fe_2Si_3O_{12}$) and nearly 20 per cent grossularite ($Ca_3Al_2Si_3O_{12}$). The pyroxene of the diopside amphibolite contains 40-45 per cent of the hedenbergite molecule; this percentage rises 80 to 90 in the garnet layers. The plagioclase is made up of 60 to 70 per cent anorthite; it may partly alter to epidote and/or scapolite where the skarn contains ore minerals. These latter two minerals also are present in barren skarn, and there they are pre-ore. Epidote of this type locally is penetrated by ore minerals. Scapolite in the barren skarn is composed of 70 per cent meionite [$3CaAl_2Si_2O_8 \cdot CaCO_3$ (or $CaSO_4$)], the same composition it has in the diopside amphibolite, but the ore-rock epidote is appreciably richer in pistacite (ferric-iron-rich epidote) than that formed during the regional metamorphism. Magnetite develops in various parts of the garnet-skarn layers in pockets a few meters long; these pockets were broken during folding.

Amphibolites are present in the diopside amphibolite, and the contacts between the two rock types are poorly defined. The thickest layer of amphibolite is 100 to 150 meters long and is located on the northeastern side of Hällinmäki. On the northeast side of this layer, the amphibolite gradually is converted to quartz diorite or diorite, but in which residual amphibole fragments remain. During folding, these amphibolites were brecciated, and the fractures are in many places penetrated by diorite or quartz diorite. The principal minerals of the amphibolite are plagioclase and hornblende, plus small amounts of biotite; the percentages of plagioclase and hornblende range widely.

Biotite-amphibolite dikes cut across the bedding of diopside amphibolite; both of these rocks were folded at the same time. The plutonic rocks of the area are younger than either of these amphibolites that they intruded after the folding. The main minerals in the biotite amphibolite are hornblende, biotite, plagioclase, and in one instance, anthophyllite; again the relative amounts of the various minerals differ appreciably from one location to another.

Hypersthene amphibolite is a minor rock in the center of the long diopside-amphibolite zone on the west side of Hällinmäki. The usual plagioclase and hornblende are present as is hypersthene; the hypersthene and hornblende are essentially never in the same layer.

A minor intercalation of cummingtonite amphibolite is in diopside amphibolite north of Narila (which is about 5 kilometers S-SE of Hällinmäki); the cummingtonite makes up over 50 per cent of the rock with the rest being mainly plagioclase and minor biotite.

The materials from which these various amphibolites were produced probably were mafic volcanics (fresh and partly weathered) and sediments; some of the sediments may have been carbonaceous. In most cases, however, the metamorphism of these rocks has been so severe that the original structures cannot be determined, but, in the southern portion of the area, distinct banding of alternately hornblende and diopside amphibolite occurs, suggesting that the material originally was a sediment. It is possible, as was believed by Härme (1954), volcanic and calcareous muds were deposited simultaneously. Mica gneiss intercalations are most abundant in the lower diopside amphibolite layers; in such locations, however, the diopside amphibolite may have originally been marl. The other amphibolites, except biotite amphibolite, also may have been produced from similar materials. The varied composition of the biotite amphibolite probably resulted from the original filling of fissures in the surrounding rocks with sedimentary material that differed in composition from one place to another.

The ore minerals in the skarn were, for a long period, thought to have been introduced metasomatically by the attack by metal halides and silica on limestone. A later suggestion was that the skarn rocks and their ores were produced by recrystallization of sedimentary material during regional metamorphism. This concept seems to suggest that the ore elements were already present in the sediments. Hyvärinen thinks that the process was one of recrystallization of sedimentary material through regional metamorphism to form the first skarn phase by reactions among iron minerals, silica, and carbonates. The ores were introduced hydrothermally later, hydrothermal reactions producing a second skarn phase as well. Magnetite in the garnet zones differs appreciably in chemical composition from that in the quartz diorite; thus, the garnet-zone magnetite probably was a product of regional metamorphism and that in the quartz diorite was deposited from the magma. During the ore-forming phase, the fractures in the garnet zone were invaded by the ore-forming fluids that, in addition to depositing ore, attacked the plagioclase in the country rock and converted it to epidote and scapolite thereby modifying the original skarn developed by regional metamorphism.

Limestones in the Hällinmäki mine area are abundant, particularly in, or near, the contacts of diopside amphibolites and mica gneisses. In most places, the limestones are only a few meters thick. An exception to this is the one large mass some 10 kilometers long and from 300 to 400 meters wide. A rather large limestone lens occurs on the northeastern side of the Hällinmäki ore body. Actually, the limestone is a diopside-wollastonite limestone; this type contains several other silicate minerals such as epidote, scapolite, garnet, olivine, humite, serpentine, sphene, amphiboles, micas, quartz, and local and minor amounts of chalcopyrite and magnetite grains.

The major limestone zone is bounded in the southwest by diopside amphibolite and in the northeast by amphibolite or diorite. The limestone zone contains intercalations of diopside amphibolite and garnet skarn and, in a few places, quartz-feldspar gneiss. These intercalations range in width from a few to several tens of meters with the contact zones between limestone and these other rock types being only about one centimeter thick. The minerals of these contact zones are mainly andradite and hedenbergite, plus local wollastonite and epidote.

The actual limestone is almost pure calcite in quite coarse grains, from five millimeters up to one centimeter. The diopside-hedenbergite in the limestone is much finer grained than the calcite (.02 millimeters). If the pyroxene occurs alone, it normally is pure diopside; in company with other pyroxenes, it contains appreciable hedenbergite molecule. Wollastonite commonly is at the contacts between limestone and country rock in veins about one centimeter wide. It also is present as clusters or disseminations in the limestone away from the contact. The scapolite has the composition of nearly pure meionite [$3Ca_3Al_2Si_2O_8 \cdot CaCO_3$ (or $CaSO_4$)]; sphene occurs in well-developed crystals up to three millimeters long. Epidote is rich in pistacite. Olivine and humite [$Mg_7(SiO_4)_3(F,OH)_2$] minerals (the structure of which is closely related to that of olivine) are rare in Hällinmäki limestone, and they may be partly or entirely serpentinized. Garnet-skarn zones in the limestone are identical to the garnet-skarn zones in the diopside amphibolites. Hyvärinen thinks that both the zone-types formed in the same manner, that is, through regional metamorphism, and this suggests that the garnet-skarn zones were formed from limestone lenses or were the most intensely metamorphosed portions of carbonate rocks converted mainly to diopside amphibolite.

Also in the diopside amphibolite are various thicknesses of hornblende-gneiss intercalations. The main minerals in these are plagioclase, quartz, biotite, and hornblende; the proportions of these differ considerably from one lens to another. The original material of this gneiss probably was tuffaceous material, where it is porphyrytic, but the even-textured material near mica gneisses and quartz-feldspar gneisses may have been, Hyvärinen thinks, carbonaceous sandstones.

Minor amounts of uralite-plagioclase porphyrites and metabasites are in diopside amphibolite; these rocks are of no importance worth mentioning.

Hyvärinen believes that the metamorphism of the mica gneisses and quartz-feldspar gneisses took place at a temperature of between 640° and 680°C with a pressure being exerted of between 2500 and 4000 bars. On the contrary, he points out that, adjacent to diopside amphibolite, epidote is equilibrium with quartz. This mixture will react, at fitting temperature and pressure to form anorthite and some grossularite/andradite solid solution, plus some hematite and H_2O. At 680°C, the pressure required for this reaction to go forward is 6100 bars, so the pressure involved at Hällinmäki could not have been equal to, or over, this amount. He thinks that the reactions that took place in carbonaceous rocks during regional metamorphism may have prohibited the development of cordierite in the adjacent mica gneisses by the formation of Co_2 and H_2O in the wollastonite reaction that raised the pressure in the area to beyond the limits of the stability field of cordierite.

According to Hyvärinen, the peridotites, gabbros, diorites, quartz-diorites, and trondhjemites are syntectonic. In a rough way, these rocks are arranged in a regional pattern; the trondhjemites and the granodioritic varieties associated with them are principally in the northwest corner of the district; peridotites and gabbros are most prominent in the area N-NE from Narila and between the Ankele quarry and Luomanen, about 5 km east-southeast of the quarry. Although the granodiorites usually are in crystalline schists, they also are quite close to the large quartz-diorite bodies. The quartz diorites are, in many places, appreciably contaminated by schist and mafic plutonic-rock fragments. Essentially all the plutonic rocks of the district are folded, the only exception being the silicic dikes. Thin diorite dikes that penetrate the schists show boudinage structures. All of the rocks of the area are cut by the aplite and pegmatite dikes.

The peridotites do not form independent intrusions but are closely related in space to the gabbros; they normally occur as tongues and lenses in the crystalline schists or in the more silicic igneous rocks. These bodies average about 0.5 kilometers long and 100 to 200 meters wide. The boundaries between the two mafic rock types are vague and indistinct as the change from gabbro to peridotite is accomplished by a steady increase in dark minerals. The minerals around the borders of these bodies are parallel to these, but the central portions of them are massive. Hornblende is the main mineral in the peridotites with lesser amounts of plagioclase, biotite, olivine, hypersthene, augite, and serpentine, plus a variety of accessory minerals. Although the proportions are different, the various individual minerals are quite similar in composition from one place to another. The plagioclases run about 60 to 70% An. If these rocks contain pyroxene, they include no primary biotite, and the hornblende is different in pyroxene-bearing than in biotite-bearing rocks.

The diorites are in small, irregularly shaped bodies associated with the gabbros or in the quartz diorite. These rocks also have foliated margins, and the principal minerals are hornblende, plagioclase, biotite, and, in places, cummingtonite, plus a wide variety of accessories that include the major ore minerals. The plagioclases have an average An content of 30 to 40 per cent.

It is impossible to draw sharp boundaries between diorites and gabbros with the proportion of dark minerals becoming smaller as the rocks are followed from gabbro to diorite to quartz diorite. The main difference, as would be expected, between diorites and quartz diorites is the presence of quartz in the latter. Interestingly, the diorites contain two to four volume per cent of magnetite, a mineral lacking in the gabbros and peridotites. Ilmenite includes apparently exsolution bodies of magnetite. The pyrrhotite in the diorites does not contain exsolution blebs of pentlandite as does the former mineral in peridotite and gabbro.

The main igneous rock of the district is the quartz diorites, and it occurs as elongated bodies that, in most places, parallel the foliation. These massive bodies are over 10 kilometers long and several kilometers wide. Small areas in the central portions of these bodies may be massive, but all other quartz diorite shows mineral foliation; the diorite normally is an even-grained rock. The late age of the quartz diorite, relative to the other igneous rocks, is shown by the fragments of the mafic rocks and of the crystalline schists that it contains. The contacts between the host rock and the fragments usually are sharp, although the fragments of quartz-feldspar gneiss may be somewhat rounded. The quartz diorites brecciates the diorite and the more mafic rocks as well. Even though the quartz diorites in many places are mixed rocks, large areas of homogeneous quartz diorite exist. The main minerals are plagioclase (An = 25 to 35 per cent), quartz, hornblende, biotite, plus the usual accessories. Along the margins of the quartz diorite bodies north of Narila and northwest of Litmanen, the rock is porphyrytic and foliated; towards the central parts of these masses, the porphyroblasts are more abundant and apparently less foliated.

Trondhjemite is closely associated with quartz diorite and is most common northwest of the Virtasalmi area. Locally, the trondhjemite forms narrow lenses between the diorite and the schists. All of the trondhjemite shows some mineral lineation and is made up principally of plagioclase (20 per cent An, mainly but locally as much as 30 per cent), quartz, and biotite with minor quantities of hornblende. Microcline is present in some of the rock, converting it to granodiorite; most of the microcline appears to replace plagioclase, and some of it is myrmekitic, suggesting a secondary alteration produced the microcline.

Although pegmatites and aplites are not common in the Hällinmäki area, they do cut all other rocks and were not folded. Some pegmatites are, however, older than others, as the intersecting relations among them show. Quartz and plagioclase are the main minerals in the pegmatites with minor muscovite and even less tourmaline. The principal minerals of the aplite are microcline and quartz plus a little oligoclase and minor muscovite.

Hyvärinen says that the diorites and quartz diorites are the dominant rocks of the area (in quantity at least) in a suite that ranges from peridotite to trondhjemite. A very low K_2O content (well less than one per cent) is characteristic of all the igneous rocks of the peridotite to trondhjemite suite, and an above-average content of Na_2O marks the diorites and quartz diorites at least.

The copper ores are, in all instances, associated with the garnet skarns that occur in, or near, the large quartz-diorite body on the southwestern side of Virmasjärvi, a large lake than trends NW-SE, following the direction of the movement of glacial ice. These skarn-copper deposits usually are quite small. The peridotite body, which is a little over one kilometer southeast of the southeast end of Virmasjärvi, contains some chalcopyrite and pyrrhotite in which are exsolution blebs of pentlandite; these minerals are not present in economic amounts.

Of the three copper-bearing bodies in garnet skarn in the Virtasalmi area, (Hällinmäki, Karhuniemi, and Lari) only the first named is of economic importance. The ore body (Hällinmäki) is located on the southwestern edge of a formation of schistose amphibolite (about 750-800 meters long and 200-250 meters wide) that is bounded and, to some extent, intruded by diorites. The schists strike between 50° and 60°W, and the dip is generally 65° to 75°NE. Intense drag folding is present in the schists, the axes of which folds plunge to the southeast from 10° to 20° to 60° to 70°. Along the northeast margin of the schist body are amphibolite and limestone formations and along the southwest are diopside-amphibolite horizons with intercalated skarn lenses. Of the plutonic rocks that cut across and follow the foliation in the schists, diorite is the most abundant rock, gabbro is not common, but what is there is at the southwest margin of the ore occurrences. Quartz diorite and trondhjemite are minor in amount and are located in the southeastern part of the ore area. Except for the silicic dikes, all the Hällinmäki igneous rocks were deformed at the same time that the drag folds in the schists were produced.

The zone that contains the ore is about 500 meters long and two to 50 meters wide and is at the contact between diopside amphibolite and diorite. The largest ore body (I) has been traced to a depth of 350 meters, and other and deeper extensions may be found. The other ore bodies are shallower and show less continuity. The separate ore bodies stand almost vertically and probably are parallel to the axes of the drag folds. In many instances, the tops of small drag folds are mineralized in line with the axis whereas the surrounding rock is entirely lacking in ore minerals.

In 1966, the Geological Survey of Finland estimated that the Hällinmäki deposit contained about two million tons of copper ore with an average grade of 0.95 per cent.

The skarn zones are located in diopside-amphibolite masses, but the mineable ore bodies are not in all garnet skarns, and, where they are in garnet skarns, the ore in many instances also spills over into the surrounding diopside-amphibolite.

Copper is the only economically recoverable material in the ore at Hällinmäki, the copper being evenly distributed through the ore lenses. The ore contains traces of cobalt and nickel with the highest nickel content being 0.07 per cent and that of cobalt being 0.03 per cent. In the peridotite-gabbro massif of Hulkkonen (about 3.5 kilometers NE of Narila) that rock has a Ni/Co ratio of 2.2, but, in the Hällinmäki ore, that ratio is 2.5. Häkli (1963) suggests that the sulfide phase of the Hällinmäki copper ore came from the same magma from which came the diorites, the quartz diorites, and the trondjhemites. The nickel and cobalt contents in the diorite and gabbro are much lower than in the ores or in the peridotites.

It appears that, in the Hällinmäki ore, silver is one to five gm/ton; where the copper grade rises over two per cent, the silver content rises slightly to seven to eight gm/ton. Gold is randomly distributed through the ore; on the average, it is about 0.1 gm/ton. Molybdenum is in only trace amounts, and zinc is no more than a few fractions of one per cent. Sulfur is no more than three to four per cent in the areas of its highest concentration but averages below one per cent. Iron, excluding that in magnetite, probably is present in amounts of two to five per cent.

Hällinmäki ore is either disseminated or brecciated. The disseminated ore is principally in diopside amphibolite in which it is evenly distributed or confined to specific layers, ordinarily those that are diopside-bearing. In places in which the diopside layers are folded, the quantities of ore minerals in the apices of the folds appears to be somewhat greater than in the remainder of the bed. This may be due to the open space developed in these fold crests over that in the straight portions of the layers. Grain size of the ore minerals ranges from 0.1 to 0.4 millimeters. The principal metallic minerals are chalcopyrite, pyrite, pyrrhotite, and magnetite with chalcopyrite making up over 70 volume per cent of these minerals. Cubanite does occur, is irregularly distributed, and makes up only a few per cent of the metallics. If, however, the disseminated ore is more abundant than normal, cubanite may rise to 20 volume per cent of the total with pyrite being equally reduced. Exsolution bodies of cubanite are common in chalcopyrite. Pyrrhotite composes five to six volume per cent of the disseminated metallics; normally the content of magnetite is below one volume per cent; in the magnetite lenses, however, the magnetite is as much as 60 volume per cent. Traces of sphalerite and pentlandite are present as exsolution bodies in chalcopyrite and pyrrhotite, respectively.

The brecciated ore is mainly in garnet skarn zones or in their immediate vicinity. Veinlets of sulfides five to 10 centimeters across surround the garnet or amphibolite fragments; they also occur in lenses one to three centimeters thick that, in many cases, contain small fragments of the enclosing rocks. Contacts of the lenses with the wall rocks are not always sharp as metallic minerals replace the outer 0.5 to 1.0 centimeters of the fragments. Sulfides also fill narrow fractures in the fragments.

In this breccia ore, the main metallic minerals are chalcopyrite, pyrrhotite, and cubanite with minor pyrite and magnetite. The metallic minerals in the richer ore consist of about 50 volume per cent of chalcopyrite, cubanite 30 to 40 volume per cent, and pyrrhotite 50 to 10 volume per cent. Magnetite is irregularly distributed in small amounts whereas pyrite may be one to five per cent by volume. The minor minerals in the brecciated ore are present in microscopic amounts and normally are in exsolution bodies in chalcopyrite and pyrrhotite; these bodies include: bornite, mackinawite, pentlandite, bravoite, sphalerite, molybdenite, linnaeite minerals, gersdorffite, and millerite. The minerals in the brecciated ore, particularly in the lenses, may be as much as three to four centimeters in diameter. Usually, the exsolution minerals are only 0.01 to three millimeters long.

At Hällinmäki, cubanite is present as an economically valuable mineral only in ore body II; it appears in large intergrowths with the chalcopyrite, with the cubanite being large-sized exsolution bodies. Pyrrhotite also is an exsolution product of chalcopyrite and appears to have come out of solution at the same time or slightly later than cubanite. Pyrrhotite and bornite, an uncommon mineral, do not occur together. The pyrrhotite associated with cubanite in chalcopyrite contains exsolution bodies of pentlandite and sphalerite.

The abundance of exsolution textures in the Hällinmäki minerals and the wide variety of them indicates that the ores crystallized at high temperatures, probably well within the hypothermal range; in the minor amounts of exsolved bornite that were formed by exsolution, minor chalcopyrite seems to have exsolved in turn from this exsolved bornite. Some of the pyrrhotite appears to have crystallized directly from the ore fluid and not by exsolution processes within the chalcopyrite; it is not clear how much of the pyrrhotite was a primary mineral and how much resulted from exsolution; centainly much of it exsolved from chalcopyrite. The exsolution sequence in chalcopyrite and in the minerals exsolved from chalcopyrite is extremely complex with, for example, Co-pentlandite exsolving directly from chalcopyrite and also exsolving from pyrrhotite that exsolved from chalcopyrite. Similarly, sphalerite appears to have exsolved directly from chalcopyrite, cutting across cubanite lamellae, but also exsolved directly from pyrrhotite that exsolved from chalcopyrite. Minor amounts of linnaeite minerals, pyrite, and magnetite seem to have deposited directly from the ore fluids.

The rather complex suite of high-temperature minerals appear to place the deposition of the chalcopyrite (and the associated pyrrhotite) in the hypothermal range, and the deposit should be categorized as Hypothermal-1.

Two minor deposits (Karhuniemi and Lari's, named for a dog trained to sniff out buried sulfides) are located in the general Hällinmäki area, the former lying to northeast and the latter to the northwest of the main ore body. They appear to be reasonably similar to the main deposit of Hällinmäki.

Hyvärinen thinks that the ore fluids probably came from the same portion of the regional magma chamber that produced the quartz-diorite magmas.

HAMMASLAHTI

Middle Precambrian Copper Mesothermal

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Ridge, 1980, Ore formation I - isotopes of sulfur, oxygen, and hydrogen: the view from the bend in the road: 5th IAGOD Symp., v. 1, p. 1-33 (Snow Bird)

Notes

The Hammaslahti deposit is located some 20 km southeast of Joensuu (62°32'N, 29°42'E); it is the newest mine in the North Karelian ore district and the one farthest to the east. It is in the same general area as those of the Outokumpu type, but is about 60 km southeast of that mine and is markedly different from Outokumpu both in the geology of the region in which it lies and in the paragenesis of the minerals. Disseminated chalcopyrite was first discovered on an arkose outcrop in Hammaslahti in 1966; from 1967 to 1970, the Finnish Geological Survey carried out geophysical surveys and core drilling in the area. This work outlined a body of copper mineralization with some zinc, but it was, because of the low grade estimated by the Survey, considered to be of no more than marginal economic interest.

In 1971, however, the property was transferred to the Outokumpu Oy for the preparation of feasibility studies and for a re-estimation of the reserves of ore. From this work it was estimated that the ore would run about one per cent copper and that only one concentrate (copper) was worth making. Early in 1972, work was begun on preparing the body for mining, and mining actually began, by open-pit methods, only one year after work was begun on the mine. This ore is recovered from two open pits, both having lain under shallow over burden. At the same time, work was commenced in preparation for underground mining by longitudinal sub-level stoping.

The design of the concentrator permits the processing of 400,000 tons of ore per year, and concentrate production in 1974 was 14,000 tons. The ore mined in that year averaged about 0.76 per cent copper; 0.1 per cent zinc and 3.8 per cent sulfur. The mean copper content of the ore, however, is estimated at about 1.1 per cent with a total copper content of the deposit being between 40,000 and 50,000 tons of metal.

The major structures of the Hammaslahti region are two synclines, the Höytiäiinen (east) and the Pyhäselka (west). Between these two basins is an anticlinorium, and the ore bodies are located in a secondary syncline in that structure. So far as the principal synclinal basins are concerned, their main rocks are Kalevian schists that Isokangas (1978) believes to have been deposited in a subsiding geosyncline; these schists show the effects of strong turbidity currents and contain well-developed graded bedding. In the anticlinorium, the most important rock types are metaarkoses, para-conglomerates, phyllites, and dolomites. Within the ore-containing secondary syncline is a 15-km-long zone of impure arkose and graywacke that contains the ore and is bordered on its west flank by a narrow band of black schist and, beyond that, by phyllite; to the east the arkose abuts directly against phyllite. These phyllites strike N20°W and dip 80°W; their fold axes plunge from 25° to 40° S. Intercalated in the phyllites are quartzites, black schists, arkosite, and carbonate concretions; the schists also contain porphyroblasts of staurolite, andalusite, and garnet. Thus, in the broad sense, the ore-bearing arkose is simply an intercalation in the dominant phyllite. Hyvärinen and his colleagues say that the rocks in the anticlinorium are lower Karelian in age, but that they are underlain by marine Jatulian black schists, the oldest rocks known in the region.

The arkose, the country rock of the ore, is quite heterogeneous in both grain size and composition and in quantities of feldspar and quartz it contains. During the formation of the ores, the country rock underwent intense silicification so that quartz is most abundant nearest the ore and decreases consistently away from it.

The marked differences in grain size mean that the rock ranges from fine-grained to conglomeratic; the conglomerates are composed of pebbles of several kinds of rocks (polymictic) and the matrix material is abundant. Intercalated with these conglomerates are, most commonly, black schists and phyllites, although interbedded skarns, carbonates, and amphibolites also are present. Although tectonic activity has been strong, the rocks have been metamorphosed only to a slight extent. No evidence of granitization, such as pods or dikes of granite, are to be seen.

The Hammaslahti ore bodies are located in a shear zone that cuts the western edge of the minor syncline in the Tohmajärvi anticlinorium. This shear zone is at least 15 km long, has a width of 2.5 to 3 km; its axial plane trends at N20°W and its axis plunges at 25° to 30° generally south. Locally, the syncline has been isoclinally folded. The copper ores are along the margin of the sheared arkosite on the west flank of the anticline and are close to the contact between arkose and phyllite. The ore bodies are located in a left-handed en-echelon fold pattern that plunges to the south, and their dip parallels that of the fold axis, and the plunge is parallel to the fold axis (30°S). The ore bodies run parallel to each other (Hyvärinen and his colleagues, 1977, say there are eight of them; Isokangas says four) and to the lineation of the shear zone (N20°W). The ore outcrops at the northern end of the series of ore bodies, and the most southerly of the line of ore bodies reaches a depth of 500 m.

Hyvärinen and his colleagues say that tectonic (dynamic-metamorphic) effects have segregated the micas and feldspars plus quartz into separated bands over much of the area of the shear zone. Much of the plagioclase feldspar has been replaced by quartz in the arkosites and tremolite by quartz in the amphibolites. Hyvärinen and his colleagues believe that the crystallization of the gangue minerals has been in the following order - albite, biotite, chlorite, quartz, calcite, and/or siderite with fluorite. Some scheelite also is present; the albite and fluorite are classed as accessory minerals. Quartz normally bears fluid inclusions and, in a few places, rutile needles.

Isokangas says (1978) that the Hammaslahti orefield consists of four overlapping ore bodies. Hyvärinen and his colleagues, however, say that the shear zone contains eight ore bodies that run parallel to one another and to the lineation of the shear zone. Further, they say that the zone was traced by drilling for 1.5 km and probably extends for 8 km. On the contrary, Isokangas quotes the mine geologist as saying that the arkose and graywacke zone intercalated in the phyllites is 15 km long; he does not say if the shear zone extends for the full 15 km or not. Both authorities agree that the ore was emplaced in a breccia in which the sulfides are present in veins and stockworks in the arkose; there appear to be places (Hyvärinen, *et al.*) where disseminated sulfides are quite abundant. Normally, the breccia ores are of relatively high grade, the disseminations of low. The boundaries of the ore are gradational with some sulfides being found throughout the 50 m width of the shear zone, but far from all of the shear zone is ore. Stoping is guided by assay boundaries.

Hyvärinen, *et al.* say that the northern part of the ore field contains two ore bodies and the southern six with their dimensions and respective metal contents covering a wide range. They say that the largest body in the southern sector is up to 200 m wide, 20 m thick, and 700 m long.

The principal ore mineral is pyrrhotite with chalcopyrite being the most abundant (and essentially the only) ore sulfide. Sphalerite and pyrite are quite common in the northernmost ore body, but sphalerite is rare in the other ores. Lead and the precious metals are low; the nickel content normally is less that 80 ppm and that of cobalt less than 450 ppm. Selenium is quite high in relation to sulfur. Mackinawite, cubanite, and tetrahedrite-tennantite occur in exsolution bodies; arsenopyrite is exceedingly rare. The order in which the minerals crystallized is (Hyvärinen, *et al.*) pyrrhotite, sphalerite, pyrite, chalcopyrite, and galena, although all five of these sulfides essentially are never found together in appreciable amounts. The sulfides, together with secondary quartz (primary as far as the veins and veinlets are concerned), are present largely as fracture fillings in which the sulfides crystallized after the gangue minerals. Three directions of fracturing are recognized in the ore zone, E-W, N-S, and N5° to 15°E. The E-W fractures appear to have been formed before the other two sets.

The average content of copper in the northern part of the deposit was determined by Hyvärinen and his colleagues as being 0.6 per cent; in the southern part they thought that content to be about 1.0 per cent. Isokangas reports, however, that the mean copper content of the deposit as a whole is 1.1 per cent. The average zinc content (Hyvärinen, *et al.*) is 510 ppm and ranges between 100 and 3600. In what these authors describe as the surface section (600 m^2), they report the zinc concentration as almost 3.5 per cent. The lead content in the ore averages only slightly above that of the surrounding arkose. The Ni:Co ratio is below 1 in the ore but above that in the non-ore arkoses. Silver averages only 6.6 gm/ton which appears to be uneconomic even at 1982 prices. The silver content is directly related to that of copper which causes Hyvärinen and his colleagues to assume that it is in the crystal lattice of the chalcopyrite, but how it is emplaced there they do not indicate. Gold is very unevenly disseminated in the mineralized rock volumes and is of no economic value.

Hyvärinen and his colleagues have studied the relationships between the metals in the ore bodies proper and the surrounding rocks. They report that the ore bodies have haloes from 70 to 100 m in width in which copper concentration is, of course, lower than it is in the ore but also lower than it is in the rocks beyond this 70 to 100 m halo. Thus, the copper anomaly around the ore bodies is a negative one. Cobalt behaves, on a much lower ppm level, much as does copper. Nickel, on the contrary, is present at about 20 ppm in the ore volumes, but in the hanging wall the nickel runs 86 ppm and in the arkose 44 ppm. The ore body, therefore, is enclosed in a 30 to 40 m-wide zone in which nickel is enriched from 0.6 to 1.0 times. Iron also seems to exhibit a negative anomaly, particularly on the hanging-wall side. Lead and zinc are neither enriched nor impoverished in the host rocks, but one series of drill core contains 10 ppm more zinc and 2 ppm more lead in the 20 to 70 m portion than it does in the 70 to 150 m, and the concentrations of both of these elements rise to a maximum in the ore bodies themselves. These authors conclude that these metal aureols are the results of remobilization of primary, sedimentary sulfides (Ni, Co, Cu, and, to a lesser extent of Fe, from the host rocks to in the shear zone. They explain the positive nickel aureole by nickel moving toward the shear zone during ore formation but having been picked up before reaching that zone by replacing magnesium in such silicate minerals as existed outside the shear zone. They believe that copper and cobalt show negative anomalies immediately outside the ore zone because both of these elements were not nearly as likely to replace magnesium as nickel and survived to reach the ore zone without entering the available silicates encountered on their way there. The gradual change in nickel, copper, and cobalt contents beyond the negative or positive anomalies (as the case may be) indicates, to Hyvärinen and his colleagues, that nickel moving in rose in the immediate environment of the ore zone whereas copper and cobalt dropped. Actually, examining their Figure 8, it appears to me that the changes in copper, cobalt, and nickel contents can equally well (or even better) be explained by assuming that the bulk of the copper and cobalt were deposited hydrothermally in the ore zone with little of either metal being left to be deposited outside the ore zone. On the contrary, nickel did not deposit in as high a proportion in the ore zone from the hydrothermal fluid as did copper and cobalt but that nickel did deposit to a considerable extent immediately beyond the ore zone. This probably resulted from the reason advanced by Hyvärinen and his colleagues, namely that it replaced magnesium in silicates. The relatively flat curves for all three elements away from the vein probably represent the original amounts of these elements in the sediments involved.

Certainly, the so-called positive and negative anomalies do not prove beyond any doubt that the metals came from the country rock into the ore zone or from the ore zone into the country rock, but I think the probabilities favor the latter over the former concept.

Fluid inclusions in the ores were studied by heating under the microscope. By this method, the temperatures at which the inclusions ruptured were measured. It was found that most of what Hyvärinen and his colleagues call secondary inclusions ruptured at 310°C. These "secondary" inclusions are those found in healed microfractures that generally radiated out from ore mineral microveins into quartz. These fractures were filled by quartz, "secondary" inclusions, and chalcopyrite. Apparently these microfractures were formed during the ore-emplacement process, so the decrepitation temperatures give some idea of the temperature at which both the inclusions and the ores with which they are indirectly associated were formed. Of course, the temperature at which the inclusions rupture is dependent not only on the gas pressure generated in them by heating but also on the strength of the quartz which they were entrapped. The average temperature of rupture of most inclusions (310°C) almost certainly is too high as the strength of the quartz would resist breaking until some temperature above inclusion-filling was reached. Even allowing for this, however, the actual temperature of inclusion-filling must have been over 250°C and perhaps not much below 300°C. If these temperatures are guides to the temperatures at which mineralization took place, as they almost certainly are, the temperature of vein formation appears to have been far too high to have been achieved through dynamometamorphism unaided by hydrothermal solutions (perhaps in the broad sense) at much higher temperatures than those produced in the enclosing rocks through "remobilization" processes.

Hyvärinen and his colleagues believe that the various ions that they determined to have been present in these "secondary" inclusions indicate that all the elements necessary for the formation of the ore minerals were present in the solutions that were the parents of those entrapped in the "secondary" inclusions. They reason that these solutions were generated in the wall rocks surrounding the shear zone and carried there through numerous microfractures that existed in the host rocks to the much more impressive fractures in the shear zone where deposition of the ore and gangue minerals (mainly quartz) took place. The high temperature of ore deposition, however, would seem more probably to indicate that the ore fluids rose through the shear zone, deposited their burden of ore and gangue minerals and then, to some extent, worked their way out into the wall rocks, perhaps to deposit minor additions to the wall-rock content of copper, cobalt, and to a great extent, nickel. The concentrations of lead and zinc outside the ore zone behave somewhat differently than those of the other metals. Hyvärinen and his colleagues say that lead and zinc are not enriched or impoverished in the host rocks, but they do, of course, rise to a maximum in the ore zone. This relationship can be interpreted to mean that the "remobilizing" solutions were essentially incapable of remobilizing lead and zinc or that the upward-ascending ore fluids deposited practically all of their meager load of lead and zinc in the ore zone and had little ability to add these metals to the wall rocks (or to subtract these metals from them). Obviously Hyvärinen and his colleagues favor the first possibility and I the second.

Hyvärinen and his colleagues studied the composition of sulfur isotopes in the ores and the wall rocks from 34 samples. Their delta-^{34}S values range from +1.4$^{\circ}/_{oo}$ to 14.2$^{\circ}/_{oo}$, their arithmetic mean being 5.5$^{\circ}/_{oo}$. The delta-^{34}S values of the ore sulfides have a mean of 3.8$^{\circ}/_{oo}$ with a standard deviation of 0.8. Those sulfides outside the ore zone have a mean delta-^{34}S value of 6.3$^{\circ}/_{oo}$ and a standard deviation of 2.6. [Should I emphasize that standard deviation is a statistical measure of dispersion? I suppose not in these days.] Hyvärinen and his colleagues adopt an estimate of 14.6$^{\circ}/_{oo}$ for sea water sulfate contemporaneous with the original sedimentary precipitation of the sea-water sulfides. They believe this confirmed by the delta-^{34}S value of 14.2$^{\circ}/_{oo}$ determined from pyrrhotite in the coarse-grained arkose on the footwall side of the ore body. They think that this delta-^{34}S value for this pyrrhotite confirms that the pyrrhotite was deposited with its sulfide sulfur having come from sea water sulfate with a change of only 0.4$^{\circ}/_{oo}$ in the process of reducing it from the S^{+6} to the S^{-2} state. Certainly, the fractionation in the ore-forming process should have been much greater than this! On the contrary, Ridge (1980) suggests that the reactions that occur in a hydrothermal fluid, on its way upward toward the surface of the earth or the bottom of the sea floor, easily can convert the S^{-2} of primordial sulfur from a delta-^{34}S value of 0.00$^{\circ}/_{oo}$ to one even higher than +14.2$^{\circ}/_{oo}$. The mean delta-^{34}S value of 3.8$^{\circ}/_{oo}$ of the ore sulfides in the Hammaslahti ore bodies is well within the ability of the reactions that affect hydrothermal fluids (and the rock through which they pass) to achieve. In short, the delta-^{34}S values of the sulfides in the Hammaslahti ores is more favorable to their having been derived from appreciable depths and have not, in general, been nearly as heavily oxidized as to bring the bulk of them to the delta-^{34}S value of Precambrian sea water. Further, it is difficult to conceive of sea water sulfate sulfur having been converted from a delta-^{34}S value of 14.6$^{\circ}/_{oo}$ to one no lower than 14.2$^{\circ}/_{oo}$ for the resulting S^{-2} in pyrrhotite. Whereas it is much more reasonable to suppose that, on being involved in the proper reactions, some S^{-2} could have been increased in delta-^{34}S value from +3.8°/oo to +14.2°/oo (Ridge, 1980, p. 17)

For remobilization of the type envisioned by Hyvärinen and his colleagues, three mechanisms may be employed: (1) the plastic flow of sulfides, (2) solid-state diffusion, and (3) the formation of metahydrothermal solutions. These authors base their ideas of the genesis of the Hammaslahti deposits largely on the third of these mechanisms. They point out that the isotopic composition of the sulfur in a metahydrothermal fluid will have the average value of the delta-^{34}S values of the sulfides dissolved in that fluid. Their reasoning is that an apparent negative correlation (correlation factor = -0.593) prevails between the delta-^{34}S and the sulfur content. Both results support, in their opinion, the theory that once-disseminated synsedimentary sulfides were remobilized to form the ore.

In favor of the first two mechanisms, they argue that the plastic flow of sulfides and solid-state diffusion both are affected by PT gradients; therefore, the remobilization grade and the grade of metamorphism should be to some extent analogous. On the basis of this and on their finding that in biotites a decrease in Fe^{+2} and an increase in Mg^{+2} can be correlated with increasing grade of metamorphism, they made microprobe analyses of biotites and chlorite. They reason that the similarity (shown in their Fig. 14) between the sulfur isotope patterns and the FeO:(FeO+MgO) pattern accordingly implies that remobilization was of importance in concentrating sulfides to form the ore. They interpret work by Mäkelä on Outokumpu and Vuonos to mean that the depletion of ^{34}S is a function of the distance from the ore proper, a relationship that does obtain in the Hammaslahti ores.

Hyvärinen and his colleagues agree that the Hammaslahti copper deposit lies in a transverse shear zone that is several kilometers long and that the ore has a distinct time-place relationship to the shear tectonics, that is, the ore must have been emplaced after the shear zone had been created as the epigenetic relations of the ore to the fractures and brecciated shear zone definitely show. They follow this with a statement that I cannot understand, namely that "the positive nickel aureole surrounding the ore indicates primary origin also for the negative aureoles of copper and cobalt because the geochemical behavior of nickel and copper should be similar and thus they should both enrich into secondary geochemical aureoles."

These authors quote Kouvo's lead age for the Hammaslahti deposit as being 2300 m.y. This cannot be the age of the shear zone, however, because the marine Jatulian formation is only 2050 m.y. old. Possibly the ore lead represents, they suggest, the original age of the Karelian sediments. The 2300 m.y. age of the Hammaslahti lead also supports their idea that the ore metals and sulfur of the Hammaslahti ores was provided by the surrounding schists. It might equally well mean that the hydrothermal fluids included 2300 m.y. old lead that was present in the (probably anatectic) magma chamber in which those fluids were developed.

These authors contend that the ore minerals must have been formed by fluids that were no lower than 310°C in temperature, whereas, I think this determination means that the temperature of the ore fluids at the time they were entrapped in the fluid inclusions in the gangue quartz were no higher than 310°C and probably appreciably lower because of the force needed to rupture the quartz containing them.

I certainly agree with their concept that the Hammslahti ores are epigenetic in relation to the schist zone in which they are found, but I do not believe that any of the points they raise demonstrate certainly that the ore metals and sulfur came from the schists that enclose the shear zone.

From the probable temperature of deposition of the ore minerals and the lack of any high-temperature silicate skarn associated with them, I suggest that the ore was formed under mesothermal conditions and should be so classified.

Isokangas is far less positive about the genesis of the Hammaslahti ores than are Hyvärinen and his colleagues. He points out that the Hammaslahti ores are a type new to North Karelia and that the data about them are scanty. They do say, however, (which hardly can be denied) that the ore exhibit epigenetic features and evidently crystallized under hydrothermal conditions. They say nothing about where the hydrothermal fluids came from or how they acquired their metals and sulfur (probably the wise thing to do at the present time). Isokangas points out that the lead-model age for Hammaslahti is 2300 m.y. (the same as Outokumpu) but that the age of the sedimentary schist that contains the shear zone is 2100 m.y. so, whatever else the lead model age may mean, it does not say that the ores are 2300 m.y. old in their present environment.

HITURA-MAKOLA - KOTALAHTI

Middle Precambrian	Nickel, Copper	Magmatic-2a Magmatic-2b Metamorphic-C

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Notes

The Kotalahti and Hitura (and the exhausted Makola) mines are separated by 195 kilometers in a belt of nickel mineralization that extends for about 430 kilometers from slightly north of Ylivieska (about 64°4'N, 24°26'E), which town is some 35 kilometers from the Gulf of Bothnia, to Parillala (61°35'N, 29°30'E) near the border with the USSR. The Hitura-Makola mines are about 40 kilometers southeast of Ylivieska; the Kotalahti mine is about 40 kilometers south of Kuopio (62°54'N, 27°40'E). About 20 prospects, in addition to the three mines named, are known in this belt of which the largest are Laukunkangas (about midway between Kotalahti and Parikkala), Ylivieska, and Parikkala. The Makola mine is only about four kilometers south of Hitura.

Actually, it is possible and has been suggested that the Hitura-Kotalahti nickel belt is part of a larger arrangement of nickel-copper deposits in southern Finland. It appears possible to integrate the Hitura-Kotalahti belt into the Ahlainen-Sääksmäki nickel belt by curving south and southwest through Laukankangas, Kitula, and Telkkälä to connect with the Ahlainen-Sääksmäki belt in the vicinity of Kylmäkoski. From Ahlainen the rude circle of nickel deposits follows the shore of the Gulf of Bothnia north and then northeast to Långbacka (near Korsnäs) and then back southeast to the Hitura-Makola area.

This rough circle surrounds the granitoid rocks of central Finland, and the deposits scattered around its circumference all are associated with ultramafic-mafic rocks that have ages that lie in the rather narrow range of 1925 to 1850 m.y. ago. All of these deposits, mutatis mutandis, also appear to have been formed in much the same manner. Thus far, no full explanation has been given as to why the parent intrusions are located on the circumference of this circle or why so many of them contain mineable bodies of nickel-copper ore.

The oldest rocks, in the complete circle or in the narrower confines of the Hitura-Kotalahti nickel belt, are migmatitic ortho- and para-gneisses (also denoted as granite gneiss by some authorities) that contain large numbers of metadiabase and amphibolite veins; this group makes up the pre-middle Precambrian basement in the belt. The basement rocks underlie much of the area and extend upward as numerous brachyanticlines and domes that are surrounded by younger schists. The rocks of the basement are dated as from 2,800 to 2,600 m.y. old. The immediately overlying Svecofennian rocks are quite thin, from a few 10's of meters to 200 meters, and are an epicontinental series that is made up of the products of the weathering of basement rocks such as weathering breccias and saprolites. Overlying these rocks in many places are glassy othoquartzites, above which are calc-silicate rocks that contain intercalations of carbonate rocks (Gaál, 1972). Above the epicontinental series are mica gneisses that contain graphite and sulfide and are thought to have been saprolites before they were metamorphosed. These gneisses are the transitional rocks between the epicontinental series and a geosynclinal facies; the rocks of the eugeosynclinal facies were metamorphosed pelites and psammites and are now migmatitic veined gneisses that contain thin intercalations of calc-silicate material. Outside the nickel belt, the basal part of the geosynclinal series contains many bodies of serpentinite that may be ophiolites; these rocks occur with black schists, metamorphosed cherts, and calc-silicate rocks and dolomite. Volcanic rocks also make up a major portion of the upper part of the eugeosynclinal facies; mafic to acid lavas are present that range from amphibolites with pillow lava structures to leptites. During and after this last volcanic activity, erosion produced conglomerates with pebbles of the volcanics and of the gneisses as well. The unconformity marked by this conglomerate also indicates the beginning of a period of synorgenic sedimentation, with the process resulting in metaturbidites that rapidly accumulated in flysch geosynclines (Gaál, 1972).

The cycle of igneous intrusion with which the nickel ores are associated began during the orogenic period, the first to be introduced having been mafic to ultramafic plutonic rocks. Determinations on these rocks indicate that they were emplaced about 1925 million years ago. These were followed by additional syntectonic and by late tectonic intrusions, most of which entered the area about 1800-1900 million years ago (Kouvo and Tilton, 1966). These later rocks are variously described as granodiorites or granites.

The mafic-ultramafic rocks of the Kotalahti area are made up of a differentiation series that ranges from olivine peridotite to quartz diorite; the main portions of this igneous mass are pyroxenites, perknites (a term not now commonly used but employed for rocks in which the main components are clinopyroxene and amphibole, with accessory orthopyroxene, biotite, iron ores, and essentially no feldspar), and hornblendites. The parts of the igneous mass composed of olivine-bearing pyroxenites, on the one extreme, and gabbros and diorites on the other, are much smaller in size than the pyroxenites, perknites, and hornblendites. Completely olivine dunites are most uncommon, and the pyroxenites are mainly lherzolites (olivine, clinopyroxene, and orthopyroxene, in which olivine normally is the most abundant mineral) and harzburgites (mainly olivine and orthopyroxene). Most of the pyroxenites are fairly common near-surface, but drop off appreciably with depth (Papunen, 1970). The bodies of granite in the area are not large but small amounts of granite bear intrusive relations to the ancient gneiss complex. Massive bodies of granite generally have diffuse boundaries, but bosses, dikes, and sills with sharp boundaries also occur; much of the granite is aplitic, but the grain size ranges up from medium to pegmatitic.

The central igneous feature of the Kotalahti district is an elongated mafic intrusion; this mass outcrops in four distinct and separate bodies, and the three most northerly of these (Mertakoski, Välimalmio, and Vehka) join not far beneath the surface. The most southerly body (Huuhtijärvi) is (at least to a depth of 700 meters) separate from the others. A fifth body of ore (Jussi) is east of, and essentially parallel to Vehka; it is not associated with ultramafic rocks. All of these intrusions are composed of peridotite, perknite, gabbro, diorite, and quartz diorite. The individual igneous bodies are irregular in form and their inter-relations are most complex; mapping these bodies in any satisfactory manner has been difficult and often impossible. The relationships of one ultramafic or mafic igneous body to another is complicated by the presence of much amphibolite and granite contained in, traversing, and brecciating the more mafic rocks;

some of the mafic rocks occur as inclusions in amphibolite and granite. Much of the amphibolites is present in the mafics as dikes or lenticular bodies, but some lies on the peripheries of mafic masses. Granite shows similar relationships to the mafics. In contrast to Papunen, Haapala refers to the peridotite as saxonite, but this does not matter much as harzburgite and saxonite are essentially synonomous. The olivine is largely serpentinized, with appreciable amounts of amphibole, mica, and iron oxides and sparse talc and carbonate. The altered perknites are made up of amphibole as the major component, with minor cummingtonite and anthophyllite and locally abundant pyroxene; in places, the perknite grades into gabbro.

The igneous complex at Hitura consists of two bodies of serpentinite, the north and south that are separated by a narrow gneiss band. Ore mineralization is confined to the northern margin of the northern serpentinite body. Both of these bodies were covered by thick Quaternary formations, and they were discovered only through geophysical surveys and diamond drilling. The altered ultramafic plugs are contained in banded migmatites that locally grade into banded or porphyritic amphibolites. Although some sulfides are known in the gneisses, they are quite different from the ore sulfides and cannot have formed in the same way. Along the contacts between gneisses and ultramafics, cordierite, staurolite, anthophyllite, biotite, and graphite have been developed.

The ultramafic body containing the ore has a core of massive serpentinite, in which the serpentine is antigorite and with which accessory chlorite, talc, carbonates, magnetite, chromite, and sulfides and abundant finely disseminated graphite are associated. Surrounding the serpentinite is nonhomogeneous peridotite composed of varied quantities of olivine, enstatite, augite, and amphiboles. Near contacts or in shear zones, the peridotite has been altered to an essentially all amphibole or an amphibole-chlorite rock that in places contains talc. If the olivine has been serpentinized, the orthopyroxenes generally have been altered to commingtonite or anthophyllite; the amphiboles are secondary in respect to the pyroxenes or the olivine. The highest-grade sulfide development is most abundant in the zone in which serpentinite grades into peridotite. Some coarse-grained pegmatite veins interest the serpentinite, and reaction rims occur between the two rock types. Near the contact zones, the serpentinite is much fractured, and the fractures may be filled by chrysotile or centrally located sheared chlorite, bordered by anthophyllite, actinolite, and talc (Papunen, 1970).

The first indication of the Makola body was given by two nickel-copper-bearing serpentinite blocks found in the area in 1936. Later several anomalies were found by electric and magnetic surveys. The main rock type around the serpentinite is a gneiss-like mica schist that locally is distinctly migmatitic; the schistosity is clear and strikes about N30°E and dips steeply on either side of the vertical.

The serpentinite body that contains the ore is a flattened cone that tapers downward. In 1954, the cone was known to reach at least 180 meters below the surface. Since the mine is now worked out, a more exact figure on the depth reached by the serpentinite obviously is known, but I have found no statement of it in the literature. The average horizontal length of the serpentinite mass, in a generally north-south direction, is 120 meters, and the width ranges between seven and 20 meters. The various structural characteristics of the cone, longitudinal axis, lineation, and plunge, correspond to thos of the surrounding mica schistose gneiss. The contact between schist and serpentinite is sharp and quite regular. In places, all the serpentinite is ore, in others, the ore occupies only a fraction of the serpentinite; locally, sulfides enter into the bordering schist (gneiss) (Huhta, 1954).

The structure of the nickel belt is fractured by a large group of long, essentially straight, faults, the lines of which strike more or less northwest-southeast. The dominant fault in the nickel belt is the Haukivesi fault; it has a right-handed displacement, a N40°W strike, and can be traced for 340 kilometers. Two other major faults are closely parallel to the Haukivesi; they are the Suvasvesi (to the northeast) and the Pihlajavesi to the southwest. They also have a right-handed displacement, and, although they are not as continuous as the Kaukivesi, they are both impressive and almost equally long faults. In addition, the area contains a large number of minor faults that are of the same type and strike direction as those that surround the major structures. Thus, Gaál (1972) refers to the major faults as shear zones rather than as single fault planes; these shear zones may be as much as several kilometers wide, and their tectonics are complex and the varieties of rocks they cut are many. Much folding accompanied the faulting with earlier folds being distorted and superposed by later vertical drag folds. The displacement along these faults can only be approximated, 20 kilometers for the Suvasvesi fault, 10 kilometers for the Haukivesi, and 25 kilometers for the Pihlajavesi. The Haukivesi and Pihlajavesi faults are cut by a younger strike-slip fault that has a left-handed lateral displacement of about 20 kilometers. This Kerimäki fault is younger than the northwest-southeast faults and results from a different stress field; otherwise it is quite similar in its effects on the rocks it cuts and also contains similar mafic and ultramafic intrusions and nickel-sulfide ore or mineralization.

The various igneous intrusions were introduced along the shear zones in several stages. The first were the ultramafic-mafic rocks; in many instances, these intrusives were broken and the fragments dragged along the faults. Brecciated bodies that contain fragments of these ultramafic-mafic rocks may have silicic migmatitic material as a matrix or it may be composed of the same material as the fragments (Gaál, 1972). From these fault, fold, and intrusive relationships, it follows that all three occurred over a considerable time interval with intrusions taking place both before and after the left-handed displacement faults, and folds accompanying and following the right-handed displacement faults. The ore appears to have been emplaced both before and after the left-handed faulting.

The fold systems of the area were produced by the same tectonic forces that caused the faulting; the fold patterns resulted from at least three stages of folding, with the latter two being the more directly related to the faulting. The effects of the first folding stage largely have been obliterated by the other two, and of this first stage essentially nothing is known. The second stage (the first related to faulting) produced east-west striking antiforms and synforms, with steep to vertical axial planes. In the southeast part of the nickel belt (that is, in northern Karelia) the east-west strike changes to a northeast-southwest one. The strike of these structures is that of the Svecofennian deformation phase (in the narrow sense). The third stage developed folds having an essentially consistent north-south strike; these folds also change strike in the southeast part of the belt, in this instance to a north-northwest to south-southeast direction and is called the Karelian phase deformation although in the nickel belt it affected Svecofennian rocks. The superposition of the third-stage folding on that of the second converted the Svecofennian antiforms and synforms to domes, brachyantiforms, and basins.

From the close interrelationships of faulting, folding, igneous intrusion, and ore-body formation, it appears that all of these effects were produced in geologically contiguous time. Thus, although these events took place over perhaps as much as 200 million years, all of them can be dated as late middle Precambrian and are so categorized here.

Although the general principles that govern the formation and location of the nickel-copper ore bodies are the same in each of the three districts, each is different enough from the others to justify enough description to show how they are similar and how different.

In general, Gaál says (1972) the following facts are significant in controlling the nickel-copper mineralization in the nickel belt: (1) the ores occur almost exclusively in ultramafic to mafic plutonic rocks; (2) sites of the intrusives containing the ores are controlled by steep strike-slip faults; (3) the ore-bearing intrusives are localized in domes and brachyantiforms that were developed at the same time as the fault-shear zones; this localization by folding is most characteristic of the northwest portion of the belt; and (4) interaction between the mafic magmas and the country rocks was needed to provide much of the sulfur that is now in the nickel-copper-iron sulfides. Later, it will be shown that this last factor (4) is probably not as important as Gaál and Häkli (1971) thought it was.

At Kotalahti, there are five major ore bodies; four of these lie along a general, 1,000-meter-long northeast-southwest line - Mertakoski (NE), Välimalmi, Vehka, and Huuhtijärvi (SW) - the Jussi ore body is located 150 meters due east of the southern part of the Vehka ore body. The Huuhtijärvi body contains a much higher proportion of massive to disseminated sulfides than do the others in the associated rock masses constitutes ultramafic-mafic rock complex (Gaál's perknite). The Huuhtijärvi is an almost vertically dipping pipe. The dip appears to be generally steeply north. This ore body was known, in 1970, to reach a depth of about 700 meters. The Vehka, Välimalmi, and Mertakoski ore bodies join at shallow depth. On the 150-meter level, the Vehka ultramafic-mafic mass is thin and elongate in the northeast-southwest direction, with the massive, pipe-like bodies of ore being irregularly scattered through the igneous rocks but with most, but not all, of the ore being along the contacts with pegmatite, gneiss, and amphibolite. The Mertakoski ore body, at least on the 150-meter level, does not have the elongate character of the Vehka and Jussi ore bodies, but it is an irregular mass of dense sulfides, and what elongation it has is in an east-southeast to north-northwest direction; the ore body and the ultramafic mass that contains it dip steeply north. The Välimalmi ore body and its containing mafics are quite narrow at the surface but widen considerably with depth, but certainly by a depth of 400 meters, if not higher, it has joined the Mertakoski and Vehka mafic bodies. The Jussi ore body appears to have much the same form as the Vehka, but it probably extends to a greater depth than does Vehka.

At Hitura, the ore is in a tube-like body at the northern end of the north serpentinite mass and is near the contact of the serpentinite tube with the enclosing rocks, these being banded migmatites that locally may grade into banded or porphyritic amphibolites. The adjacent country rocks (mica gneisses) at the western border of the ultramafic mass contain sulfides, but these are much different from the sulfides in the ore, and the sulfides in the two rock types probably are genetically unrelated to each other. The ore dips nearly vertically in the ore and ultramafic pipe.

At Makola, the ore body forms a flattened cone, with its taper downward; the longitudinal axis, the lineation, and the plunge are in accord with the strike and dip of the surrounding country rocks (mica schists and gneisses). At the surface, the ore occurs in three irregular masses that are arranged roughly from north to south. Not far beneath the surface, at least the two northern ore bodies coalesce (and the third probably does also) into a single cone. The contact between the ultramafic body and the country rocks is quite sharp and fairly regular. Usually, the limits of ore correspond to those of serpentinite, but the serpentinite is not always ore; sulfides do penetrate into the country rocks.

At Kotalahti, all the ore bodies, except one (Jussi) are composed of parts of the ultramafic-mafic masses, the boundaries of the bodies cutting across both the structural and compositional features of the rocks. Mineralization below ore grade, even though uneconomic, still is located in the mafic rocks.

The first ore type is composed of sulfides disseminated through the various differentiates of the ultramafic magma, with essentially all rock types being ore in places, but the granite and amphibolite seldom contain disseminated sulfides, even in their inclusions of rocks more commonly sulfidized in more massive bodies. Typically, the disseminations are made up of sparse coarse specks and blebs, up to five millimeters in diameter. These are scattered mainly in the interstices among the silicates but also may intersect, or be enclosed in, silicate minerals. Minor shears and cleavages in the silicates also may contain sulfides. The intensity of dissemination differs widely from one place to another with the extreme case being where coarse sulfides unite into a continuous phase that contains silicate islands or even into massive sulfide veinlets and larger, more irregular bodies.

In contrast to the disseminations, sulfides may occur in minor masses, as breccia cements, and in stockwork veinlets. These more massive sulfide areas are found, normally but irregularly, along rock contacts, particularly where the ultramafic-mafic rocks cut across amphibolites and granites. In the latter rocks, the sulfides form veinlets and surround breccia fragments, and sulfide veinlets are associated with quartz-plagioclase veins that they may in part replace or for which the sulfides substitute. Thus, massive sulfide material grades in two directions, one into disseminations in ultramafic-mafic rock types and the other into fractures in the granites and amphibolites that surround the ultramafics; sulfides also appear to assimilate amphibolites (Haapala, 1969).

The Jussi ore shoot is located in a tongue of the quartzite rocks, and this massive sulfide body is not associated directly with any mafic body. This shoot is controlled by fractures in the quartzite and appears to have been introduced from a mafic magma in which all the silicates had crystallized, leaving only molten sulfides to be forced out by a fairly late-stage burst of tectonic activity. These sulfides were emplaced as major-sized fracture fillings, minor stringers, and very minor disseminations in broken quartzite (Haapala, 1969).

As is to have been expected, the principal minerals in the Kotalahti ores, disseminated or massive, are pyrrhotite, pentlandite, and chalcopyrite; accessory minerals include sphalerite and gersdorffite, which are found in most specimens. Skeletal crystals of pyrite occur separately, and locally molybdenite has been seen. Alteration products include mackinawite, violarite, and melnikovite. In places, graphite is associated with the sulfides, and some minor magnetite accompanies the sulfides.

The Jussi ore body is unusual because it contains considerable millerite and bornite; the bornite, as would seem reasonable, is associated with chalcopyrite. In addition, this ore body contains more pyrite than the others, and the pyrite is higher in cobalt than elsewhere and is present in rather large patches.

In disseminated ore, the disseminated blebs are normally composed of a single crystal of pyrrhotite, containing varied amounts of pentlandite and chalcopyrite. Locally, leaf-like lanellae of troilite are found in the pyrrhotite; this is particularly true in peridotites and olivine gabbros. Most chalcopyrite and pentlandite are located as scattered grains along the rims of pyrrhotite individuals, but some of each occurs in veinlets in cross fractures or in tabular bodies parallel to pyrrhotite cleavage. Feather- or flame-like inclusions of pentlandite are present in pyrrhotite along contacts and fractures.

In the more continuous sulfide phases, pentlandite is found in coarse veinlets between pyrrhotite grains, their sinuous trends paralleling the boundaries of pyrrhotite crystals. Small sulfide individuals in this ore much resemble the blebs of the disseminated type and are composite groups of sulfide minerals. Chalcopyrite is mainly in veinlets controlled by fractures and cleavage planes and is quite abundant.

In actual massive ore, the sulfide material is principally an irregular mosaic of pyrrhotite crystals; pentlandite is either intergranular or as individual crystals. Tongues and islands of pentlandite also occur in the pyrrhotite. Chalcopyrite is associated with pentlandite as veinlets and as fillings in cleavage fragments of the latter mineral. Magnetite is scarce in the massive ore.

In all ore, the nickel content is quite constant, although nickel is slightly higher in disseminated ore in peridotite. The copper content is appreciably more varied; in massive ore the ratio is 10 nickel to 1 copper, in the more continuous sulfide phase of the high-sulfide disseminated ore, the ratio is 5 nickel to 1 copper, and in the scattered-bleb disseminated ore, 2 to 3 nickel to 1 copper. Nickel content of pentlandite has a considerable range, with the flame-like inclusions in pyrrhotite being lower in nickel than the coarse grains. The pentlandite in the troilite-bearing pyrrhotite is higher in nickel than the types just mentioned (Haapla, 1969).

At Hitura, the sulfide mineralization is strongest in the outer part of the sertintinite plug; less abundant and more weakly disseminated sulfides are found further within the serpentinite. Massive ore is present mainly as a small lens at the western contact zone of the ultramafic-mafic mass. The rocks surrounding the plug contain some nickel- and copper-poor disseminated iron sulfides. The ore types are (Papunen, 1970): (1) disseminations of sulfides in peridotite, (2) disseminations of sulfides in serpentinite, (3) massive sulfides in the serpentinite, (4) massive sulfides in the contact zone, (5) disseminated sulfides in granite veins and gneiss fragments, (6) disseminated sulfides in wall rocks. Type (1) is a high-grade ore and is located in the transition zone between serpentinite and peridotite. The sulfides constitute a continuous intergranular network between pyroxenes and relics of olivine. In amphibole-rich portions of the peridotite, the sulfide content is not as regular since the sulfides fill the interstices in acicular amphibole prisms. The main sulfides are pyrrhotite, pentlandite, and minor chalcopyrite; mackinawite and valleriite are most common in serpentine-rich rocks. Fractures in idiomorphic pentlandite crystals are quite usually filled with magnetite or serpentine. (These facts suggest that the probably secondary mackinawite, valleriite, magnetite, and serpentine are produced by the alteration of olivine-rich peridotite). Further, the alteration of pentlandite to mackinawite is most abundant in highly disseminated serpentine-rich rocks, where sulfides (mainly pyrrhotite) also may have been replaced by chlorite, actinolite, magnetite, and dolomite. In serpentine-poor amphibolitized peridotites, the pyrrhotite and other sulfides are quite unaffected by alteration. Some of the pyrrhotite is troilite-bearing.

In contrast to disseminated ore in peridotite, this type (2) of ore in serpentinite shows (Papunen, 1970) replacement textures on a large scale, mackinawite and valleriite disseminations, and only minor amounts of sulfides; such primary sulfide textures as can be seen show the sulfide grains to be interstitial. Much of the primary sulfide has been replaced by chlorite flakes with interlayers of talc, magnetite, dolomite, and mackinawite. Pentlandite crystal forms can still be recognized, but the nickel sulfide has been entirely changed to mackinawite with magnetite filling fractures in sulfide grains; magnetite may coat sulfides. Pyrrhotite and rare chalcopyrite grains may be found as residual inclusions in carbonate; cubanite occurs as fine prisms around sulfide grains. Graphite coats slickensides and with magnetite and carbonate replaces sulfides. (All these relationships seem to be compatible with the disseminated sulfides having been introduced in the ultramafic magma from which the serpentinite was later formed and having been extensively altered and replaced during the serpentinization process).

Massive sulfide bodies (type 3) are in pure serpentinite inside the zone of abundant disseminated sulfides in the eastern part of the sulfide mass; the sulfides are altered by the abundant production of chlorite flakes, mackinawite, valleriite, and dolomite, with a border zone of dolomite and chlorite from which thin chlorite sheets intrude coarse-grained pyrrhotite; mackinawite is largely, if not entirely, confined to pentlandite and valleriite to chalcopyrite. Locally, some sulfides in sheared rock volumes have been converted to serpentine and valleriite that are so closely intermixed as to be mechanically inseparable; some monoclinic pyrrhotite remains unaltered.

Massive ore in the western contact zone is about two meters broad; type (4) contains rounded fragments of wall-rock gneiss and peridotite that are strongly sheared and coated with slickensided chlorite. The principal constituent of this massive ore is much deformed pyrrhotite with abundant exsolution pentlandite; other pentlandite is in large, roundish grains that contain magnetite-filled fractures. Where pentlandite was formed near the border of massive ore with serpentinite, it has been altered to mackinawite, but, near gneiss contacts, the pentlandite is quite fresh. Where pentlandite has exsolved from pyrrhotite; it also is not much altered. Chalcopyrite is most common near the wall rock, and veins of it fill fractures in the adjacent gneiss; a silver-bearing pentlandite has been found in this chalcopyrite. This chalcopyrite contains abundant exsolution sphalerite stars. Valleriite is present only where ore fragments have been caught up in the serpentinite.

Some gneiss fragments contain disseminated sulfides as do some granite veins in serpentinite; chalcopyrite is the predominant sulfide here. All such sulfides are unaltered (type 5 ore).

The vein ore type (6) occurs in wall-rock gneisses near the western border of the ultramafic plug in which sulfides are disseminated; pyrrhotite is most important near the ultramafic-gneiss contact with some pentlandite accompanying it; pyrite and chalcopyrite, and some sphalerite, have been found beyond the pyrrhotite. This sulfide assemblage differs completely from that in the other five ore types, and this suggests that these sulfides may have come from another (and earlier?) source.

In a recent paper, Häkli, and others (1976) say that the quantity of platinum minerals in the Hitura ore body is small, and they are irregularly distributed. The minerals of this group are anomalously high only in the tectonically sheared marginal areas of the deposit that contain abundant chlorite-rich rocks. At Hitura, the tenor of platinum in essentially pure serpentinite is low, but much higher quantities are present in fresh olivine-bearing peridotites, suggesting to Häkli and his colleagues that some platinum is incorporated in the olivine and that what platinum was present in serpentinized olivine-rich rocks was moved elsewhere during the serpentinization. This "elsewhere" may be the chlorite-rich rock types that are high in platinum arsenides. Almost no platinum entered into the structure of the ore sulfides, the silicates, or the oxides, so it must have been incorporated in platinum group mineral (sperrylite) exclusively. Palladium behaves somewhat differently since it is enriched in the sulfide phase, particularly in the copper-rich residual melt, where it is found as tellurides and bismuthides. Palladium actually is more abundant than platinum, the latter averaging 0.027 grams/ton as opposed to 0.048 grams/ton of the former. Sperrylite is the only independent platinum mineral, whereas palladium is present in michnerite and Pd-bearing irarsite. The post-magmatic hydrothermal alteration of the Hitura deposit has controlled the deposition and mode of occurrence of the platinum-group minerals.

Although the Makola body is now mined out, Huhta's 1954 paper gives some information about the mineralization there. The principal ore minerals are pyrrhotite, pentlandite, and chalcopyrite, plus some cubanite, magnetite, pyrite, bravoite, and sphalerite; in places, cubanite is the only copper mineral, particularly in the lower part of the ore body. The most important non-metallic mineral is serpentine; talcose tremolitic amphibole is common along the margins of the deposits. Numerous fractures and shears contain chlorite, biotite, and calcite. The ore is cut by coarse-grained silicic dikes in which plagioclase (Ab_{73}) is the main mineral. Where the dikes are bordered by serpentinite, the contact zone is marked by chlorite and talc-bearing anthophyllite.

Huhta divides the Makola ore, on the basis of non-metallic minerals, into four types: (1) A-ore, which is a fine-grained, pure serpentine with sulfides, the grains of which are less than two millimeters in diameter; (2) B-ore, which is a more coarse-grained pure serpentine with sulfides, the grains of which are from four to 200 millimeters in diameter; (3) C-ore, of which the main rock mass is either serpentinite or amphibolite in which the sulfides are in widely scattered grains; the grade was, in many places, below the economic mining limit; and (4) D-ore, in which major mineral is amphibole (the rock is amphibolite) and the orientation of amphibole crystals can be seen; the sulfides are irregularly distributed and are in cracks in the amphibole grains or in compact, rounded, or ragged patches - this ore is higher in copper content than are the other three types.

Huhta thinks the sulfides were introduced after the ultramafic magma had solidified, molten sulfides having been separated from the ultramafic material before it reached its present position. This seems unreasonable, on the basis of more detailed studies of the largely similar Hitura and Kotalahti ore bodies. Rather, it appears that the ultramafic magma contained the sulfides in molten solution when it was introduced. Although most of this sulfide material probably separated from the dominant silicate melt in the molten state before all the silicates were crystallized; all, or almost all, of it appears to have been trapped among silicate crystals before the molten sulfides could settle into a massive deposit. The lack of massive sulfides in this ore body suggests that the category magmatic-2b does not apply to Makola except for minor amounts of apparently molten sulfides that penetrated into the surrounding gneiss. The sulfides in the amphibolite ore also may have been forced into that solid rock after it (the amphibolite) had been fractured, although the presence of patches of sulfides among the silicate minerals indicates that the sulfide more probably separated from the original magma from which the amphibolite (and its enclosed sulfides) were formed. The sulfides in cracks in the amphiboles may still have been molten when the rock was fractured; this is strongly suggested by the higher copper content of this ore type (4). The fact that the ore is almost entirely concentrated in the altered ultramafic rock would seem to mean that the sulfides separated from the peridotite magma after it had reached its present location.

None of the evidence that has been presented seems incompatible with the concept that the ores in all three deposits derived from the ultramafic magma in which they were introduced in molten solution. Considerable amounts of massive sulfide material appears, at least at Kotalahti and Hitura, to have settled to the base of the magma mass while still in the molten state and that some of this pool of molten sulfides was later forced into fractures and brecciated masses of the ultramafic rocks and, to a much lesser extent, into fractured and broken portions of the country rocks. Papunen's suggestion (1970) that much of the sulfur was derived from sulfide-bearing black schists in the area seems to lack merit because of the large number of similar ore deposits in the world where such a source for sulfur does not seem necessary. The deposits should be classified as Magmatic, 2a and 2b, plus Metamorphic-C.

KEMI

Middle Precambrian Chromite Magmatic-1b

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Notes

In June of 1959, a local resident of the town of Kemi (65°46'N, 24°34'E) sent some pyrite-bearing samples to the Geological Survey of Finland. On an inspection trip by geologists of the Survey, one non-magnetic sample was discovered that, on further study, was found to be chromite. This sample came from narrow chromite lenses 20 to 50 cm thick that lay at the bottom of a water channel dug for a local paper mill. On the basis of the chromite lenses and of the general geology of the area as reported by Mikkola (1949) and Härme (1949), the Survey decided to carry out gravimetric, magnetic, and electromagnetic investigations of the drift-covered area around the discovery site; the work was begun in the Fall of 1959. From the measurements made, a distinct positive anomaly (3 to 1.2 mgal) zone that was several kilometers long as found on the southeastern side of the discovery lenses; with this positive anomaly was associated a negative one (1000 to 2000 gammas). After making density and susceptibility studies of the chromite lenses and of the slightly magnetic serpentinite rocks surrounding them, the opinion was reached that the probability was high that the anomalies were caused by a chromite bed within the serpentinite. The first drill hole was put down at the western end of the anomalous zone and went through a chromite ore body some 20 m thick; further verification was provided by 25 additional drill holes, all of which encountered the chromite seam. Drilling continued until the end of April in 1960 and found that the seam ranged between 20 and 100 m in thickness and was estimated to contain 17.5 million tons of over 20 per cent Cr_2O_3 and 4.4 million tons over 30 per cent in that oxide. The Cr:Fe ratio, however, was found to be rather low, actually ranging between 0.84 to 1.87. In April

1960, the Survey conveyed the deposit, and the research data collected about it, to the Outokumpu Oy.

Eight separate ore bodies were found by diamond drilling in the 5 km-long favorable anomaly zone. The largest ore body was located in part on the western shore of, and in part underneath, Elijärvi (lake). By 1964, the reserves had been determined to be 33 million tons of 26 per cent Cr_2O_3 ore with a Cr:Fe ratio of 1.6 to 1.

All eight ore bodies outcrop; development work on these was begun in 1965 and, in 1967, a test concentrator was finished. After the test period, and annual production of 350,000 tons was scheduled; by 1974, the concentrator throughput was 410,000 tons of ore that averaged 25.8 per cent Cr_2O_3. From this ore, 130,000 tons of chromite concentrate and 28,000 tons of foundry sand were produced. During this same period, a stainless-steel plant was built about 10 km southeast of Tornio and in the immediate vicinity of the town of Kemi; this plant went into production at the end of 1976 with an annual capacity of 50,000 tons of stainless-steel plates and strip. The mine is 40 km by road from the plant. The nickel also needed in the production of stainless steel comes from the mines of the Outokumpu Oy described elsewhere in this volume (e.g., those of the Kotalahti and Ahlainen-Sääksmäki belts). Chromite concentrate above the 140,000 tons needed each year at Tornio is marketed in Finland and abroad. The annual production of foundry sand, now 15,000 tons, is used in various steel foundaries.

If a narrow triangle formed by a line connecting Tornio and Ylitornio as its base and the iron district of Misi at its apex, the distance from the base to its apex (in a northeast direction) is about 150 km; the geological unit enclosed by the figure is known as the Peräpohaja schist. This area is bordered on the north by the granite area of central Lapland and on the south by Pudajärvi granite gneiss. This so-called Peräpohaja schist area is said by Kujanpää (1980) to be a bowl-shaped structural unit made up of two depressions that strike about east-west and a ridge that separates them. These depressions are known as the Marttimojoki syncline (north) and the Kemijoki syncline (south); the ridge separating them is called the Korpikylä anticline (Mikkola, 1949). The Peräpohja area is made up of supracrustal rocks, associated with which are abundant igneous and volcanic rocks of much the same age; all of these rocks have been subjected to low-grade metamorphism.

In the southern part of this area, it is possible to establish, with considerable certainty, the stratigraphy of the sedimentary rocks; portions of this sequence can be identified in other parts of the area. The geological maps of the area (Perttunen, 1971, 1972, 1975) show (as the oldest rocks of the Supracrustal series) a basal conglomerate that lies on a basement gneiss complex, and it has been found at several points in the schist area. Overlying the conglomerate are rocks produced in the first stage of volcanism, for example, the agglomerates known west of the Kemi airport and at Sompujarvi. Over these volcanic rocks is a quartzite (the Kivalo orthoquartzite of Jatulian age) that averages at least 2 km in thickness and extends northeast from Kemi for about 100 km. Above the Kivalo is about 1 to 5 km of thick Karelian spilitic greenstones; these rocks belong to the second stage of volcanic activity. The rocks are typical amygdaloidal mafic lavas that contain local beds of agglomerates and metadiabase; no silicic differentiates are included in this formation. Above these lavas is a second orthoquartzite that may be as much as 0.5 km thick and contains several intercalations of dolomite. This second quartzite is overlain by the uppermost (third) series of volcanic rocks; these were originally tuffites and now are metamorphosed into mafic schists that still exhibit graded bedding and contain the largest dolomite bodies in the area as the uppermost horizon of the schist zone. The combined schist-dolomite formation has a maximum thickness of 1 km. In the central part of the Peräpohja area, black schists are included in the green schists; the mafic schists merge upward into graded-bedded phyllitic schists. An interformational conglomerate (the Taivalkoski) is contained in the phyllitic schists.

Between the basement gneiss and the basal volcanics and the Kival quartzite is a continuous albite diabase sill (0.5 to 1 km thick) and locally stratified sill-like ultramafic rocks (known as the Kemi and Penikat formations in the Kemi area and as Suhanko and Särkikämä formations in the Ranua area). The intrusions have displaced parts of the old basal conglomerate and the volcanics that overlie it from their original position to one between the albite diabase and the mafic plutonic rocks.

The Kemi sill, which intruded tetween the basement and the lowest supracrustal rocks, is a typical example (Kujanpää, 1980) of a stratified sill-like mafic-ultramafic body that differentiated in situ. Beginning at the town of Kemi on the Gulf of Bothnia, the Kemi zone can be followed for 15 km to the northeast with a width that ranges between 300 to 1900 m. The shape of this formation is that of a bent plate; it trends northeast and dips at an average of 70°NW. The region in which the Kemi zone is located is a flat, glaciated area no more than 20 to 25 m above sea level; the area is almost entirely covered by bogs and shallow lakes, and outcrops are sparse.

The Kemi zone, at its western end in the town of Kemi, is some 500 m wide; followed to the airport, it narrows to about 250 m; at Elijärvi (lake) it has gradually widened to about 1900 m. Northeast of Elijärvi the zone narrows over a short distance and dies out completely north of Kirvesjärvi (about 10 km from Elijärvi). In the few outcrops that are to be seen, no contacts of the zone are exposed, but one was made visible in 1977 in the footwall of the Viia open pit on the east side of Elijärvi. About 8 km northeast of Elijärvi, what may be a continuation of the Kemi zone appears. This zone continues northeast for about 21 km and with a width of 2 to 3 km; it is known as the Penikat intrusive, but it contains no economically workable chromite bodies.

The upper part of the differentiation sequence of the Kemi zone is made up of amphibolitized and saussuritized pyroxene gabbro arranged in what were anorthositic and pyroxenitic layers. In the middle of the formation, these layers (immediately above the ultramafic lower part) are unaltered and are chiefly noritic in composition. The total thickness of the altered and unaltered gabbro is as much as 1 km. The lower part of the Kemi zone consists solely of ultramafic rocks; the layers are mainly peridotite, pyroxenite, dunite, and chromite ore. The connected chromite-rich layer is situated from 50 to 200 m from the base of the Kemi zone.

Alteration, resulting from autometamorphism, has converted the peridotites to serpentinites and talc-carbonate rocks in the vicinity of the ore; nevertheless, the primary features of the peridotite can still be distinguished. The development of magmatic layering is best to be seen just above the ore horizon and is from 20 to 50 km thick; this layering is caused by the alternation, locally rhythmic, of the various layer-types, peridotite, pyroxenite, serpentinite, talc-carbonate rock, and chromite layers. At the base of the Kemi zone, the talc-carbonate rock has been intensely deformed immediately above the contact with the basement-gneiss complex; in this position the talc-carbonate material has been converted to a mylonitic talc-chlorite schist that is from 5 to 50 m thick. The total thickness of the ultramafic unit is about that of the gabbro unit, that is, up to 1000 m.

The economically valuable chromite layer ranges between a few centimeters and several meters in thickness. The intrusion is thickest in the Nuottijärvi-Elijärvi area and for 4.5 km in a northeasterly direction from there, the chromite zone incorporates six successive enlarged volumes in which the bulk of the ores are contained. In addition to these six ore bodies, two other bulges are known in a horizon 100 m above the main one, making a total of eight ore bodies. These are from southwest to northeast in the main ore horizon: Matilainen, Surmaoja, Elijärvi, Viianranta, Viianmaa, and Perukka - the two ore bodies in the upper horizon are Nuottijärvi (directly northwest of Surmaoja) and Viianlahti (northwest of Viianranta). The largest of these chromite masses is Elijärvi, and it was the major open pit in production; this pit was 670 m long and 95 m wide at the maximum. The hanging wall (northwest) side has a layered structure that extends downward into the upper segment of the ore body. The lower part of this pit and its western end have been highly brecciated. As a result of the breaking, the ore contains many inclusions of wall rock in a wide range of sizes; this makes for problems that require selective mining of the ore. On the contrary, the footwall of the ore body is mainly a talc-carbonate rock in which numerous bodies of chromitite are embedded.

The ore is divided into two main varieties: (1) a soft type in which the matrix is tremolite or talc-carbonate and (2) a hard type in which the matrix is much stronger serpentinite. The chromite is present in idiomorphic crystals that average 0.2 mm in diameter; it is the only ore mineral that is worth being, or can be, recovered. Other oxides present in the ore include: magnetite, ilmenite, hematite, and rutile. Sulfides occur as accessory minerals, mainly being located in fractures with pyrite, chalcopyrite, and millerite being present in the greatest quantity.

The soft ore (with the tremolite matrix) makes up a persistent and generally homogeneous layer along the hanging-wall contact of the ore body; this ore has the highest grade. This ore contains pseudomorphs of large pyroxene grains (about 5 cm in diameter) that is evenly filled with poikilitic chromite grains. Relics of the primary pyroxene are met with only in a few places. In addition to the tremolite as an alteration product of the pyroxene, chlorite (kämmererite), and talc also were so developed. Very fine-grained uvarovite is present in the upper part of the ore body.

Below the tremolite ore, the ore body largely has a talc-carbonate matrix. This ore type is produced from chromitite with an olivine-pyroxene or talc matrix, and it is the type in which autometasomatism of the silicates has been carried farthest toward completion. Whether the ore contains or does not contain serpentine, in addition to chromite, talc, and carbonate, rests on the degree to which metasomatism has progressed; the less the alteration, the greater the content of serpentine. This talc-carbonate ore is very soft, the more the carbonate content, the softer the ore and the greater the degree to which it breaks down. One of the products of the alteration process is the development of magnetite as a fine dust and as microscopically sized stringers in the carbonate. The ore with a dominant serpentine matrix has a structure much the same as that with one of talc and carbonate; in the matrix with abundant serpentine, however, the primary silicates have not been as much affected by alteration as in the talc-carbonate ore. The ore with the serpentine matrix includes, of course, chromite and serpentine as well as antigorite, some talc, and locally carbonate; this ore type also includes unaltered olivine relics.

Although the serpentine ore is very hard and resistant to surface processes, an increasing amount of talc makes it less so. Because the chromite grains normally are enclosed by a very thin talc coating, the ore breaks readily along the margins of the ore grains, making their concentration easily accomplished.

Kujanpää (1980) considers the Kemi chromite-bearing mafic-ultramafic sill to be typical of stratification by gravity and chemical differentiation.

The major ore body, that at Elijärvi, has been drilled on 100 m-centers with the deepest intersection of ore being at 450 m. In the Viianranta and Viianlahti ore bodies the deepest drill hole intersections have been at 250 m (twice). No drill data are availabe for the other ore bodies below the bottoms of the open pits. From the Elijärvi ore body 3.5 million tons were mined between 1966 and 1977. The surface plan of the Elijärvi pit shows an ore body with a most irregular outline surrounded in major part by talc-carbonate rock and in lesser part by serpentinite. In section, the ore consists of one steeply dipping major mass with local bulges or isolated satellite bodies of appreciably smaller size. Tiny ore lenses are found in the wall rocks, even in the peridotite and talc-carbonate schist. A typical feature of the Elijärvi ore zone is a layering in the hanging wall. The lower part of the ore body, to the northwest and the entirety of the western end, is brecciated with the ore containing much serpentinite and fragments of the country rock; obviously, this structure acts to require selective and expensive mining. Talc-carbonate rock makes up the footwall of the ore; it also contains small patches of serpentinite that include disseminated chromite grains and nodules and other chromite bodies of different sizes and shapes. The hard serpentine ore makes up about 15 per cent of the total reserves. The ore mined from the Elijärvi ore body alone averaged 26.5 per cent Cr_2O_3 (or 65 per cent chromite); the density of the ore was 3.45 and of pure chromite 4.65. FeO = 12 per cent ($FeO:Fe_2O_3$ = 3.5); Al_2O_3 = 9.5 per cent; MgO = 19.5 per cent; SiO_2 = 18.5 per cent; and Cr:Fe = 1.54. The chromite contained 0.15 to 0.35 per cent Ti; 0.15 to 0.20 per cent Mn; and less than 0.1 per cent V. No platinum group metals have been found.

At the end of 1976 and the beginning of 1977, mining was shifted to the Viianranta and the Viianlahti ore bodies through the Viia open pit which encompasses them both. In the next two years 1 million tons of ore was taken out, and then the ore reserves were about 5 million tons. In general, the surface plan of the Viia open pit is much the same as that of the Elijärvi, although the details, of course, differ widely. The Viia open-pit mining, however, is complicated by wide albite-diabase dikes that cut more or less northwestward across the pit. The geology of the pit differs from Elijärvi in that the Viia hanging-wall peridotites have been largely converted to talc-carbonate rocks in which the layered structures are lacking or are almost completely destroyed. Rock altered to serpentine is uncommon in both hanging and foot walls nor is there much of the serpentine-type ore in this pit. Talc and dolomite are essentially the only gangue minerals. The Viianranta ore bodies have been fractured in several distinct parts and exhibit a breccia structure much like that of the lower part of the Elijärvi ore body. Near the diabase dikes, the ore and the wall rocks are heavily broken. The chromite grains have been broken and rotated so that the grains of chromite are much smaller than they are in Elijärvi ores and are much more difficult to concentrate. This is true especially for the Viianlahti ore body whereas the Viianranta chromite is unbroken over considerable parts of its area.

The Viia open-pit ores contain about 1.0 per cent more chromium, and the Cr:Fe ratio is about 0.1 per cent higher or about 1.64; except for these differences, the differences in the main components are about the same as in Elijärvi.

The other five ore bodies have been examined only by diamond drilling, but they have much the same structures and occur in much the same types as those at Elijärvi. These grades of the five ore bodies range between 24.5 and 30.5 per cent Cr_2O_3, and the ranges of Cr:Fe are 1.52 to 1.59. The average of Cr_2O_3 for the entire ore zone is 27.5 per cent and the Cr:Fe ratio is 1.58. The Cr_2O_3 content of chromite averages about 34 per cent but may be as high as 51 to 52 per cent.

There appears to be no reason to doubt that the Kemi deposits were produced by magmatic processes. Isokangas (1978) believes that the magma from which the chromite deposits were derived was introduced between the Presvecokarelian basement and the Svecokarelian sedimentary and volcanic rocks. Complicating the study of the chromite ores and their genesis is the presence of an albite-granite sill that, in places, is in close spatial connection with the ultramafic lower portion of the Kemi intrusive. This sill probably was introduced after the ultramafic-mafic body, and its complex geometry in relation to the ultramafic rocks probably was caused by the tectonic activity that tilted the ultramafic-mafic body into its present 70°NW dip. The sill can be traced for more than 100 km northeast of Kemi; where the Kemi intrusion is lacking, the sill occurs either between the basement and the overlying Jatulian quartzite or within the quartzite.

Initially, the basal portions of the Kemi intrusive consisted mainly of pyroxenite and peridotite with probably are minor amounts of dunite in a layered sequence. About 50 to 200 m above the base of the ultramafic body, the main layer of chromite appears. This irregularity of the elevation of the chromite layer above the basement probably is due largely to elevations and depressions in the upper surface of the basement. The initiation of chromite precipitation probably occurred when the $Cr_2O_4^{-2}$ content of the magma became sufficiently high to do so (relative to the silicate content) and continued until that concentration had dropped low enough for deposition of chromite no longer to be possible. Locally, about 100 m above the main chromite layer another chromite layer has been found. It appears that this layer is normally too thin (or even is lacking) to be considered as ore, but two areas, Nuottijärvi (west of the Elijärvi ore body) and Viianlahti (north of the Viianranti ore body) are thick enough and rich enough to be mined (Viianlahti) or mineable (Nuottijärvi).

The major zone of wall-rock alteration both above and below the chromite ore bodies and between probably most, if not all, of the chromite layer(s) has been a major problem to explain. Piirainen (1974) and Isokangas (1978) agreeing with Piirainen believe that the ultramafic-mafic magma was high in water content because water was incorporated in the magma from the water-saturated clastic sediments immediately above the layered intrusion. They consider that the raised water content increased the fugacity of oxygen in the magma. Crystallization, therefore, began with the precipitation of ferromagnesian silicates and spinel. Apparently these two authors believe that spinel was an accessory mineral in the early crystallization of the ultramafic portion of the magma. With the concentration of $Cr_2O_4^{-2}$ steadily increasing relative to that of the silicates and of other anions acceptable in the spinel structure, deposition of actual chromite began. This hypothesis would suggest that the ultramafics under the ore should include accessory spinel that increased in amount of chromium content as the chromite ore zone was approached. Nothing that I can find in print indicates that this is so; instead, it appears that spinel deposition began suddenly and in large amounts in comparison with that of the various silicates. This precipitation appears to have stopped more gradually than it began, but the rise in concentration of silicate ions in the melt relative to those of chromite quickly changed the rock being formed back to the usual ultramafic sequence of pyroxenites and peridotites that is best developed in the sill immediately above the ore.

Another problem that needs explanation is the extreme degree of autometamorphism undergone by the ultramafic silicates both within and immediately above and below the ore seam. Water is necessary for these autometamorphic changes, and this is probably the reason for Piirainen and Isokangas to favor the incorporation was water from overlying water-rich sediments. On the contrary, it is possible that the original magma was high enough in water when it entered that, by the time that crystallization of anhydrous silicates had accumulated to a depth of 50 to 200 m, the water content had risen high enough that autometamorphism was possible on the scale observed in the ore zone and its immediate environment. Further work seems to be indicated on this point.

Veltheim (1962) says that he can find no direct connection between the ore zone and the country rock that surrounds it. He does not think that the enrichment in chromite in the ore zone was derived by the in situ crystallization of the overlying ultramafic-mafic magma. Instead, he suggests that, before the complex was intruded, an immiscibility developed between the silicate and chromite-rich phases of the parent magma took place. The main portion of the magma, the silicate-rich portion, had the composition essentially of olivine gabbro; this portion was intruded as a sill along the contact between the gneiss and the overlying Svecokarelian sediments. This chromite-poor portion of the magma then crystallized to form the differentiated ultramafic-mafic rock sequence that is the Kemi sill, minus the chromite seam(s). He considers that, during the accumulation of the ultramafic portion of the magma of the Kemi sill, partial alteration of the olivine took place plus some production of tremolite from the monoclinic pyroxenes. Still later, the ultramafic complex was intruded by the chromite-rich fraction of the original magma along the zone of autometamorphism in the solidified ultramafics. Veltheim is not certain whether or not this intrusion took place before or after the tilting of the complex nor is he clear as to why the intrusion chose this particular plane of entry because this alteration zone was not a geological boundary in any sense. It was, however, a zone of weakness in the ultramafic rock sequence, and this may have been provided enough ease of entry to the chromite-rich melt as to encourage its entry along this zone. He contends that, after the intrusion of the chromite occurred, the rocks of the ore zone and its immediately neighboring ultramafics underwent strong hydrothermal alteration to produce the ore-zone alteration as it now exists. Whenever these hydrothermal solutions entered the ultramafic-chromite rock zone, the chromite was the only mineral not affected by these solutions. The shear zones at the contacts of the country rock and the ore zone suggest to Veltheim that definite earth movements took place after the intrusion of the ore zone material.

Although the concept of Pirrainen and Isokangas still leaves some points unanswered (as mentioned above), Veltheim's idea seems much less logical. First, the possibility of molten chromite separating from a chromite-silicate melt is unknown in association with ultramafic rocks. Veins filled with chromite are present in the anorthosite of the Bushveld complex, and these definitely appear to have been introduced in the molten state. The chromite veins, however, are in anorthosite, stratigraphically far above the ultramafic rocks at the base of the Bushveld.

Secondly, the shear zones on the margins of the ore zone at Kemi seem much more probably to have developed after the autometamorphism and were localized by the minerals thus formed that were readily susceptible to shearing during the tectonic activity that tilted the Kemi sill. Shearing would have taken place during that movement, along the margins of the alteration zones around the ore zone simply because that zone was a zone of weakness in which stresses well could have been relieved during the tilting.

Thus, it seems far more reasonable to adopt the concept of one-magma that differentiated to form the present rock-ore sequence rather than that of two intrusions, the second of which had to confine its entry into the (incompletely developed) zone of autometamorphism.

If the one-magma concept is accepted, then the ores were formed under magmatic conditions by early separation, early crystallization (relative to the whole thickness of the magma) and should be classified, under the modified Lindgren classification as Magmatic-1b.

Because the Kemi sill was not intruded until after a considerable thickness of almost certainly lithified Jatulian and Karelian rocks; this means that the age of the ore formation must have been appreciably younger that that of the Presvecokarelian basement and the age of 2440 m.y. proposed by Kouvo. The Kemi ores, however, probably are of essentially the same age as Mustavaara (2100 to 2200 m.y.) and about that of the ores of the Outokumpu type. Thus, the Kemi ores are middle Precambrian but an appreciably older middle Precambrian than those of the circular nickel belt of south-central Finland.

MUSTAVAARA

Middle Precambrian Vanadium in Magnetite Magmatic-3b

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Notes

The Porttivaara mafic intrusion, of which the Mustavaara gabbro and its included vanadiferous magnetite body are parts, is located in the narrow waist of central Finland. The town of Oulu (and home of the University of that name) lies about 180 km by road to the southwest. Most of the intrusive in question is in the commune of Taivalkoski. The intrusive has been eroded into several hills, the highest three of which crest at about 390 m ASL; these are Porttivaara, Raiskiovaara, and Mustavaara. The total area of the Porttivaara intrusion is about 600 sq km.

In 1959, aeromagnetic measurements carried out by the Otanmäki Oy discovered a continuous magnetite gabbro horizon in the Porttivaara intrusion. This work had been the result of samples of vanadium-bearing gabbro sent in by various citizens (prospectors, professional or amateur) in 1957 and 1958 probably from the Porttivaara or Haukivaara intrusions. In the aerial investigations, an aeromagnetic anomaly was found in these two intrusions, but the vanadium content of this zone in the intrusions was too low to warrant further work. Later, the Otanmäki Oy investigated the aeromagnetic anomaly more thoroughly and found higher vanadium contents than had been reported from earlier work. Only in the Mustavaara area, however, were these contents high enough to justify diamond drilling. In 1967 and 1968, drilling was done on 200 m profiles, later narrowed to 100 m over a distance of 1800 m; to these were added 8 holes at the western end of the intrusion. On the results of this drilling, it was decided, in 1973 to develop a mine, and production began in 1976. Production was planned to be 1,500,000 tons of ore a year from which it was expected to extract 3000 tons of V_2O_5. This is done by leaching the magnetite concentrate, and the leached material is stockpiled against the day when a process will be available to separate the titanium from the iron.

The dip of the various layers in the igneous masses, and of the tops of these layers as well, is toward the north. The mafic bodies are far from uniform in composition, ranging from ultramafic to anorthositic as they are followed upward. These layered rocks have been introduced between the

Presvecokarelian basement and the Svecokarelian volcanic and sedimentary rocks above them. The contacts of the host rocks of these intrusions, both lower and upper, with the Porttivaara have undergone considerable deformation. Actually, between most of the uppermost basement rocks and the mafic layered intrusion, a quartz-albite-rich rock was introduced. This silicic body cuts the lower margin of the intrusion, and locally a hybrid between the quartz-albite body and the intrusion has been developed. The students of the area (members of the Koillismaa Research Project) believe that this quartz-albite rock was produced by a partial mobilization of basement material, this mobilization having been caused by the heat of the mafic intrusive body.

Immediately above the Porttivaara intrusive, the Svecokarelian volcanics are tholeiitic and keratophyric lavas; the sediments above them are quartzites, mica schists, and dolomites.

The Porttivaara intrusion is only one of several similar bodies in the area of the research project; both their position stratigraphically and their mineral composition is much the same as that of the Porttivaara. The largest of these are those designated as Syöte and Kuusijärvi-Lipeävaara with the Näränkävaara also being of major size. It appears possible to the research group that these bodies, at an early time, may all have been part of a single continuous sheet of mafic rock. Later Svecokarelian deformation and still later erosion appear to have caused the present isolated character of those bodies that make up what remains of the original sheet. What relationship, if any, the mafic-ultramafic intrusion northeast of the town of Kemi may have with those of the Porttivaara group is uncertain. The chromite ore of the Kemi area is, of course, quite different from the vanadium-rich magnetite ore of Porttivaara, but the stratigraphic positions of the mafic-ultramafic igneous rocks of the two areas is much the same. The two areas are over 100 km apart, so a correlation, if any, between them must be quite uncertain at this time.

Juopperi (1977) believes that the mafic-ultramafic intrusives were part of the early Svecokarelian magmatism with these rocks being younger than the mafic volcanics under which they now lie. These volcanics were emplaced subaerially.

Early on, it was recognized that the rocks of the area were too low in magnetite and the magnetite too high in inseparable ilmenite (at least inseparable with currently known processes) to be economic to mine and process. Not long after prospecting began in the area, prompted by samples sent in by members of the general public, it was apparent that, in a single and coherent magnetic gabbro layer near the upper part of the Mustavaara, sufficient vanadium was present in the magnetite to justify mining feasibility studies.

The Porttivaara intrusion is an igneous body in which rythmic and cryptic layering are prominent. [Cryptic layering consists of the regular vertical change in composition of certain minerals; cryptic layering is defined by the ratios of Fe to Mg in mafic minerals and Ca to Na in feldspars.] The Porttivaara intrusion is divided into several horizons, as shown in the following table.

The marginal zone is present around the entire layered series, but the outer contacts of this zone have been destroyed entirely by the strong deformation that the layered series has undergone. As the series now exists, the layers dip 30° to 40° to the north-northwest, but the dip at the bottom of the series is discordant by 5° to 15° to the dip of the layering.

The major primary minerals in the intrusive are plagioclase, augite, orthopyroxene, olivine, and ilmenomagnetite (in Buddington's sense of the term). Most of the pyroxenes have been uralitized, and the plagioclase has been, to a varied extent, changed to epidote. The greatest alteration has been in the marginal zones, in the magnetite gabbro, and in the layers directly above it.

Horizon	Feldspar An%	Olivine Fo%	Augite Mg	: Fe	: Ca	Approximate Thickness
Upper Marginal Zone	--					30 m
Anothositic Gabbro II	50-60		34	22	44	500 m
Magnetite Gabbro	50		37	21	42	150 m
Anothositic Gabbro I	60		40	18	42	185 m
Pyroxenitic Gabbro IV	63		41	16	43	450 m
Olivine Gabbro II	63	70	42	15	43	15 m
Pyroxenitic Gabbro III	67		43	13	44	600 m
Pyroxenitic Gabbro II	69					140 m
Pyroxenitic Gabbro I	70		45	12	43	40 m
Olivine Gabbro I	70	75				65 m
Lower Marginal Zone	--					40 m

Normally, plagioclase is the most abundant mineral, both in the layered rocks and in the zones around their margins; generally it has formed as a cumulus mineral. The crystals are in the form of laths; later growth after they had been collected at the base of the then molten magma gives them an irregular outline, generally sinuous. Usually, the plagioclases are zoned with the An content normally being 10 to 15 per cent higher in the centers; in rare instances, this difference may be as much as 20 to 30 per cent. In the lower portions of the intrusion, the zoning may be oscillatory. Plagioclase deposited from the intercumulus liquid is only weakly zoned, and such zoning is in the normal direction. As the intrusion is followed upward, the anorthite percentage drops from 70 to 50 (see the preceding table) although in the rocks above the magnetite gabbro (anorthositic gabbro II) the anorthite content locally is higher being due (Juopperi, 1977) to the cumulus manner in which the plagioclases were precipitated.

Olivine is confined to the two narrow olivine gabbros with the forsterite molecule being present as 75 per cent of the olivine in the lower rock of this type and 70 per cent in the upper one. The olivine grains are not zoned cumulus material, although they normally have been added to during adcumulus growth.

Orthopyroxene is present as intercumulus material in all rocks from Pyroxene Gabbro IV on downward; it is bronzite. It is a cumulus mineral only when it is present in those gabbros directly associated with olivine gabbro II. The olivine includes narrow augite lamellae in the (100) direction.

As is true of the plagioclase, the clinopyroxene, augite, is present in the entire vertical extent of the intrusion. It is in poikilitic grains and was the last silicate mineral to crystallize. It is not visibly zoned and is allotriomorphic in form, suggesting that it was produced intercumulusly. It contains narrow exsolution lamellae, too narrow for their composition to be determined. As the intrusion is followed upward, the augite is consistently enriched in iron in relation to magnesium.

Magnetite is an accessory mineral except in the magnetite-gabbro horizon. It is allotriomorphic to the silicate grains which it separates. It is so strongly altered that original changes in composition across the intrusion cannot be determined. The principal accessory minerals are ilmenite, apatite, and quartz; the quartz is present in a granophyric intergrowth with albite. Such sulfides as are in the intrusion are principally in the marginal metagabbro but some occur as accessories in the body of the layered rock, ilmenite is located as accessories in layers below the magnetite gabbro, being both as lamaellae in magnetite and as separate grains enclosed by silicates. Ilmenite in the magnetite gabbro is present only as lamellae in, or in intimate contact with, magnetite.

A division of the marginal zones is made into an upper zone that lies conformably above the layered rocks and a lower zone that makes up the floor of the intrusion. The lower zone has been divided into metagabbro below and metaperidotite above. The metagabbro is from 10 to 50 m thick and is distinctly heterogenous, the gabbros that make it up consisting of minerals widely varied grain sizes and contents of essential minerals. In some places, masses of metaperidotite are enclosed in the gabbro. The primary plagioclase and augite has been considerably altered, the plagioclase to albite and the pyroxene to clinozoisite, carbonate, chlorite, and amphibole. The overlying metaperidotite is even more highly altered; the primary olivine phenocrysts are now tremolite, magnetite, and serpentine, and the augite and orthopyroxene have been converted to masses of chlorite, talc, and carbonate.

The upper zone is highly inhomogeneous and is strongly altered. The metaperidotite is 10 to 30 m thick.

The Mustavaara intrusive contains two layers of olivine gabbro (I and II); the lower of the two is immediately above the marginal zone and contains olivine, plagioclase, orthopyroxene, and augite in minor amounts. Rythmic layering, due to changes in the amounts of olivine is characteristic of this layer. Olivine gabbro II is about 800 m above the base of the layered intrusion and is only about 15 m thick as opposed to about 65 m for olivine gabbro I. Olivine and feldspar vary reciprocally with olivine being in largest amounts at the base of this rock type.

Three layers of pyroxene gabbro (I, II, and III) lie one above the next and are distinguished by the presence or absence of rythmic layering. Pyroxene gabbro IV is separated from type III by olivine gabbro II. In all pyroxene gabbros, the essential minerals are plagioclase, orthopyroxene, and augite with the accessory minerals being magnetite, apatite, quartz, and suflides. Beginning with pyroxene gabbro III, the plagioclase grains are separated by a symplectic (intimate) intergrowth of albite and quartz.

Rythmic layering exhibited by pyroxene gabbro I is caused by changes in ratios of plagioclase to the pyroxenes; light layers have 70 to 90 per cent plagioclase and dark layers 20 to 60 per cent of that mineral; in both instances, the remainder is half orthopyroxene and half augite with dark layers being 0.5 to 3 cm thick and light 1 to 10 cm. Pyroxene gabbro II is homogeneous with 60 per cent long laths of pyroxene and 35 per cent poikilitic augite; orthopyroxene and other accessories make up the remaining 5 per cent.

Rythmic layering appears in the upper and lower parts of pyroxene gabbro II, layering being due to changes in proportions of plagioclase and pyroxenes as well as to changes in plagioclase grain sizes. Plagioclase ranges between 30 and 65 per cent and pyroxenes between 70 and 35 per cent. The ortho- and clinopyroxenes are about equal in amount. Layer-thickness is between 5 and 15 cm. The pyroxene grains are poikilitic in relation to plagioclase.

Pyroxene gabbro IV shows rythmic layering with the upper parts of these layers being richer in plagioclase than the lower parts; a few layers are homogeneous. Layer thickness is from 1 to 10 m. Plagioclase runs from 50 to 60 per cent and orthopyroxene and augite each are 35 to 15 per cent of the rock although augite is the only pyroxene in the upper part of this gabbro.

The two anorthosite gabbros (I and II) are in the upper part of the layered series and are separated only by the magnetite gabbro. The anorthosite is light, coarse-grained, and highly altered, layering is weak in type I and rare in type II with what layering there is being just below the magnetite gabbro. Nearly all plagioclase is now epidote and the augite is completely converted to uralite (or to chlorite in type II); no evidence has been found of orthopyroxenes in the anorthosite. The amount of plagioclase in type I anorthosite may be as much as 95 per cent (a true anorthosite); this drops to 65 per cent in the dark layers of type I rock; plagioclase grain size is appreciably smaller in anorthositic gabbro than in anorthosite. The uppermost 10 to 50 m below the magnetite gabbro is mostly anorthosite.

In anorthosite II, the rock is much more homogeneous than type I. The plagioclase is 70 to 80 per cent of the rock and altered augite is from 15 to 25 per cent. Ilmenomagnetite is present interstitially in both anorthositic types, but it is greater in type II than in type I.

Because of the lack of a fine-grained chill zone, it is difficult to calculate the composition of the primary magma from which the various layers were crystallized; Juopperi (1977) reports on an estimate of this composition derived from a mean value obtained from 165 samples gathered in five profiles across the intrusion. This analysis (obtained from Mäkelä's 1975 masters' thesis) is as follows:

Analysis		Norm			
SiO_2	51.35	ILM	= 0.90	Mg/Fe	= 0.73
TiO_2	0.64	OR	= 3.89	Co	= 0.007
Al_2O_3	16.14	AB	= 25.28	Cu	= 0.030
Fe_2O_3	1.59	AN	= 30.04	Cr	= 0.030
FeO	7.13	MAG	= 1.68	Ni	= 0.033
MnO	0.14	FS	= 6.33	V	= 0.025
MgO	8.09	EN	= 15.26	Zn	= 0.010
CaO	9.48	DIOPS	= 14.08		
Na_2O	2.78	FO	= 1.79		
K_2O	0.65	FA	= 0.74		
H_2O+ H_2O-	1.50				
Total	99.49				

What has thus far been said about the Porttivaara intrusion does not indicate that it actually has any economic value for any of the elements or minerals that it contains. This will be considered in what follows.

The magnetite gabbro of the Porttivaara intrusion in the Mustavaara area was found to form a coherent sheet within the anorthosite gabbro in this part of the intrusion. This sheet, and the magnetite gabbro in which it is contained, is cut in half by a fault into more or less west-southwest striking portions, the eastern of which is about 13 km long and the western about 6. The western portion is cut off by another fault at its west end, whereas the eastern end of the east portion narrows and, east of Mustavaara, dies out completely. Except for the narrowing in the eastern section, the horizontal width of the magnetite layer is about 300 m; the dip ranges from 30° to 40° more or less N, but it does become steeper to the east where the last hole drilled found a dip of about 70°N.

The transition upward from the magnetite gabbro to anorthosite gabbro II takes place over about 50 m with the major number of anorthosite gabbro blocks in magnetite gabbro gradually decreasing until the rock becomes an inhomogeneous anorthosite gabbro including specks of magnetite gabbro. Finally, no more magnetite gabbro remains to leave the rock as standard anorthosite gabbro II. On the contrary, the lower contact of the magnetite gabbro with the underlying anorthosite gabbro I is sharp. This anorthosite gabbro is layered for the first 3 m below the contact, with the light layers being anorthosite and the dark anorthosite gabbro. In this 3 m, the ilmenomagnetite is about 5 per cent of the rock; for the next 13 m, the rock is spotted with dark minerals to some extent, normally the spots are few enough that the rock is true anorthosite.

The magnetite gabbro is so composed that it can be divided into four layers, each of which is sharply divided from the others. The lowest three of these layers are sufficiently rich in vanadium that they can be mined as ore; these are known as: (1) ore lower layer (OLL), (2) ore middle layer (OML), and (3) ore upper layer (OUL). The fourth layer contains so sparse a dissemination of ilmenomagnetite that it is not economically mineable and is known as disseminated rock (DR). The lowest layer is the richest and contains 25 to 35 wt per cent of magnetite; the middle layer (OML) is the most poorly mineralized with about 15 per cent magnetite.

The DR is quite inhomogeneous, containing scattered anorthosite and anorthosite-gabbro fragments and tiny dikes of compact magnetite. Anorthosite fragments also are present in the ore; they are randomly oriented in it and most of them are rounded with sharp contacts obtaining with the surrounding ilmenomagnetite. Ilmenomagnetite away from the contact with anorthosite normally is altered to silicates but with the ilmenite lamellae having been preserved.

Toward the east, each of the layers of the magnetite gabbro participates in the narrowing but the thinnest layer (OLL) is least obviously narrowed. The OML thins from about 50 m on the western side of Mustavaara to 15 m in its most easternmost known occurrence. The corresponding thicknesses for the (OUL) are 40 and 10 m. The (DR) does not continue as far east as do the ore layers; thus, in the east, the (OUL) is the uppermost layer of the magnetite gabbro. Some layering within the four layers of the magnetite gabbro can be discerned; the contacts between these minor layers are quite definite and are parallel to those of the four major layers.

Where the magnetite has been converted to silicates and relict ilmenite lamellae, the affected rock is no longer magnetically susceptible.

The primary minerals of the magnetite gabbro originally were plagioclase, augite, and magnetite; no orthpyroxene has been identified. Augite has been almost universally altered to uralite, and other alteration minerals include epidote, chlorite, biotite, titanite, and leucoxene. The accessories are quartz, apatite, and sulfides.

Juopperi (1977) points out that the magnetite in the magnetite gabbro is "ilmenomagnetite" in Buddington's sense, that is, magnetite with microscale intergrowths of ilmenite. This discussion follows Juopperi in that magnetite and ilmenomagnetite here both mean ilmenomagnetite, as does Ti-bearing magnetite. The Ti-bearing magnetite normally is present as anhedral grains that fill spaces among silicate minerals with these grains being elongated parallel to the layering of the intrusion. In the (OLL), where the Ti-bearing magnetite grains are most common, they touch each other to form curved chains of several grains. The grain sizes of the ilmenomagnetite grains are 0.3 to 1.0 mm. Idiomorphic grains are common, and are even more common than the anhedral ones in the easternmost part of the Mustavaara body. In this area, the (OML) and (OUL) contain only idiomorphic grains; the (OLL) contains allotriomorphic grains also.

The ilmenite lamellae take three different forms: (1) as conforming to the (111) direction in magnetite, some few of these are much broader than the majority -0.1 mm and the majority are 60 - 1 micron; the range of 60 microns is not found in any one specimen, the greatest difference being 20 microns; (2) as splinter-like and much thinner lamellae between the longer ones but following the same (111) direction; where these are abundant, they form non-directional patches; (3) as granular grains inside the magnetite or on its border; these grains usually are less than 0.5 m in diameter. The granular lamellae are very sparse in the (OLL), more abundant in the (OML) and (OUL) but most common in the (DR). Some lamellae in the ilmenomagnetite may be spinel in the strict sense. Although the magnetite usually is replaced, where altered, by silicates and the ilmenite preserved, the reverse may occur and the ilmenite be altered to titanite. In alteration related to shear zones, both ilmenite and magnetite may have been altered, the first to ilmenite and leucoxene and the second to chlorite; here the rock silicates also are altered, plagioclase to albite, quartz, and epidote, and uralite to chlorite.

Some ilmenite grains are not in direct contact with magnetite, and Juopperi thinks they definitely must be primary; they are not, however, found in magnetite gabbro. Nor is the quantity of ilmenite as inclusions in, or at the edges of, magnetite grains as small.

Exsolution rods (usually of magnetite) are present in granular ilmenite and in the broader lamellae of that mineral. This exsolved magnetite can be oxidized to hematite, and surrounding magnetite is oxidized to martite. Juopperi points out that these relationships show that small amounts of magnetite can be taken into solid solution in ilmenite at high temperatures.

Magnetite gabbro also contains rather uniformly distributed, but minor, chalcopyrite; generally it occurs as inclusions in uralite but also may circle the uralite grains: most of it is in the (OLL). Minor quantities of other sulfides have been identified.

Juopperi (1977) gives tables showing the chemical and modal composition of the Mustavaara magnetite gabbro:

Components	DR	OUL	OML	OLL
SiO_2	44.76	38.99	42.72	34.99
TiO_2	2.06	2.11	2.00	2.79
Al_2O_3	14.69	12.49	12.27	10.44
Fe_2O_3	8.73	16.49	11.63	19.81
FeO	9.56	12.85	11.08	15.61
MnO	0.21	0.17	0.17	0.21
MgO	5.90	4.28	5.18	3.17
CaO	10.88	8.61	9.21	7.31
Na_2O	2.53	2.47	2.48	2.32
K_2O	0. 3	0.34	0.35	0.89
V	0.15	0.26	0.22	0.38
Cr	0.002	0.002	0.003	0.013
Co	0.008	0.013	0.010	0.013
Ni	0.009	0.011	0.014	0.029
Cu	0.032	0.078	0.076	0.061
Zn	0.01	0.015	0.012	0.017
Totals	99.90	99.17	97.42	98.05

The mean values of the modal analyses quoted by Juopperi are given here:

Ore Layer	Plag	Aug	Ti-Mag	Total
DR	50.2	39.4	10.4	100.0
OUL	50.4	24.2	25.4	100.0
OML	55.3	30.8	13.9	100.0
OLL	49.5	17.6	32.9	100.0

One of the immediate problems that had to be solved before the vanadium-bearing material could be mined and processed at Mustavaara was to locate the vanadium in the minerals of the ore layers. It was soon discovered that the vanadium was in the magnetite proper to a much larger extent than in the ilmenite lamellae and grains. Juopperi points out, however, that "the small differences in mean values of vanadium in the different layers of magnetite gabbro do not vary systematically in the profiles." Analyses done by microprobe of the magnetite between ilmenite lamellae showed very small quantities of titanium in the magnetite, less than 1 per cent. This, of course, means, a very limited solid solution between magnetite and ilmenite as the temperatures of the original solid solution fell, so that no more than 1 to 2 per cent of ulvospinel [$Fe_2Ti^4O_4$] at present remains in the magnetite. Actually, the concentration of vanadium is three to five times greater in magnetite than in ilmenite, suggesting that the vanadium is present in solid solution in the magnetite [probably as the mineral coulsonite ($Fe_2V^4O_4$)], an analog of ulvospinel, but one that is much more soluble in magnetite (in proportion to the ratio of Ti/V) than is ulvospinel. The magnetite samples tested from Mustavaara ran quite high in vanadium (1.6 to 1.7 per cent V_2O_3). These quantities of vanadium were quite constant throughout the ore layers but became lower in the disseminated rock (DR). Presumably what vanadium is present in the ilmenite is isomorphous with the alpha hematite structure of the ilmenite. Juopperi's work also shows that the iron in the ilmenite lamellae (above that needed to produce ilmenite) is present as hematite and not as magnetite.

In discussing the crystallization of the Porttivaara intrusion (of which the Mustavaara section is a part), the metagabbro at the contacts of the intrusion (originally pyroxene gabbro) indicates the character of the original intrusion. The contact of this magma with the cold country rock into which it was intruded almost certainly provided a sample of what the entire magma was like because there was not time for any differentiation. This layer, however, is not homogeneous, and Juopperi accounts for this by suggesting that this layer may have been the result of several injections of magma, the intervals between which were long enough that the practically solidified earlier magma was brecciated by the newly arrived material. This metagabbro layer has not been observed along the upper contact of the intrusion, partly because good outcrops are scarse and partly because the upper layer must have been quite thin.

The peridotite layer immediately above the metagabbro must have been a cumulate in which the essential minerals were olivine, augite, and orthopyroxene. The primary texture, however, was so much obliterated by alteration that it cannot now be observed. The metaperidotite probably resulted from gravitative differentiation after the magma chamber had been filled and the magma had become stationary. This produced a differential density sorting with the heaviest minerals settling to the bottom of the chamber; the gently sloping walls of the chamber allowed even settling and explain the lack of slumping along the chamber sides.

During this initial crystallization, the olivine and pyroxene, being heavier than the magma, sank to the chamber floor, but the plagioclase, being lighter, rose toward the roof. Juopperi thinks that this upward accumulation of plagioclase may have continued throughout the crystallization of the magma (despite such currents as may have existed) since some quiet intervals must have occurred during which some of the plagioclase crystals had time to rise from their places of nucleation to the roof of the magma chamber. The widespread distribution of plagioclase from the anorthositic rocks at the top of the magma chamber to the olivine gabbro near its base indicate that the rise of plagioclase crystals toward the roof during periods of low current activity largely was balanced by downward movement when currents were strong in the magma.

In the Porttivaara intrusion, Juopperi thinks that the rhythmic layering in the main body of the intrusion was caused by magmatic currents. The primocrysts [a primocryst being a cumulus crystal after it finally had settled out of the molten magma] involved in the mechanical sorting that formed the olivine gabbros were olivine and plagioclase; whereas, in the pyroxene gabbros, the primocrysts were plagioclase and perhaps smaller amounts of pyroxene. The interstitial positions of the pyroxenes, however, suggest that much (or perhaps most) of the pyroxenes crystallized among the plagioclase laths. Because these pyroxenes, therefore, crystallized only on the floor of the magma chamber, they were not very readily removed from their parent melt. As a consequence, the pyroxenes achieved a better equilibrium and enrichment in iron was less than if they had been primocrysts and had been withdrawn from the surrounding melt by magmatic currents.

Further, cryptic layering was superimposed on the rhythmic layers of the Mustavaara intrusion. Thus, despite long contact of the rock minerals with the molten magma, gradual changes did occur in the composition of the minerals forming solid-solution series, so that progressive changes did take place in their composition from the bottom of one cryptic layer to its top. This gradual change in composition would be followed by an abrupt appearance or disappearance of a given cumulus mineral (or minerals) across the width of the remaining magma to initiate the formation of a new cryptic layer. These changes, Juopperi believes, were associated with fractional crystallization. The occasional development of olivine in the magma (to produce the two olivine gabbros) follows the principal that, in multicomponent systems, a magmatic liquid can cross the boundary curve separating one mineral (olivine) from another (pyroxene) granted the proper changes in the governing parameters of the system are made. Juopperi considers that, in the formation of the lowest cumulate (i.e., the parent peridotite of the metaperidotite) the magma was on the phase boundary between olivine and pyroxene. During the development of olivine gabbro, however, the magmatic liquid was on the boundary curve between olivine and plagioclase. During the progressive crystallization of the magma, steady changes in melt composition, produced by the fractional crystallization of minerals differing in composition from the magma itself, resulted in such shifts in total composition that the magma passed from one area of mineral stability to another, and a quite different set of minerals began to precipitate.

Very simply, the second development of olivine gabbro, some 800 m above the base of the layered intrusive, occurred because the crystallization of the some 700 m of pyroxene gabbro below OLGB II changed the composition of the magma sufficiently to bring it onto the stability surface on which all four essential minerals of the olivine gabbro [olivine, orthopyroxene (apparently bronzite), augite and plagioclase] all were stable. As crystallization of this olivine gabbro continued, olivine was proportionally reduced and plagioclase increased. By the time some 15 m of this olivine gabbro had been precipitated, however, the composition of the magma had been sufficiently changed that it had passed out of the portion of the surface on which olivine was stable and entered one in which only the three essential minerals of pyroxene gabbro IV (ortho- and clinopyroxene and plagioclase) were stable. As the deposition of pyroxene gabbro continued, the amount of that mineral that could be precipitated was reduced sufficiently that the rock being formed changed from pyroxene gabbro to anorthosite gabbro.

The major change in the orderly conversion of the rocks being formed from the Mustavaara magma from olivine gabbro to anorthositic gabbro was the insertion, about 1500 m above the floor of the intrusive, of the vanadium-bearing magnetite gabbro. Juopperi discusses the various methods by which such iron-rich rock may have been formed. These are: (1) early gravitational settling of magnetite (or ilmenomagnetite) as at Ülvo in Sweden - the position of the magnetite gabbro at Mustavaara rules out the possibility; (2) Vogt's concept of the remelting of early magnetite and its movement upward through the rock column - the lack of any evidence of early magnetite or its upward movement after remelting rule out this concept; (3) the development of a residual iron-rich melt as suggested by Fenner - the very slight change in iron content of the augites in the Mustavaara intrusion indicates that the possibility of the explanation applying here is poor; (4) a metamorphic origin for the magnetite (as suggested by Ramdohr for the magnetite associated with charnockites) is ruled out by the minor to neglible metamorphism that affected the Mustavaara rocks; (5) liquid immiscibility - such immiscibility is supported by experimental studies, particularly those of Fischer in 1950, showing that such immiscibility, depending on the original and derivative compositions of the magma involved, can occur at almost any stage in magma crystallization.

Juopperi believes that the last explanation (5) best fits the situation at Mustavaara; he considers that this hypothesis is compatible with the texture and composition of the associated augite. At the time of separation of the iron-rich melt, the solidus temperature of this liquid was lower than that of the coexisting silicate melt. Globules of this liquid were caught up in the various currents moving through the partly molten, partly crystallized silicate material and thus became concentrated to various degrees in the silicate matrix of the gabbro, producing a layered structure in the gabbro. In addition to the plagioclase crystals and the magnetite globules trapped among them, the magmatic currents moved anorthosite blocks from the roof area of the magma chamber; these blocks caught up in the solidifying gabbro give the appearance of having been brecciated. Juopperi points out that the blocks and the molten material that moved them had different velocities; the long axes of the plagioclase crystals, therefore, became oriented parallel to the surface of the anorthosite blocks.

Juopperi also thinks that this immiscibility hypothesis readily explains the increase in the amount of ilmenomagnetite in relation to augite. His explanation requires only that different proportions of oxide and silicate melts existed in different parts of the magnetite-gabbro parent material. Had there been only one interstitial liquid (containing both the components of magnetite and of augite),the greater the amount of ilmenomagnetite formed, the greater would be the magnesium content of the residual liquid and the higher the magnesium content of the augites. Such a rise in augite magnesium was not observed which argues cogently for the existence of two immiscible melts.

Juopperi establishes that Fischer's (1950) experiments show that an oxide-rich melt can be separated from a silicate one if: (1) the silicate melt contains alkalies in excess over calcium; (2) the heavy metals are in a high state of oxidation; (3) the magma must contain phosphorus, and (4) the magma must contain appreciable volatiles. Only the first condition is not fulfilled at Mustavaara. The presence of volatile material is indicated by the alteration of augite to uralite and plagioclase to epidote (volatiles here, of course, including water). The sparse presence of apatite grains suggests strongly the presence of phosphorus, and the ore itself shows that the metals are highly oxidized. Alkali-rich rocks, however, are not in association with the Mustavaara layered intrusion.

The Mustavaara vanadium-bearing magnetite gabbro should be classified as "in magmas by late separation-late solidification (with or without fluid injection) immiscible melts, metal-oxygen rich, metal-phosphorus rich or, in the usual shorthand, Magmatic-3b.

ORIJÄRVI-AIJALA-METSÄMONTTU

Middle Precambrian	Zinc, Copper, Lead, Gold, Lead	Hypothermal-2 and -1

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Notes

The deposits of Orijärvi, Aijala, and Metsämonttu all lie in the Orijärvi leptite zone made famous by the classic studies of Pentti Eskola (1914, 1915) on the classification of mineral facies and on iron-magnesium-silicate metasomatism. The Orijärvi deposit is about 30 kilometers north of the town of Tammisaari (60°30'N, 23°50'E) on the southern coast of southwestern Finland and the Aijala (east) and the Metsämonttu (west) mines are located about 10 kilometers west-southwest of the Orijärvi mine. The first copper discovery on the Orijärvi property was made in 1757. Systematic regional prospecting, however, was not begun in this (Kaski) area until after the end of World War II (late 1945). The first economic ore was not found at Aijala until 1947, when the East (Itämalmi) and West (Länsimalmi) bodies were located. The Aijala and Metsämonttu ore bodies are about 1.5 kilometers apart, with the original estimates showing one million tons of ore at Aijala and 0.6 million at Metsämonttu. These tonnages probably were considerably increased because, in the early 1950s, 100,000 tons annually were being mined at Aijala and 70,000 from Metsämonttu. The literature on the mine geology was scarce until the recent publication of Latvalahti (1979); that on the rocks of the area is extensive. The several types of iron ore in the area are not considered here.

From 1932 to 1942, about 30,000 tons of ore was mined annually from the Orijärvi mine; in that period the grade was 1.4 per cent lead, 1.0 per cent copper, 4.5 per cent zinc, 0.4 grams gold, and 10 grams silver per ton; after 1942, the grade gradually went down. In 1948, mining of old dumps (accumulated when only copper ore was mined) was begun, and total production was increased to 40,000 tons per year; this ore averaged 0.7 per cent copper, 0.8 per cent lead, and 2.5 per cent zinc but no amounts for gold or silver are mentioned. In 1954, the mine was reported to have passed its prime, and, in 1963, when I visited the area, all mining had ceased. At Aijala, the ore was said (Varma, 1954) to have the following mineral content: chalcopyrite 5.2 per cent, sphalerite 0.9 per cent, pyrite 16 to 18 per cent, pyrrhotite 2 to 3 per cent, galena 0.01 per cent; at Metsämonttu, these mineral percentages were, in the same order, 0.25 per cent, 6.9 per cent, 9 to 10 per cent, 18-20 per cent, 0.12 per cent. Silicates and carbonates amount to 75 per cent of the Aijala ore, higher than the 65 per cent at Metsämonttu. The grade of the two ore bodies was, at Aijala, 1.8 per cent copper, 0.6 per cent zinc, 21.6 per cent iron, 17.4 per cent sulfur, and only arsenic (0.16%) of all other metallic ore elements was over 0.1 per cent; at Metsämonttu, only 0.1 per cent copper, 4.6 per cent zinc, 18.6 per cent iron, 16.1 per cent sulfur, and 0.1. per cent lead. Aijala ceased production in 1958 and Metsämonttu in 1974.

Eskola (1963) states that the leptite zone of southwest Finland, in which the ores are contained, is a belt of supracrustal rocks of middle Precambrian (Svecofennian) age. According to Latvalahti (1979), the volcanic activity that produced the leptites occurred about 1900 million years ago with the orogeny that later affected them being no younger than 1800 million years ago. The leptite zone is some 110 kilometers long and extends from the southwest coast to the Lohja district to the northeast. Although most of the leptites in the area of the mines are of volcanic origin, some sedimentary leptites are present. These rocks are made up mostly of quartz and feldspar, and range in grain size from fine (hälleflinta) to medium (leptite), to coarse (leptitic gneiss). Some mafic or intermediate volcanics are included in the leptite sequence. The leptites comprise the lower Svecofennian group in which the original materials were lavas, pyroclastites, and epiclastites. Some limestone and iron-formation lenses are interlayered with the dominant leptites as are some arenites. The thickness of the group is less than 3000 meters. Above the lower Svecofennian group is the middle group, mainly mafic to intermediate rocks that originally were lavas, pyroclastics, and epiclastics and also include random limestone layers; the group has a thickness of between 500 and 1000 meters. The upper group is composed of arenaceous and argillaceous rocks - phyllites, mica schists, graywackes, graywacke schists, mica gneisses, and arkosites; the thickness ranges widely but is less than 3000 meters. The volcanic rocks are sodium-rich, the sedimentary leptites potassium-rich.

The leptite zone in the Orijärvi-Aijala area is bordered by granodiorite; with this rock are associated quartz diorite, diorites, gabbros, and hornblendites. Latvalahti (1979) says that the contacts of plutonic rocks with leptites usually are conformable. In addition to these synkinematic rocks, the north and south borders of the leptite zone later were intruded by post-kinematic microcline granites.

The synkinematic intrusives southeast of the Aijala mine divide the generally westerly trending leptite zone into two parts. The southern part of the zone extends roughly eastward and contains no known ore; it is not mentioned further here. The northern part of the zone trends northeast and includes both the Aijala and Metsämonttu ore deposits; about six kilometers to the east of Aijala, the leptite schist zone separates into two parts, forming a synclinal structure that widens to the northeast. The southern limb of this structure contains no ore. The Orijärvi sulfide ore body lies near the base of the northeastern limb.

The synkinematic intrusive that divides the leptite schist belt forms a diapirically uplifted anticlinal batholith and ranges in composition from granite to granodiorite with some hornblendite sections. Locally, the granite has bent outward the schists that surround it. Partly assimilated fragments occur in the more mafic parts of the intrusive. The postorogenic microcline granite does not outcrop in the vicinity of the various ore deposits.

In the areas of the ore bodies, leptite rocks are higher in proportion than in other parts of the leptite belt.

The leptite rocks of the ore-bearing area have been affected by two phases of folding. In the Aijala-Orijärvi area, the older folding is isoclinal; the fold axes plunge gently to the east. The younger phase also produced isoclinal folds in which the subvertical axes plunge to the southwest at high angles. In both phases, the axial-plane schistosity is subvertical.

Lenses of skarn, limestone, and iron formation interlayered with the leptite show small-scale folding in many places. On the contrary, the acid volcanics only rarely show folds or other structures tectonically caused. The lineation of the second-stage folding is essentially always to be seen, but that of the first seldom is visible. Aijala is a far higher-strain region than Orijärvi; this is indicated by the more intensely deformed ejecta in the former area than in the latter. Further, the axial plane foliation is much more strongly developed at Aijala than at Orijärvi.

Post-metamorphic fault zones occur in the Aijala-Orijärvi area with the most prominent being the Jyly fault zone east of Orijärvi and the Kirkkojärvi fault zone that runs east-northeast about 500 meters north-northwest of the contact between the silicic leptites (southeast) and the intermediate leptites (northwest). The Metsämonttu ore body is cut by a fault at the 135 level that dips gently south; the fault displaces the upper part of the ore body about 270 meters north. A similar fault cuts the ore body on the 540 level and displaces it 80 meters north. The Aijala ore is cut off sharply at the 200 level by a fault and shear zone that dips steeply north. The general area of the Orijärvi ore is cut by several faults and shear zones; north-trending faults east and west of the ore body are the most prominent of these.

Latvalahti (1979) reports that regional metamorphism took place before the intrusion of the microcline granite. In the folding process, the first phase in part came before and in part after the regional metamorphism. The first folding stage had reached its peak before the second phase began. The rocks of the Aijala-Orijärvi area were metamorphosed to the low-pressure amphibolite facies; farther east metamorphism reached the granulite facies.

The diagnostic metamorphic-mineral assemblage in the ore area is muscovite-quartz-plagioclase (An_{20}) with the reaction locally producing sillimanite and andalusite. At Orijärvi, the mica gneiss actually is a cordierite-sericite mica gneiss, and it contains andalusite and sillimanite. Latvalahti calculates that the confining pressure was about three kb and the temperature 650° ± 30°C. Eskola (1915) maintained that the Orijärvi granodiorite mass caused contact metamorphism, the PT conditions of which were about the same as, or slightly lower than, those of the regional metamorphism which he contends followed the intrusion. This contact metamorphic aureole (?) surrounds the western two-thirds of the synkinematic silicic intrusive; this aureole is not duplicated, except to a minor extent, around the major mass of the granodiorite farther south from the Orijärvi area. Eskola considered that, northeast of the granodiorite mass is a belt, some 200 meters wide, of cordierite-anthophyllite rock. On its northeast, this rock is bordered by amphibolite. In turn the amphibolite is succeeded by leptite.

In vertical section, at right angles to the ore contact, the major ore lens follows the cordierite-anthophyllite amphibolite contact; further, the smaller ore lenses, as much as 50 meters from this contact, have the same attitude as this latter feature. In the two rock types just discussed, limestone and skarn lenses are intercalated. The skarns are composed mainly of tremolite and actinolite, and, in parts of some ore volumes, these two minerals are the matrix of the ore.

Into the country rocks of the Orijärvi area was intruded what Eskola called oligoclase granite in 1914 but considered to be granodiorite in 1963. Surrounding the western two-thirds of this granodiorite is a contact-metamorphic aureole, not duplicated, except to a minor extent, in the major mass of granodiorite to the south of the Orijärvi body. North(east) of the mass of granodiorite is a belt, some 200 meters wide, of cordierite-anthophyllite rock that, to its north(east), is bordered by a band, some 100 meters wide, of amphibolite, still farther north, the amphibolite is bordered by leptite. The ore and small associated bodies of skarn are located between the amphibolite (northeast) and cordierite-anthophyllite rock with one or two smaller ore lenses being entirely within the latter rock type. The major ore lenses are located at a southward bulge in the amphibolite contact.

The granodiorite, to which Eskola considers the ores (see below) to be genetically related, almost certainly was introduced into the area at the end of the middle Precambrian and during or after the metamorphism of the host rocks of the ore. The host rocks, whether denoted as Svecofennian or

Karelian, are definitely middle Precambrian, and various age dates on the igneous material appear to confirm this. The ore deposits of the Orijärvi area, therefore, are classified as middle Precambrian in age.

Eskola's original concept was that the ore fluids were the same that contact-metamorphosed the leptites and limestones and came from the granodiorite (oligoclase granite). This idea was critized orally by Brögger on the grounds that a granite was not a good source of magnesium; this criticism is less valid now that the granite has been agreed to be a granodiorite. It remains true, however, that a magnesium-rich ore fluid might have been more likely to have come from a magma more mafic than one of granodioritic composition. Geiger (1917) suggested that the ores at Falun (where host rocks, altered wall rocks, and ore are similar to those at Orijärvi) were formed in much the same way as those at Orijärvi. This argues against the concept that the amphibolites were the result of regional metamorphism (at least in the ore area) because the amphibolitized mafic dikes at Falun contain cummingtonite in the ore field but lack it outside of that area. This mineral relationship also occurs at Orijärvi where cummingtonite is a principal constituent of the amphibolite in its border facies.

In 1950, Tuominen and Mikkola reexamined the magnesium-rich rocks in the Orijärvi area and suggested that the cordierite and anthophyllite accumulated in the fold hinges in small anticlines and synclines in the altered layered leptites. They considered that such positional relationships meant that, during folding, silicates of magnesium, iron, and aluminum crystallized at low temperatures, in the presence of considerable water (fluid?) as micas, chlorite, and talc and moved plastically to the fold crests and troughs. The introduction of magnesium and iron forced the transfer of calcium, potassium, sodium, and silicon to the competent leptite layers where they are supposed to have produced feldspars as well as andalusite and sillimanite. These authors also thought that the ore-mineral constituents were introduced in the same manner as the iron and magnesium.

Eskola (1950) thought this concept, to some extent, improved on the hypothesis of a magmatic origin for the ore fluids, but he pointed out that the anthophyllite-cordierite rock and the Orijärvi ore bodies do not occur in a fold crest. Eskola, for this and other reasons such as the lack of correlation between Tuominen's geomagnetic map and the plastic folding shown in the exposures in the mine area, believes that Tuominen's interpretation is greatly exaggerated.

On the basis of the work he did in the Orijärvi area, Eskola put forward his concept of metamorphic facies. It was not until 1939, that Eskola put his ideas on this subject in his final form. At that time, he added three facies to the five he had defined in 1920 to give a total of eight. These eight facies are: (1) sanidinite, critical minerals are sanidine and pigeonite that occur as xenoliths in volcanic rocks, indicating pyrometasomatic conditions (in Lindgren's sense of the term); (2) hornfels (now called pyroxene hornfels) critical assemblages are diopside-hypersthene and orthoclase-andalusite that occur in the inner zones of contact aureoles (typically seen in the Oslo area); (3) amphibolite facies in which hornblende and plagioclase are critical, and this assemblage, occurring at Orijärvi, suggests higher pressures there than in the Oslo pyroxene-hornfels facies, but its development in the outer zones of contact aureoles indicates lower temperatures (Eskola considered water essential to the formation of this facies); (4) greenschist, critical minerals are muscovite-chlorite-quartz and albite-epidote-quartz and the facies is a product of regional metamorphism in the upper levels of the crust [pressure and temperature lower than those facies (1) through (3) above]; (5) eclogite, where the association of omphacite with pyrope-almandine and rutile are critical - the high density of this assemblage indicates formation by extreme metamorphism at great depths and high temperature; (6) granulite, critical minerals are quartz-orthoclase-plagioclase-almandine-hypersthene, while hornblende, micas, wollastonite, and grossularite are absent - probably these assemblages are due to regional metamorphism at high temperatures and pressures in a dry environment (indicated by lack of hydrous minerals), (7) epidote amphibolite (now called albite-epidote amphibolite), critical minerals are quartz-albite-epidote-hornblende, and these formed under conditions intermediate between the greenschist and amphibolite facies; (8) glaucophane schist, critical minerals are glaucophane, crossite, lawsonite, and pumpellyite; although Eskola related this facies to eclogite, other authors consider it related to the greenschist and albite-epidote-amphibolite facies.

Eskola (1939) among others, recognized that disequilibrium assemblages of minerals occur, particularly among low-grade, regional metamorphic facies, in which mixtures of minerals assemblages are developed that belong to more than one facies, the relationships in time being indicated by textural criteria such as pseudomorphs of one mineral after another, cross-cutting veinlets, and partially destroyed relics of probably earlier minerals. Because reaction rates are speeded up by increasing temperature, partial replacement of a higher-grade assemblage by a lower one is more likely than the opposite result. Such higher to lower transitions require increased hydration of the rock involved.

A more detailed discussion of the various metamorphically induced mineral assemblages can be found in Fyfe, Turner, and Verhoogen - Metamorphic Reactions and Metamorphic Facies: Geol. Soc. Amer. Mem. 73, 1958. In this discussion, p. 199-239, the authors break down the various temperature-pressure assemblages into those produced in pelitic, calcareous, mafic, and magnesian, and, in places, quartzo-feldspathic rocks. Pelitic rocks are considered by these authors to be the derivatives of aluminous sediments; quartzo-feldspathic to be those of sandstones and of silicic igneous rocks; calcareous those of limestones, dolomites, and marls; mafic as those of mafic and semimafic igneous rocks, tuffs, and some tuffaceous sediments; and magnesian as those of ultramafic igneous rocks and of highly magnesian sediments. If quartz is present in the metamorphic assemblage, the facies is characterized as silica-rich, if it is absent, as silica-deficient. It seems apparent that each of the five rock types given above, if metamorphosed under the same conditions, will produce sufficiently different mineral assemblages that the parent rock can be recognized and, at the same time, can be assigned to the same metamorphic grade.

Certain divisions of these assemblages into subfacies also were made by the authors at that time, but since then they have been decided that subfacies confuse the issue more than they clarify it. Metamorphic rocks, therefore, that contain the minerals of two adjacent assemblages are designated as transition facies, e.g., greenschist-amphibolite transition facies.

From the point of view of the economic geologist, it is unfortunate that the interest in the concept of metamorphic facies developed from Eskola's work in the district has overshadowed the geologic study of the ore deposits of the area. Although an important part of Eskola's work was based on what was then visible in the open-pit of the Orijärvi mine as it was then; the ore that was in that pit is now completely mined out with more geologic attention having been paid to metamorphism than to ore formation.

On either side of the now-abandoned open pit are both the anthophyllite-cordierite and amphibolite rocks, only the former of which contains any ore. These rocks also contain skarn and limestone intercalations, the skarn having been contact-metamorphically derived from the limestone layers. The skarn is primarily an actinolite-tremolite rock with some of the ore having skarn minerals as matrix materials.

Two types of ore have been recognized. The first is soft ore (blötmalm) that consists of galena, sphalerite, and chalcopyrite, with only accessory amounts of pyrite and pyrrhotite; this ore was developed in the skarn portions of the ore occurrence. The second type, hard ore (hardmalm), is made up of biotite-cordierite-anthophyllite rock and cordierite quartzite; these rocks contain chalcopyrite as equally distributed impregnations or as distinct bands in association with pyrite and pyrrhotite. Sphalerite also is present in this hard ore as patches and small compact masses; galena is present only in minimal amounts. Varma (1954) says that, judging from the appearance of the old open pit, the ore body must have been a connected mass near the surface; this massive body branched downward into projections and lenses. The strike of the ore mass is about east-west, and the dip was about 70°N to a depth of 100 meters. Between 100 meters and 200 meters the dip is southerly at about the same angle; below 200 meters, the dip remains southerly but apparently at a steeper angle than above 200 meters. The ore lens plunges east at from 35° to 50°. Some 100 meters west-northwest of the open pit was a small, isolated lens of ore that consisted principally of sphalerite and pyrite; it extended only about 100 meters below the surface and has been mined out. Further, a weak magnetic anomaly was found along the northern border of the amphibolite north of the open pit; here a narrow zone of sphalerite-galena mineralization was found by diamond drilling; it does not appear to have been worth an attempt to mine it.

The ores at Aijala and Metsämonttu are in the same geological environment as the Orijärvi mine, some 10 kilometers southwest of the latter mine. On the north, the ore zone is bordered by amphibolite and diopside amphibolite; it is intercalated by narrow bands and layers of mica gneisses and mica schists, and some of the mica schist bands may contain arsenopyrite.

On the south, the ore zone is bounded by a quartz porphyry (Varma says blastoporphyritic leptite), then by sericite schist, and, still farther south, by cordierite-mica schists. Pegmatites are present on both sides of the ore zone; they parallel the general strike of the schistosity.

Although both the Aijala and Metsämonttu ore zones trend generally northeast-southwest, the Metsämonttu lenses are definitely northwest of a straight-line continuation of the strike of the Aijala lenses; if, therefore, they are in the same stratigraphic unit, an appreciable fold or fault must lie between the two ore zones. The two ore zones, however, may simply be in the same major rock zone, without being in the same stratigraphic unit. In both zones, the ore occurs mainly, in contradistinction to Orijärvi, as disseminations or breccia cements in limestone or skarn derived from limestone. The dips of the ore bodies, as well as those of the enclosing schists, is quite steep, departing from the vertical by no more that 5° to 10° in either direction. In both areas, the ores are all of lens shape and plunge steeply, but Varma (1954) does not say in what direction, though it more probably is southwest than northeast. The half dozen or so Metsämonttu ore lenses trend generally southwest, but the strike of the most southerly lenses is some 35° more southerly than the major lenses lying northwest of it. The total surface length of the Metsämonttu ore lenses is about 250 meters.

Although the major portion of the Aijala ore is confined to the east (Itämalmi) lens and the west (Länsimalmi) lens, the area of this mine includes about 10 mappable ore lenses on a scale of 1 cm = 100 meters. The total length of the Aijala zone is about 400 meters.

Surprisingly, the mineral composition of Aijala is quite different from that at Metsämonttu, with chalcopyrite being 20 times more abundant at Aijala than at Metsämonttu and sphalerite 7.5 times as abundant at Metsämonttu as at Aijala. Pyrite is nearly twice as abundant at Aijala than at Metsämonttu, whereas pyrrhotite is seven to nine times as abundant at

Metsämonttu as at Aijala. Galena is low at both mines, but it is 10 times as common at Metsämonttu at at Aijala. Total sulphide content at Metsämonttu averages about 10 per cent greater than at Aijala. This difference in mineral composition is difficult to explain, but perhaps no more so than the difference between the soft and hard ores at Orijärvi where the former ore is higher in the lead and zinc sulfides and the latter in chalcopyrite. At Orijärvi, the sphalerite-rich ore has skarn as its host rock and the chalcopyrite-rich has cordierite-anthophyllite rock as its host. In this instance, the difference may be due to the different reactions occurring between the ore fluid and the two host rocks. If this is so, then perhaps the Metsämonttu ore is largely in limestone and skarn and the Aijala ore principally in mica-cordierite-garnet schist. It is difficult to be certain about the host rocks of the Aijala-Metsämonttu ore lenses because of the small scale of Varma's (1954) cross section for the Aijala ore and his lack of one for the Metsämonttu ore.

Latvalahti (1979) points out that the ores in all three deposits are strata-bound in general, but, in detail they fill fractures and cement breccia masses in the host rocks. The Aijala and Metsämonttu ores are somewhat more definitely strata-bound than those at Orijärvi; the former, she observes, occur at the contacts between silicic volcanics and intermediate-mafic pyroclastic volcanites (at Metsämonttu) or close to them (at Aijala). The Orijärvi deposit, on the contrary, is located (as Eskola says also) in a zone of silicic cordierite-sericite and cordierite-anthophyllite rocks. Latvalahti (1979) indicates that there are two mineralized horizons at Orijärvi that lie in the zone of andalusite-, cordierite-, sericite-, and anthophyllite-bearing rocks. The ore body is in the lower mineralized horizon; the upper zone has only a small Zn-Pb-Au-Ag deposit (Illampi) that apparently is too small and/or low grade to be mined. Cordierite- and anthophyllite lenses occur at several levels in the metamorphosed rocks, although she believes that only some of them are in the mineralized horizons. These metamorphosed rocks range widely in composition and texture. Despite the metamorphism they have undergone, some of these rocks still show relics of primary layering. Latvalahti remarks that a hypabyssal variant of the synorogenic granodiorite, with quartz phenocrysts, is about 200 meters southwest of the Orijärvi deposit. The greatest extent of the ore bodies at all three mines is in dolomitic limestones and skarns. Thus, although the bulk of the ore horizon is composed of siliceous metamorphosed rocks, the ore favors those that still are carbonates or have a carbonate ancestry.

In the three ore deposits, the actual ore bodies consist of several small, narrow, and elongated ore bodies in which the major axis is parallel to the b-lineation of the second phase of folding. The ore bodies are oriented either side by side or en echelon; the orientation depends on the relations of the bodies to the fold structures. At Orijärvi, the axes of the ore bodies dip 45-50° NE; further, at Orijärvi, the deposit is cut by a basaltic subvolcanic dike that greatly complicates the relationships of the ore masses to each other. At Aijala and Metsämonttu, the ore bodies dip more steeply, 80-85° SW. The ore bodies (Latvalahti, 1979) are cut by faults and shear zones that brecciate the ore bodies; dense joint networks, in many places, are filled with zeolite veinlets or chlorite-coated slickensides. The largest fault displacements are at Metsämonttu where unmetamorphosed clay coats the fault surfaces.

The ore bodies of the Aijala area conform to the axial plane of foliation of the second-stage folds. At Metsämonttu, the dolomite skarns and chlorite-bearing tremolite-diopside skarns become less abundant with depth; on the contrary, cordierite-anthophyllite rocks and cordierite-mica gneisses are more common with depth than they were at the surface. By the 510 level, the cordierite-bearing rocks dominate over the skarns. At this level, the grade and size of the ore bodies decrease markedly. The ores at all three deposits are thickest (up to 20 meters) in the skarns, but normally they are less than 10 meters; their strike lengths range from 100 to 150 meters.

The Orijärvi deposit is rather like that at Metsämonttu in wall rocks and wall-rock alteration, distribution of metals, and their modes of occurrence. The Orijärvi ore bodies are located largely in chlorite-bearing tremolite-diopside skarns; quartz- and anthophyllite-bearing rocks are present in the walls of the ore bodies but contain essentially no bodies of economic value. The Orijärvi ores are more broken and smaller than those at Metsämonttu.

Latvalahti (1979) classes the ores at all three deposits as: (1) breccia, (2) massive-vein, and (3) disseminations. Of these, those of type (1) are economically the most important. They clearly cut, brecciate, and replace the wall rock but are generally conformable to the bedding and have sharper contacts with those rocks in structure and metal content. Although the breccia ores normally are skarns and dolomite limestones, minor amounts are in quartz-rich rocks; they include only a few wall-rock fragments of small size. The ores replace amphiboles and pyroxenes along cleavages and brecciate quartz-phlogopite masses. The richest of the ore types in lead and silver are breccia ores in skarn and dolomite limestones; iron sulfides predominate in those in quartz and cordierite rocks, but some chalcopyrite usually is present in these.

The (2) ores form massive sulfide veins with widths from a few centimeters to several meters. Their major minerals are pyrrhotite and pyrite; they have sharp contacts with the wall rocks and locally include wall-rock fragments. Shear zones containing chlorite may be filled with veins of iron sulfides (Latvalahti says they are remobilized sulfides but does not say why) or coarse-grained disseminations of sulfides. Disseminated sulfides also are in places associated with breccia ores in which they replace amphibole and pyroxene. Iron sulfides may be disseminated in various cordierite and mica gneisses.

The chemical composition of the ores ranges widely. The ores at Metsämonttu and Orijärvi are much the same, but the latter ores are less discrete, forming broken, narrow, and discontinuous ore bodies. In both areas, ores in limestones and skarns have a higher (Zn,Pb)/Cu ratio than those in cordierite rocks but at Orijärvi copper is more common in cordierite than it is at Metsämonttu. In this latter deposit, a copper ore body in cordierite gneiss continues in slightly mineralized form to the Aijala deposit. Zinc-lead, zinc-iron, and copper ore bodies are present; zinc-lead ores are located in chlorite-bearing diopside skarns and dolomitic limestones, zinc-iron bodies in mica and cordierite gneisses, and copper ore bodies in cordierite gneiss.

The Aijala ore bodies are principally composed of chalcopyrite (mainly) and pyrite. The deposits down to the 220 level are predominantly chalcopyrite but with some sphalerite, galena, arsenopyrite, and iron sulfides in varied grades. The margins of the Aijala ore bodies are richer in iron than the cores. Most of the copper ore bodies grade into pyrite ores with depth.

The deposits of all three districts have simple mineralogies with pyrrhotite, pyrite, chalcopyrite, sphalerite, and galena as the major minerals; their relative abundances differ from one deposit to the next and from one part of a single deposit to another. Latvalahti says that ore in these various loci both fill open space and replace gangue minerals. She says that fault and shear zones have affected the ore, partly remobilizing and replacing it. She does not give her reasons for this belief, although the concept follows logically from her faith that the ores were originally syngenetic and strata-bound, and the only way they can show replacing and cross-cutting textures is by remobilization and resolution and replacement. Of course, such statements as she makes do not remove the possibility that the replacement and open-space filling resulted from the reactions accomlished by the solutions that also deposited the strata-bound ores. The "original" banded textures (in Latvalahti's opinion) in the ores, therefore, may be the result of replacement rather than of primary deposition. Although certainly the primary ores were somewhat affected by metamorphism attendent on the second phase of folding.

In addition to the minerals already mentioned, the ores also contain cubanite, magnetite, tetrahedrite-tennantite (fahlerz), native silver, and various sulfo-salts, the latter as accessories in association with galena. At Metsämonttu, sphalerite in many places contains exsolution bodies of chalcopyrite and pyrrhotite. Boulangerite is present as inclusions in galena; chalcopyrite contains cubanite lamellae and tetrahedrite inclusions. Orijärvi ore contains, normally, more chalcopyrite than that at Metsämonttu enough to class it as a major mineral. Orijärvi contains more high-temperature minerals (ilmenite, magnetite, gudmundite, molybdenite, and uranitite) than the other two districts.

Below the ore at both Orijärvi and Metsämonttu funnel-shaped alteration pipes of cordierite-anthophyllite rocks are present below and connected with ore above. Further, largely surrounding the ore deposits are alteration envelopes of cordierite-bearing sericite and muscovite schists that, Latvalahti says, are located close to the volcanic discharge channels. Folding, faulting, and "remobilization," Latvalahti believes, have to some extent affected the original locations of ore relative to alteration zones. These relations she cannot imagine as being the result of anything other than remobilization of primary ores and of metamorphism-induced wall-rock alteration.

The Aijala and Metsämonttu ore deposits have a common outer alteration zone that is about three kilometers long. Its maximum width is 100 meters, but it drops down to 0.5 to two meters in the areas between the two deposits. The rocks in the outer margins of this alteration zone have been sericitized, exactly the effect that would be expected from the ore-bearing solutions working their way outward after depositing the ore bodies of the two deposits. Latvalahti considers that the internal heterogeneity of the inner portions of the alteration zones in all three deposits largely is due to the original heterogeneity of the rocks that host the ore, but she also thinks that differences in the altering fluids from one point to another may have been partly responsible for these variations. The alteration of the inner zone is primarily the magnesium alteration that Eskola recognized early in this century. The principal types of rocks in the inner zone are dolomitic limestones and chlorite-bearing tremolite-diopside skarns derived from limestones and cordierite-biotite gneisses and cordierite-anthophyllite rocks produced from silicic tuffs with quartz and plagioclase phenocrysts. Quartz-rich rocks are the result of silicification of silicic rocks.

The quartz-rich rocks and the chlorite-bearing tremolite-diopside skarns occur only in the wall rocks of the ore bodies, whereas, the cordierite-anthophyllite rocks and the cordierite-bearing gneisses make up the alteration pipes below the Orijärvi and Metsämonttu deposits. The former pipe is known to go down 250 meters, without bottom having been found, and the better explored latter pipe reaches at least 600 meters beneath the ore. No alteration pipe is known at Aijala.

In the inner alteration zone, some retrograde metamorphism is shown by sericitization of plagioclase and pinnitization of cordierite.

Eskola still considered, in 1950, that the ores were produced by hypothermal mineralization by ore fluids derived from the synkinematic granodiorite. The same ore fluids, in his opinion, also produced the various types of wall-rock alteration surrounding and intermixed with the ores. On the contrary, Latvalahti (1979) appears convinced the ores were deposited on the sea floor and that the altered pipes were produced by the same solutions during their upward journey. How she explains the envelopes of alteration around the ores is less clear. These alteration zones could not have been formed at the same time as the ores were spewed out on the sea floor, because rocks to be altered in this manner did not exist at that time around the ores. She can explain these envelopes only by reactions generated in the volcanics (largely subsequent to the ores) by the metamorphism attendent on the second stage of folding. To me, even after the lapse of so many years and the present popularity of volcano-genic processes, it seems that Eskola is more nearly correct than Latvalahti.

The contact metamorphism of the host rocks of the ores in the Orijärvi district appears (to me) to have taken place under hypothermal conditions, and, since the ores are found in both calcareous and noncalcareous rocks, the wall-rock alteration alone would be categorized as hypothermal-1 and -2. If, however, it could be shown that the ore sulfides were emplaced under sufficiently lower temperatures than the gangue minerals, the hypothermal category would not be used in the classification scheme employed in these volumes. The presence of cubanite [Fe_2CuS_3] in the Orijärvi ore suggests that the chalcopyrite, with which it is closely associated in time and space, at least formed under hypothermal conditions. The presence of exsolution textures of chalcopyrite and pyrrhotite in sphalerite adds confirmation to the proper place of the Orijärvi area deposits in the modified Lindgren classification being hypothermal. What is known, including the major amounts of pyrrhotite, the small content of galena, and the presence of gudmundite [FeSbS] in the deposit, indicates nothing against the deposit having been formed under hypothermal conditions and probably goes a considerable way to confirm it. The Orijärvi deposits, therefore, are here categorized as Hypothermal-1 and -2. The Aijala and Metsämonttu deposits are similar enough, both in gangue and ore minerals, to those at Orijärvi that these two groups of ore bodies also should be classified as Hypothermal-1 and -2. The minor amounts of sulfosalts associated with galena suggests that some of the ore was formed under mesothermal conditions.

OTANMÄKI

Middle Precambrian	Iron as Magnetite	Magmatic-3b
	Titanium as Ilmenite	Metamorphic-C
	Vanadium in Magnetite	

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Notes

The central-Finland district of Otanmäki (64°07'N, 27°06'E) is about 5 km south of a major lake (Oulujärvi). The nearest town of any size is Kajaani, 32 km to the NNE. Otanmäki is another of the ore deposits in Finland to have been discovered by tracing the source of mineralized glacial boulders (ilmenite-magnetite ones in this case) for some 40 km to the NNE. These boulders were found in 1937 and pointed the way to the locating of a 500,000 m^2 magnetic anomaly on Otanmäki Hill; diamond drilling was begun in 1939. In addition to Otanmäki, this geological work also found the ore-bearing areas of Vuorokas, Pentinpuro, Isonkivenkangas, Honkamäki, and Itäranta, all within 12 km of Otanmäki.

The mining of ore and the preparation of magnetite and ilmenite concentrates began at Otanmäki in 1953. Of the five ore-bearing areas just mentioned, mining began at Vuorokas in 1976 with Pentinpuro and Honkamäki still being explored at that time. The annual production at Otanmäki in 1980 was 1 million tons of ore that contained 2400 tons of V_2O_5, 265,000 tons of iron-concentrate pellets (from which the V_2O_5 was leached), 135,000 tons of ilmenite concentrates, and 6000 tons of pyrite concentrates. Pyrrhotite apparently contributes significantly to the pyrite concentrates. In places, sulfides amount to as much as 20 per cent of the ore, but none of these (except the iron sulfides) is worth recovering; chalcopyrite is the most abundant of the non-ferrous sulfides. In 1978, reserves were reported to be 38 million tons with a grade of 22 per cent iron, 5 per cent titanium, and 0.2 per cent vanadium. Plans have been made to raise production to 1.5 million tons each year; this ore will contain 3000 tons of V_2O_5.

A state-owned company, Otanmäki Oy, was founded to exploit the deposit; in 1968, this company was merged with Rautaruuki Oy, the major state-owned steel producer in Finland. The magnetite concentrate is delivered to this company's furnaces where it is used to make pig iron and steel. Vanadium, the most valuable product of the Otanmäki operation, is removed by leaching from the magnetite concentrates; the V_2O_5 mainly is exported. Ilmenite and pyrite are recovered by flotation.

Because all the bedrock of the area is covered by 1 to 3 m of glacial debris, surface and underground diamond and percussion drilling were and are the major exploration tools. The original surface drilling located 10 million tons of ore in numerous separate lenses, each of which contained an average of 100,000 tons; further exploration raised the number of lenses found to over 200 by 1960.

The basement rocks of the Otanmäki district mainly are gneissic granites that belong to the lower Precambrian (pre-Svecokarelian or Katarchean); they are by far the oldest rocks in the region. In the southern and western parts of the district, the basement includes striped gneisses. These gneisses are structurally most heterogeneous and contain inclusions of amphibolites (originally interleaved with the gneisses as mafic volcanics but since altered to their present state), veins of pegmatitic granite, and fragments of mica gneiss. The outcrops of these gneisses and their inclusions are so sparse that even today their exact age and extent are uncertain. The gneisses range in texture from fine- to medium-grained gneissose granites to strongly compresses and contorted veined gneiss.

The Otanmäki formation that contains the ore bodies of the district consists of gabbro (more or less altered to amphibolite) and anorthosite. This formation is located near the eastern end of a zone of deformation that is more than 100 km in length. This zone cuts through not only the basement gneisses but also through the basin-shaped mass of supracrustal rocks that overlie the basement and are mainly quartz-feldspar schists, mica schists, and metavolcanics. The zone of deformation strikes northwest-southeast and cuts across Lake Oulu. Where this deformation is most marked, the Otanmäki complex is most strongly differentiated.

The Vuorokas ore zone almost certainly is in the same rock-complex as the Otanmäki ore zone proper and probably is a faulted continuation of the Otanmäki ores. The Honkamäki and Pentinpuro prospects are in similar rocks located on a separated linear magnetic anomaly that runs NNE with Honkamäki about 10 km WNW from Otanmäki (at the southern end of this second anomaly) and Pentinpuro about 12 km northwest of Otanmäki at the northern end (of this second anomaly). Isokivenkangas is located on the second anomaly also but is appreciably nearer Pentinpuro than Honkamäki. If any structural connection exists between the ore prospects of the second linear anomaly and those on the Otanmäki anomaly, it is not readily apparent on the aeromagnetic maps available to me.

The only amphibolite of any economic importance in the Otanmäki district is that produced from the alteration of gabbro that was one phase produced by differentiation of the mafic magma intruded into the district long after some of the mafic volcanics in the basement had been converted to amphibolite. Although two ages of amphibolite almost certainly exist, there is no reason to confuse one type with the other.

The predominant light mineral of the Otanmäki amphibolite is plagioclase (ca. An_{35} on the average), whereas the main dark mineral is hornblende. The most common type of amphibolite is a dark rock that contains abundant hornblende, but some of the amphibolite is plagioclase-rich and much lighter. Some of the plagioclase in the various amphibolites is sericitized. The hornblende normally is arranged in closely spaced streaks, and some biotite and apatite are known in this rock type. In addition to sericite, epidote, chlorite, and zoisite occur as alteration products. Traces of pyrite and occasional grains of ilmenite, surrounded by small grains of sphene, have been seen in the amphibolite. The schistosity of the amphibolite is quite distinct, and quartz is present locally in the rock and probably is secondary.

None of the amphibolite in the area of the Otanmäki Hill is of the ancient variety; all of it developed from a much younger gabbro that was introduced synchronously with the Karelian orogeny. This gabbro with which the ore is directly associated has been radiometrically dated (from zircons in the gabbro) as 2060 m.y. A microcline granite that is extensively intruded into the eastern part of the district is dated by the same method as 2000 m.y., and it cuts all rocks in the area. Thus, the ores must have been introduced in this interval of 60 m.y., probably essentially co-equal in time with the intrusion of the gabbro. The age of the Otanmäki ore, therefore, is middle Precambrian, about that of Vihanti and Pyhäsalmi but appreciably older than the sulfide deposits in the Kotalahti nickel belt.

Lindholm and Anttonen (1980) say that "vanadium-bearing magnetite-ilmenite ores are associated with layered gabbro-anorthosite intrusives," not only at Otanmäki but also at Vuorokas, Honkamäki, Pentinpuro, and Isoivenkangas. The gabbros of this association have been converted to a large extent to amphibolites composed (in addition to the original plagioclase) of uralitic hornblende with relict olivine and pyroxene, "The ore zone is located (Lindholm and Anttonen, 1980) in a heterogeneous ortho-amphibolite zone which is the contact zone between gabbro and anorthosite." This heterogeneous zone lies on the northern flank of a large mass of hornblende gabbro and is bordered on the north by pink alkalic gneiss. South of the gabbro is striped gneiss that, close to the contact with the gabbro, is strongly schistose.

This heterogeneous ore zone is about 2 km long (east to west) and, at its maximum width is about 600 m (north to south). The orthoamphibolite zone consists of metagabbros, gabbros, and anorthosites, the metagabbros being, essentially, amphibolites. The ore lenses lie in this heterogeneous amphibolite with most of the ore lenses being concentrated along its southern and eastern margins. From the northern end of the eastern margin, the ore curves gradually back to an east-west direction in which it continues for about 800 m to a point about 300 m to the east of the western end of the southern-margin ore lenses. Some of the ore near the north border of the orthoamphibolite is a few tens of meters south of the contact with the gneissose granite to the north of the ore zone but appears to be connected by a narrow, vein-like structure with the ore lenses (or vein) along the actual contact. Intermingled with the ore lenses are many masses of anorthosite of considerable size. Even larger bodies of anorthosite lie in the center of the ore oval, many of them at an appreciable distance from any ore.

The Otanmäki complex of ore lenses, anorthosite, and gabbro is contained in a roughly east-west-trending zone, being about, as stated, 2.0 km long and 0.6 km wide. The ore lenses are discontinuous both in plan and in section; some of them are in direct contact, certainly in section and probably in plan, with anorthosite bodies. There does not, however, seem to be any direct relationship in size between amphibolite bodies and the ore lenses in contact with them. In some places between ore and amphibolite, reaction rims (usually 10 cm wide but occasionally as much as a meter in width) are developed. These are composed of garnet and epidote. Where anorthosite xenoliths are included in the ore, the margins of these consist of epidote and lesser amounts of very dark chlorite. Feldspar associated with these reaction zones may take on a pink-orange color.

If, as Lindholm and Anttonen say, this now orthoamphibolite (with irregularly arranged anorthosite and ore-lens inclusions) was initially a layered gabbro-anorthosite body, post-crystallization deformation must have broken the anorthosite layers into masses of varied sizes and shapes. Certainly, the patterns shown in Lindholm and Anttonen's sections can not be those of a layered intrusion unaffected by strong deformation. The probability of such deformation having taken place is shown by the relationships among the rock and ore minerals. The amphibolite contains two types of hornblende, earlier, moderate-sized, ragged grains of that mineral with which are associated smaller, clear hornblende grains. The earlier plagioclase grains are larger than the later ones, the latter being clear as opposed to cloudy and lower in An-content than is the earlier feldspar. Added to this, intergrowths of ilmenite and magnetite are rare; both minerals occur as independent grains that easily are separated in the concentration process. Thus, during deformation, ilmenite in solid solution in magnetite was driven out to form nearly pure grains of $FeTiO_3$. If these small-scale effects were produced by such deforming forces as acted on the "layered gabbro-anorthosite intrusion," then it seems quite probable that these forces also would have been capable of fracturing the more brittle and layered anorthosite and of distributing these broken blocks into the irregular pattern they now display. There is a distinct flow structure in the ore around the anorthosite masses.

If this reasoning is correct, it is almost certain that the layered intrusive occupied the entire rock volume that contains the ore lenses and that it probably was emplaced before the ore pods were intruded into the heterogeneous mixture of altered gabbros (now amphibolites, but once gabbros that ranged the gamut from leuco- to mela-gabbros) and blocks of fractured anorthosite of a wide variety of sizes and shapes.

The plagioclase content of the amphibolite normally ranges between 35 to 50 per cent, but as much as 70 per cent has been observed in some places. Locally, however, the amphibolite is made up almost entirely of hornblende. This high hornblende rock is schistose enough to be designated as hornblende schist. The contact between the various types of amphibolite, with their various percentages of plagioclase, are highly gradational. Further, the amphibolite, in some places, is light green due to alteration to epidote. In amphibolite high in feldspar, the color is much lighter than normal.

The relationships of ore to amphibolite and anorthosite appear to be the same in the similar, but much smaller, Fe-Ti-V deposits at the other four localities mentioned above, indicating that the same system of rock and ore formation was repeated in the area at least as many times as there are ore bodies present in it.

In the Otanmäki area proper, the amphibolite is intruded by two later varieties of granite. One of these is an alkalic granite gneiss, a large, very heterogeneous mass of rock that appears (Lindholm and Anttonen, 1980) to have under gone appreciable metasomatism. It has been considerably metasomatized, as its mineral composition suggests. It contains not only dominant quartz and feldspar, but also alkalic amphibole, alkalic pyroxene, abundant fluorspar, zircon, and sphene, local carbonate, and many rare minerals such as columbite, basnaesite, danalite [$Fe_4^2Be_3(SiO_4)_3S$], thorite, and enigmatite [$Na_2Fe_5^2TiSi_6O_{20}$].

In the eastern section of the geological map of Lindholm and Anttonen, there are extensive areas of another microcline granite intrusive (Kajani granite); farther east it is bounded by the Karelian schist belt.

The Otanmäki and Vurokas gabbros are dated from their zircons as 2060 m.y. old. The alkali gneiss at Honkamäki is about 2000 m.y. old. Veins of this alkali gneiss cut into the amphibolite, and amphibolite fragments form xenoliths in the granite.

The volume of ore-containing rock includes over 200 separate ore pods with the largest of these bodies being some 200 m long and 3 to 50 m wide. The dip of the pods is now vertical or northerly along the southern side of the hill, but, along the eastern edge of the ore area, the departure from the vertical gradually changes with distance from steeply north to steeply west. The average pitch of the ore bodies is 50°W; this also is the pitch of the entire ore-containing rock volume as is indicated by geophysical measurements. Because the ore lenses could hardly have settled out of their molten source material in their present more-or-less vertical positions, this is another argument in favor of the ore volume having undergone deformation after the introduction of the Fe-Ti-V melt. This means that the originally flat-lying ore lenses and their associated igneous rocks must have been rotated through about 90° with the tops of the former layered lenses having been moved in directions that ranged from north to west, depending on their location in the folded area. This would indicate that the structure of the ore volume is synclinal rather than anticlinal. The Geological Staff (1960) simply says that the structure may have been folded.

The ore lenses definitely are associated with the sheared portions of the amphibolite-anorthosite complex with many of the high-grade lenses being directly at the contact with gabbro and anorthosite lenses. The contacts of the ore lenses with the gabbros normally are sharp, although, on the southern side close to the contact, the gabbro contains considerable magnetite and ilmenite but the contact with anorthosite generally is brecciated. The ore contains fragments not only of the anorthosite but of the amphibolite as well. Most of the ore lenses are located on the south sides of the anorthosite-gabbro clusters. Despite the high proportion of amphibolite to unaltered gabbro in the ore zone, such masses of gabbro do exist. Perhaps these unaltered gabbro masses avoided the general and thorough-going amphibolitization of the bulk of the gabbro by having been intruded after this alteration had occurred.

In places, the ore has penetrated the country rock along planes of schistosity, and amphibolite fragments in the ore have been penetrated by ore along their planes of schistosity. Thus, the host rock certainly was schistose when the ore was finally emplaced. In places, particularly in low-grade ore, schistosity cuts across the ore.

According to Pöschl (1964), the Otanmäki ore is composed of three types: (1) impregnation ores; (2) banded ores; and (3) massive ores. As would be expected, the impregnation ores are of the lowest grade; in this type magnetite and ilmentite make up less than 20 per cent of the total weight. The host rock of these impregnations is amphibolite (how these impregnations get into the amphibolite and where the amphibolite came from is discussed below).

The banded ores are typical of Otanmäki, and this structure is thought to have resulted from a rhythmic alteration in the precipitation of high- and low-grade material. The massive ore is what its name implies, a massive mixture of magnetite and ilmenite crystals; the gangue minerals are chlorite and hornblende. The hornblende is more abundant to the east and the chlorite to the west. Also the ore is lower grade toward the east. The pyrite is the more abundant iron sulfide to the west and pyrrhotite is so to the east; the change is gradual and some of both sulfides are present in most of the ore.

From the richest (massive) portions of the ore, tongues of the ore material have intruded into the amphibolite parallel to its planes of schistosity. In low-grade ore, where the banding is well preserved, the schistosity cuts through the ore and probably developed during or after its emplacement. The massive ore is composed of both coarse- and fine-grained-saccharoidal types that cut across the ore bands.

The lack of any uniform relationship between the size of the anorthosite masses and the ore lenses that, in places, are in sharp contact with them, indicates that the ore was not crystallized (or separated in the molten state) from the magma that produced the originally layered anorthosite. Nor does the gabbro from which the amphibolite was developed show any evidence of having been the direct parent of the ore pods. Rather, it seems that the various ore lenses were intruded into the mixture of broken anorthosite and amphibitolized gabbro as pods of material immiscible in the parent gabbro-anorthosite magma that were retained in the original magma chamber appreciably after the gabbro-anorthosite magma had been introduced into rock volume now containing the ore lenses. Probably these pods contained varied amounts of silicates dissolved in them, depending on when they left the magma chamber where they were generated. Those that left that chamber early were those that eventually formed the low-grade impregnation ores. These were so low in magnetite and ilmenite (less than 20 per cent of the ore pods of this type) that banding of the ore minerals did not develop during their primary crystallization nor later when they underwent the latter stages of the dynamic metamorphic effects. Pods of higher grade ore material that left the magma chamber somewhat later, with a lower silicate content and a higher one of ore elements, crystallized either as banded ore in the main or as the more richly disseminated ore that was turned into banded ore by the still later results of the dynamic metamorphism. The pods that turned into massive ore lenses contained little silicate material. The massive ore contains, in addition to magnetite and ilmenite, minor amounts of hornblende (to the east) and chlorite (to the west); these two minerals were derived from the minor amounts of molten silicate material dissolved in the Fe-Ti-V melt. Some biotite was a late mineral in some of the massive ore.

The average size of the magnetite grains is 0.2 to 0.8 mm; in a few places in the high-grade ore, however, the size of the grains may be well over 1 mm. Lineation is not apparent in the massive ore (probably because this ore does not hold enough silicate minerals to provide layers contrasting with the ore minerals).

The blocks of broken anorthosite in the gabbro-derived amphibolite also contain minor amounts of the ore minerals, particularly ilmenite. And some of the amphibolite includes minor disseminated ore minerals that probably were primary crystallizations from the various types of the later gabbro. The high-grade ore contains 35 to 40 per cent of magnetite (with vanadium) and ilmenite between 28 and 30 per cent. In the lower-grade ore, the proportions of ilmenite to magnetite change with the disseminated ore having twice as much ilmenite as magnetite. Also in the disseminated ore, the proportions of plagioclase to hornblende differ widely. The major sulfides, pyrite, pyrrhotite, and chalcopyrite, are not of econmic importance, but locally they amount to as much as 20 per cent of the mass of host rock and ore.

Although most magnetite normally lacks inclusions of ilmenite (and vice versa), in a few places, such inclusions are abundant; these probably were areas where, for one reason or another, the effects of deformation in promoting exsolution were not felt to any effective extent. In magnetite from which ilmentite does exsolve, a later $MgAl_2O_4$ spinel also may exsolve after the ilmenite has done so. It is so small in amount as to have nothing more than vague scientific interest. Vanadium in the magnetite, on the contrary, is of major economic importance as the vanadium content of the magnetite averages 0.90 per cent V_2O_5 and ranges between 0.80 and 1.05 per cent. So far as has been reported, no evidence of exsolution of coulsonite ($Fe^2V^3_2O_4$) has been reported, indicating that the solubility of coulsonite in magnetite is something above the percentages of V_2O_5 reported for Otanmäki. In the layered ore, no differences between vandium content in the various layers has been noted. The magnetite also averages 0.03 per cent chromium and between 0.06 and 0.12 per cent zinc.

In high-grade ore, the sizes of the ilmenite grains are about the same as those of magnetite, but in low-grade disseminations of the ore, a large number of ilmenite grains are less than 0.15 mm in diameter. Rounded hematite exsolution bodies are present in ilmenite, being about 4 to 5 microns in diameter. In the coarsest hematite-exsolution bodies, ilmenite exsolves from the Fe_2O_3. Some magnetite exsolution lamellae also are seen in the ilmenite; the magnetite may have been produced by shearing-induced reduction of hematite.

Lindholm and Anttonen (1980) believe that the Otanmäki complex is to be considered as a magmatic intrusion in which the ore bodies are in their original sites in relation to the nearest wall rocks. It is not certain if they believe that metamorphism participated in the formation of the deposits, although how the major anorthosite layers could have been so broken and blocks so produced moved around relative to each other without appreciable metamorphic effects is unclear to me at least. They also say that the pattern of the geologic map of Otanmäki changes with the progress made in the geologic study of the ore district. An examination of the various maps presented between 1954 (Pääkkönen) and Lindholm and Anttonen (1980) certainly show this to be so.

I do, however, believe in the applicability of the scheme I outlined. This is that of the layered gabbro-anorthosite having first been produced with essentially no magnetite-ilmenite having been involved and with the later introduction of those minerals (and their associated accessories) in the molten state into the solidified and metamorphosed layered gabbro-anorthosite rocks. The location of the ores without appropriate volume relations to the anorthosite bodies makes it highly probable that the ores came from a magma chamber at depth. It is likely that such disparities in size ratios between ore lenses and anorthosite blocks means that the two materials could not have been developed in the same milieu at much the same time. So many magmatic sulfide deposits contain intrusion of molten sulfides that entered the rock system after all silicates (and minor sulfide accessories) had crystallized that it seems reasonable that the same can be true of molten oxides. The contact of many of the ore lenses, but far from all of them, along the margins of anorthosite blocks suggests to me that these contacts provided the easiest channels of movement of the ore melts, particularly when the rock-ore system is considered in three dimensions.

On the basis of this concept of their formation, the Otanmäki ores are here classified as Magmatic-3b, Metamorphic-C.

Note: It is 25 years since I visited Otanmäki under the guidance of Ole Lindholm, and I regret having to disagree with such an excellent guide and host. Our disagreement is, however, minor, and further work on the ores should remove it.

OUTOKUMPU GROUP*

Middle Precambrian	Copper, Pyrites Cobalt, Nickel	Hypothermal-1 to Mesothermal Metamorphic-C

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Notes

The deposits of the Outokumpu area of North Karelia include the Outokumpu mine, proper, the Vuonos mine, 4 km northeast of Outokumpu, and the Luikonlahti mine, 27 km northwest of Outokumpu. One other mine now produces in North Karelia, the Hammaslahti mine, 55 km southeast of Outokumpu; it is different from the other three mines in geology, mineralogy, and chemical characteristics and is not discussed further under this heading. The Outokumpu (Strange Hill) deposit was discovered in 1910 through the finding of a 20-ton boulder and tracing it back to the locale of its glassy quartzite host rock, 50 km to the northwest. The Vuonos deposit was not found until the 1960s; it being a blind ore body, lithogeochemical methods were needed to find it. The outcrop of the Luikonlahti mine was known since the time of World War I, but since the grade in copper was only a little over 25 per cent of that at Outokumpu, production was begun only in 1972. The ore at Outokumpu and Luikonlahti is essentially the same, whereas at Vuonos not only is the Outokumpu ore-type present, but there also is a low-grade concentration (0.20 Ni) of nickel ore in the hanging wall of the copper-cobalt ore body.

Although the nickel-cobalt ratio in the serpentinites surrounding the Outokumpu ore is 20 to 25:1, essentially no copper is present in the altered ultramafic, but anomolous copper-cobalt halos exist around the Outokumpu ores. This fact was used to re-evaluate old drill core, and indications of the Vuonos ore body were thus found, and the body itself was cut by drilling in 1965.

The North Karelia area of Finland is one in which Precambrian outcrops are abundant; this, added to copper deposits already known in the area, make it a favorite target for Finnish exploration efforts.

Total reserves of copper metal at Outokumpu are about 1 million tons; between 1913 and 1974, 22,657,000 tons of ore (grade 3.51% Cu) were mined and 426,000 tons in 1978. At Vuonos, between 1970 and 1974, 603,000 tons of ore (grade 2.1% Cu) were mined and 600,000 tons of ore in 1978. Some nickel ore was mined at Vuonos between 1972 and 1976. At Luikonlahti, 3,408,000 tons of ore (grade 0.97% Cu) were mined.

The entire North Karelian area is underlain by a Presvecofennian basement composed mainly of granodioritic gneisses; these rocks outcrop in the southeastern part of the district with the contact with the younger rocks running along a north-northeast line well to the east of the Outokumpu district proper. These younger (Jatulian) rocks are of epicontinental origin and lie with a marked unconformity on the basement; the Jatulian rocks consist of basal conglomerates and arkosic quartzites that were followed by much purer epicontinental quartzites. These epicontinental rocks, in turn, are covered by a thick accumulation of geosynclinal Kalerian sediments that are made up mainly of phyllites and mica schists. The mineralogical and chemical compositions (Peltola, 1980) of these mica-bearing schists indicate that they were formed from incompletely weathered debris and probably were flysch sediments in the Karelian geosyncline.

In the western part of the Karelian schist area is a long, ribbon-like zone that is composed mainly of serpentinites, carbonaceous black schists, and quartzitic rocks. These quartzites are appreciably different from the Jatulian epicontinental ones and are best designated as Outokumpu quartzites, and the entire sequence of the ribbon body is called the Outokumpu zone. This zone is intensely folded, and the sinuous ribbon conforms to the enclosing mica schists and curves around the Prekarelian gneiss domes that rise through the mica schists. As the Karelian rocks are followed from east to west, their metamorphic grade increases as the phyllites and mica schists grade into migmatitic mica gneisses and granites toward the west.

In the eastern part of the Outokumpu district, the beds dip to the west, and the folding is isoclinal and overturned so that the fold axes dip east. In the western portion of the district, the major Maarianvaara granite cuts through the Karelian rocks and outcrops over a large area. Two major anticlines in this part of the district were developed around two domes - the Maairanvaara (to the north) and the Juojärvi (to the south). Between these two domes on the west and the Sotkuma dome to the southeast is the Outokumpu synclinorium; the latter two domes are composed of basement gneiss and the Maarianvaara of the granite of that name. In the Outokumpu synclinorium the main plunge is to the southwest; local reversals, however, have been produced by local axial depressions and culminations. The Outokumpu and Vuonos ore bodies are in an axial depression in which the bedding dips southeast.

Pegmatite dikes given off in the late stages of the intrusion of the Maarianvaara granite cut not only the surrounding gneisses and schists but also reach into the Outokumpu zone itself.

All of the Outokumpu-type deposits are located in the mica schist of the Outokumpu zone. Three types of mica schists are recognized: (1) bedded, (2) homogeneous, and (3) cataclastic; in these rock types may be layers of black schist that are actually mica schist layers containing more than one per cent of both graphite and sulfur. In these mica schists are major masses of ultramafic material now largely serpentinite; these ultramafics may have been intruded into the mica schists or they may have been poured out as lavas on the sea floor while the original sediment of the mica schist was being deposited. These altered ultramafics contain a core of serpentinite surrounded from the inside out by (1) carbonate-bearing serpentinite;

(2) carbonate rock; (3) tremolite skarn; (4) diopside skarn; and (5) quartzite. The copper-cobalt ores of the Outokumpu zone are localized in quartzites, skarns, and carbonate rocks - all these are surrounded by mica schists. The nickel mineralization of the Vuonos type is in quartzites and skarns and in the black schist where that rock is at the contact with serpentinite.

The Maarianvaara granite occupies much of the area between Outokumpu-Vuonos (southeast) and Luikonlahti (northwest); its age is estimated at 1850 to 1800 m.y.; therefore, it is late middle Precambrian.

The epicontinental sediments range from 1 to 130 m in thickness, and these rocks are composed of various now metamorphosed rocks plus weathering breccias and conglomerates. Although there is some general resemblance between these metamorphosed rocks and those surrounding the serpentinites, in detail they are readily distinguished.

In addition to lying exposed at the surface to the north-northeast of the Karelian mica schist, the basement complex also underlies the Karelian formation at depth. A few major and some minor domes of basement gneiss also pierce through the Karelian schist and are exposed to the east and southwest of the Outokumpu-Vuonos area. Later than the Karelian schist is the Maarianvaara granodiorite-granite that occupies much of the area between Outokumpu-Vuonos (southeast) and Luikonlahti (northwest); its age is estimated at 1800-1850 m.y. and is, therefore, late middle Precambrian. This granite is reported to cross-cut the ore at Outokumpu.

The Karelian schists in North Karelia are subdivided into the Jatulian (older) and the Karelian sedimentary (younger) groups. The Jatulian rocks are mainly well-sorted and show the effects of strong chemical weathering; they are cut by tholeiitic and spilitic lavas and by hypabyssal intrusions. The Karelian beds are mainly turbidites into which intrusive mafic igneous rocks, now highly altered, have been introduced.

The mica schists cover most of the surface in the area of the Outokumpu-type ore deposits and have a somewhat varied mineral composition, but average about 30 per cent quartz, 35 per cent plagioclase, and 27 per cent other minerals of which the main one is biotite. These rocks increase slightly in grain size from east to west, with the primary textures toward the west being more completely destroyed, although local exceptions to this rule exist. Microscopically, the rocks are granoblastic to lepidoblastic and are quite equigranular; only mica and garnet are in porphyroblasts. The plagioclase (An_{20}-An_{35}) normally is not twinned, but it is altered to various degrees to sericite and saussurite, and its grains average (as does quartz) 0.15 mm in diameter. The biotite averages 0.5 mm, ranging between 0.1 and 2.0 mm. Biotite is in many places partly chloritized and, in shear zones, may be completely so. Among the minor constituents are almandine, retrograde chlorite, K-feldspar, and a little graphite. Accessory minerals include apatite, tourmaline, zircon, epidote, magnetite, and sulfides. The mica schists may be bedded, homogeneous, and mylonitic or cataclastic. In the bedded mica schist, dark, mica-rich layers alternate with lighter ones rich in quartz and plagioclase. The lighter layers contain calc-silicate intercalations from 0.5 to 2 cm thick. Primary graded bedding has been seen in this bedded schist. The lower portions of the bedded rock originally were quartz-plagioclase-rich graywackes, the higher ones largely mica-rich pelitic material.

The homogeneous mica schist contains no observable bedding, but the mineral composition differs considerably from one place to another. In some, it has so low a mica content as better to be designated as a mica-quartzite. In another type, however, mica is the predominant mineral. In most instances, it resembles fine-grained igneous rock.

In the several areas of mylonization and cataclastic shearing a coarse-grained flaky biotite schist has been formed; it is thought that silica originally present has been recrystallized in quartz veins and lenses in zones of minimum pressure.

The mica schist probably was originally deposited in a marine geosyncline, with the high-feldspar content indicating rapid sedimentation of poorly weathered material. The total thickness of this highly folded formation is thought to have been at least 3000 m.

In the black schists, the silicates are low in iron and have a lower content of alumina and silica than the mica schists with graphite and the sulfides providing the dark color of the rock; the sulfides are mainly pyrite and pyrrhotite. Microcline is common and the anorthite content of the plagioclase higher than in the mica schist, being between 25 and 70 per cent. It has been suggested by Peltola (1960) that the black schists are the metamorphic descendants of sapropelic sediments deposited in a marine environment.

Probably the most important characteristic of the Outokumpu ore area is the regular occurrence of the carbonate-bearing serpentinite - carbonate rock - tremolite skarn - diopside skarn - quartzite envelopes around the serpentinite bodies - this is the Outokumpu association. Beyond this envelope and between it and the mica schist, black schist commonly is present.

The elongate serpentinite lenses range from a few meters to several 100 meters thick; they are thought originally to have been ophiolites introduced into a geosynclinal environment. An analysis of their folding suggests to Gaál and his colleagues (1975) that they were tectonically emplaced in zones of pressure minima, particularly in the hinge zones of isoclinal folds. These men believe that the envelopes of altered rocks around the serpentinites were the result of regional and not of contact metamorphism. Gaál and his colleagues think that the serpentinites were injected tectonically after they had already been solidified and were not magma intrusions in their present settings. They do not say if the altered rocks that surround the serpentinites were developed before, during, or after their emplacement in their present positions. The serpentinites were divided by Haapala (1936) into western and eastern phases on the basis of the major mineral being chrysotile in the west and antigorite in the east. The two types are divided by a north-south line in the Horsmamaho area in the far-east portion of the Outokumpu area and well east of the ore bodies of Outokumpu and Vuonos.

In the western zone, the serpentinite, because the regional fold axis is subhorizontal, is elongate in mapped sections. The major lenses, in turn, are made up of many smaller lenses that are divided by minor sheared zones. Otherwise, the serpentinite is massive, and the component minerals show no dimensional orientation. The rocks from which the serpentinite was derived were: (1) dunite, monomineralic with very little residual olivine, (2) saxonite, now mainly composed of chrysotile with a wide variety of accessories; the original minerals were mainly olivine, plus various amounts of enstatite; the rock is now mottled; (3) porphyritic rock, probably originally dunite but with the olivine crystals being considerably elongated, and they, or their serpentinite pseudomorphs, give the porphyritic appearance to the rock. The three varieties are complexly interrelated, with the central parts of any lens being most altered, whereas what well preserved olivine there is is in the lens margins. The black color of most of the contact zones is due to finely divided magnetite; sulfides in the serpentinite are most abundant near the contacts with the alteration envelopes.

The eastern serpentinites have been more complexly folded and metamorphosed since they were strongly influenced by a third deformation that did not affect the western serpentinite. In this eastern body, the serpentinite lenses have heen further folded and now show a distinct foliation. The lenses are strongly deformed and are bordered by a mixture of talc-magnestite rock, tremolite skarn, zoisite-tremolite skarn, and the chlorite schist.

It would appear that the greater folding of the eastern section acccounts for its more intense alteration, but the lack of mineable sulfide bodies in this area can hardly be charged to the greater metamorphism of the eastern area. The only possible explanation for the lack of sulfides, if the sulfides were introduced with the parent magmas from the solid products from which the serpentinites were formed, is that they were driven off from the eastern serpentinites during the last, most vigorous stage of metamorphism.

The close spatial association of serpentinites, skarns, carbonate rocks and quartzites of the Outokumpu association is well established. The most recent opinion (Huhma and Huhma, 1970, and Gaál and his colleagues, 1975) is that the rocks also are genetically associated, with all of these rock types having been developed from the original ultramafic masses by autometamorphism. Earlier workers had other opinions such as considering the quartzite to have been a normal sedimentary rock in the black schist - mica schist series or that, the ultramafic masses having been extruded on the sea floor, the parent silica of the quartzites was deposited colloidally around it in the relationships now seen. How the colloidal silica was able to surround entirely the masses of ultramafic material on the sea floor completely is not clear; the part of the ultramafic mass on the sea floor surely was protected from being invaded by the colloidal silica. Against a primary sedimentary method of formation for the quartzite is the lack of any possibly detrital minerals in the quartzite. Thus, the most likely explanation for the presence of the quartzite around the serpentinite masses is that they, as well as the carbonate and skarn rocks, were produced by autometamorphism of the ultramafic material. Certainly, much silica is removed from the ultramafic material in its conversion to serpentinite.

The closest of the envelope of rocks to the serpentinite is the carbonate rocks; this material may be from a negligible amount to several meters thick; inward, the carbonate rocks grade gradually into the serpentinite. In places, carbonate-bearing talcose rock or pure talc rock may occupy the position of the carbonates; chlorite may be abundant in this talc. In the western (ore-bearing) area, the carbonate is dolomite, in the east it is magnesite; again this indicates that the degree to which the process that produced this group of rocks operated was greater in the east than in the west. Near the serpentinite, the carbonate may contain dots of serpentine; as the skarn is approached the carbonate rock includes increasing amounts of tremolite, and talc and chlorite may be locally quite abundant. Accessories in the skarn quite commonly include chromite and pyrrhotite and pentlandite.

The skarn separates the carbonate rock and the quartzite and ranges from a few centimeters to several meters in thickness; the thickest skarn occurs where quartzite tongues pinch into the serpentinite. Chromite-bearing (green) diopside is located against the contact with quartzite. This outer skarn rock is essentially monomineralic but may contain tremolite, and pyrrhotite is a common accessory. On the inner side of the skarn, between the diopside skarn and the carbonate rock (dolomite or dolomite-bearing serpentinite), is a tremolite (actinolite) skarn; it may contain some diopside but normally is essentailly tremolite; here pyrrhotite also is an accessory. Immediately against the carbonate rock, the skarn may be a tremolite-carbonate or talcose-tremolite rock that grades locally into chlorite schists with tremolite. The skarn may either be massive or foliated. The skarn contains a small quantity of several chromium-bearing minerals of which eskolaite was discovered at Outokumpu. Feldspar and biotite are conspicuous by their absence.

Finnish geologists now generally appear to be agreed (Haapala, 1936, Huhma and Huhma, 1970) that serpentinization was the process that caused the formation of the envelopes of rocks that surround that rock type. Thus, it seems that the serpentinization was an autometamorphic process in which the fluids made available during crystallization of the original ultramafic magma caused a radical change in the mineral composition of the ultramafic mass to produce, in what probably was essentially the same volume it initially occupied, the concentric arrangement of the rocks that have just been described. I would suppose that the process of autometamorphism was carried out in stages, so that the transference of material went in both directions until the various elements needed to produce the concentric layers had reached a stage of essential equilibrium.

It seems reasonable to assume that the ultramafic masses were intruded into the mica schist (and black schist) formation and were autometamorphically altered to the present arrangement of much serpentinite (perhaps 60 per cent of the altered ultramafic rock volume), quartzite (about 30 per cent) and dolomite and skarn (perhaps 10 per cent). These percentages estimates are rough but cannot be radically wrong. Certainly (Isokangas, 1978), the volume of quartzite and skarn is directly proportional to that of serpentinite, strongly suggesting that these rock types were all formed by the same process.

The Outokumpu ores are contained in the Karelian mica schist formation that is dated broadly, but certainly, as middle Precambrian. The minimum age of the ore is fixed by the ore being cut by dike-like body of Maarianvaara granite; this rock has given an age of about 1845 m.y. by the rubidium-strontium method. Actually dating the ore within the over 700 million years of middle Precambrian time before the Maarianvaara granite was intruded is less easily done. A considerable period of time can be subtracted from the lower end of this 700 million years because time must be allowed for the formation of the sediments of the Jatulian group that were laid down and lithified before Karelian sedimentation began. This should definitely be no less than 300 m.y., which would put the beginning of Karelian time at 2300 m.y. An age determination using an isochron plot of $^{207}\mathrm{Pb}/^{204}\mathrm{Pb}$ and $^{205}\mathrm{Pb}/^{204}\mathrm{Pb}$ gives a model age of 2300 m.y. This coincidence in ages from these two methods (one about as accurate as the other) does not mean much except that 300 m.y. for Jatulian time is not a radically inaccurate guess. Gaál and his colleagues accept this span between 2300 m.y. and 1845 m.y. (455 m.y.) as the time available for the four phases of deformation they recognize in the Outokumpu rocks and ores. Almost certainly, the ore was not emplaced until some considerable period after the end of Jatulian time, but the Maarianvaara granite was intruded in these authors' third phase, and the fourth (and final) phase must have been considerably after that date. As has been pointed, it is not completely certain that the Outokumpu ore was introduced during the first deformation phase, but it certainly was pre-second phase at the latest. This would give the Outokumpu ores an age of between 2200 and 2250 m.y., and this gives the ores a classification at middle Precambrian.

The Outokumpu ore is contained in two northeast-southwest-trending plates that overlap slightly about one-third of the way northeast from the southwest end of the ore body. The two plates probably at one time were a single mass but have been separated by a high-angle reverse fault. This fault (the Kaasila) strikes N25°W and dips 80°SW and has dropped the northeastern tip of the ore body downward by about 100 m. The ore northeast of the Kaasila fault is known as the Kaasila ore body. The Kaasila ore body overlaps the southwestern (Kumpu-Lietukka) ore body for about 225 m of the total length of both ore bodies of 4000 m. A second reverse fault, which strikes N25°E and dips about 75°SE, cuts the southwestern portion of the ore body into two parts, the larger of which is the northeastern fraction and is known as the Kumpu ore body; the smaller, southwestern fraction (appreciably smaller than the northeastern) is the Lietukka ore body. The displacement on this smaller fault is 5 to 6 m, with the northeast side being down. In addition to these three ore bodies, a small ore body lies above the Lietukka ore and is called the Jyrin Lietukka. Below the ore are two more satellitic ore bodies, the Poikanen and Turunen.

From the northeast portion of the Outokumpu ore body (the Kaasila plate), northeast of the Kaasila fault, to about the center of the Kumpu ore body, the axial pitch is from 15° to 25° southwest. From there on south, to the southwest end of the Lietukka ore body, the ore bodies are generally horizontal. The entire ore lens parallels the structure of the Outokumpu association and dips between 65°SE and horizontal through most of its length, although portions of the ore body may dip as much as 15°NW. The ore extends down the dip of the plates from 250 to 400 m, and the average thickness perpendicular to the dip is 7 to 9 m, although locally it may be as much as 40 m.

All the faults contain ore fragments, thus indicating that they are post-ore.

The Jyrin Lietukka ore body, although vertically above the Leitukka ore at its southwest end, has not been located certainly in the stratigraphy and may be in the same quartzite as the main ore body, folding accounting for the relative positions of the two.

The upper edge of the ore generally is in contact with quartzite, although rarely mica schist may be in this position but the lower edge of the ore penetrates into the serpentinite. Isokangas (1978) does not think the ore shows any contact-metamorphic effect on the host rocks, and the contacts of the ore with the wall rock are sharp, although along the contact with the serpentinite, that rock is sheared and chloritized. There are, however, tongues of ore that enter the wall rock at angles that are not concordant with the structure of the rocks. The longest of these tongues, is 500 m and is from 0.5 to 1.5 m thick; this tongue is in the footwall of the Kumpu ore body. Another such body of about the same size is in the footwall of the Lietukka ore body. How much of the character of these tongues (or satellitic ore bodies) is the result of initial deposition and how much to folding is uncertain; Isokangas (1978) and Gaál and others (1975) are convinced that the folding is completely responsible.

Three types of ore have been recognized at Outokumpu by the various geologists who have studied the deposits, with the designations of Gaál and his colleagues being used here; although those of Disler (1953) and Vähätalo (1953) are easily reconciled with those used here. These three types are: (1) layered (disseminated), (2) massive (structural normal), and (3) brecciated type (type with quartzitic inclusions).

According to Gaál and his colleagues (1975), in the layered type, the layers are distinguished by changes in compositon and are more or less regular and parallel to the foliation of the host quartzite. These authors recognize two subtypes within the layered type: (a) low-grade material in which the quartzite is predominant and (b) a high-grade ore in which sulfides are dominant over quartzite to form a compact ore with layers being distinguished by greater or lesser amounts of quartz in relation to sulfides. The low-grade layered type of Gaál and his colleagues appears to be equivalent to the disseminated type of Vähätalo, in fact, Gaál and his colleagues state that low-grade layered ore grades into an unoriented disseminated ore through to massive ore. Against this Vähätalo argues that the disseminated and normal (massive) ore types are separated by a zone of type (3) (brecciated ore). In the places where massive ore and disseminated ore border directly on each other, the contact is sharp. In one example, where a normal ore lies against a disseminated zone, the former cuts through the schistosity of the disseminated ore with sharp contacts. In still another area, dike-like bodies of massive ore cut perpendicularly through the preserved schistosity of the disseminated ore. These statements (those of Vähätalo and Gaál and his colleagues) are difficult to reconcile. If Gaál and his colleagues are correct, then ore emplacement took place in one continuous process, with the density of ore in a given rock volume depending on the duration over which the process acted. Contrarily, if Vähätalo is right, then very real hiatus occurred between the formation of the disseminated ore and the introduction of the normal (massive variety). This problem will be considered further after the character of the ore is examined.

The Outokumpu ore contains a considerable variety of sulfides that average 3.5 to 4.0 per cent copper, about 28 per cent iron, slightly less than 25 per cent sulfur, and about 1 per cent zinc. The content of nickel is between 0.02 to 0.50 per cent and of cobalt from 0.20 to 0.40 per cent. The ore contains small amounts of gold, silver, and selenium. Before any mining had been done, the total tonnage of ore in the deposit was about 28 million.

Although the ore is predominantly in quartzite, microscopic examination shows that the skarn minerals in the quartzite (mainly diopside and tremolite) were attacked before quartz grains by the replacing sulfides, that in the zones in which partly replaced fragments of skarn minerals are concentrated are rich ore, and that quartzite bands low in sulfides also are low in skarn minerals, replaced or not.

The massive (normal) ore is concentrated in the central parts of the ore plate and is nearer to the hanging- than the footwall. In some areas, although the massive ore types usually are separated from the wall rocks by brecciated and then disseminated ore, massive ore may lie directly against the wall rock, again suggesting, if Vähätalo is correct, that the massive ore does not grade through lower and lower grade material to die out against essentially unmineralized wall rock. This seems to confirm his position that ore types (1) and (2) were formed in different ways the one from the other.

In addition to grade differences, two mineralogically distinct types of ore occur at Outokumpu; these are pyrite-dominant and pyrrhotite-dominant. The pyrite ore forms the central part of the ore lens and normally is surrounded by the pyrrhotite type. The thickness of the pyrrhotite envelope ranges from several meters to zero; further, at the ends of the ore body, the ore is entirely of the pyrrhotite type. If the ore was formed in skarn, the pyrrhotite type is dominant.

In some brecciated ore, Gaál and his colleagues believe that remobilization of pyrrhotite into spaces around the breccia fragments may raise the ore to the pyrrhotite-dominant class.

In the pyrite-predominant ore, the pyrite is present as euhedral grains, the size of which ranges between 0.01 and several millimeters. This mineral may be as much as 40 per cent of the massive ore. Although pyrite normally is missing from the pyrrhotite-dominant ore, it is found as separate porphyroblasts up to 30 cm in diameter. These pyrite crystals usually are fractured, and the fractures are cut by micro-sized veins of chalcopyrite and pyrrhotite and, less commonly, of sphalerite and quartz. The pyrite cubes may be bordered by rims of sphalerite or chalcopyrite. The cobalt content of the pyrite ranges from a few tenths of a per cent to several per cent.

Pyrrhotite may constitute 10 to 20 per cent of the pyrite-dominant ore in grains from minute to 1 or 2 mm. In the pyrrhotite-dominant ore, that mineral may exceed 50 per cent of the ore volume. Chalcopyrite is closely associated with pyrrhotite in both types of ore, and sphalerite normally is found with the chalcopyrite and in a few places as exsolution stars in the copper mineral. On the southeastern edge of the ore body, a few areas in the pyrrhotite-dominant ore contain abundant cubanite, but it is rare in the pyrite-dominant variety. Cobalt pentlandite is essentially always present with pyrrhotite; it occurs both as grains surrounding the pyrrhotite and as exsolution lamellae in it. Mackinawite [$(Fe,Ni)_9S_8$] is common in the cubanite occurrences and there replaces pantlandite. The silver in the ore is contained in argentian pentlandite that is thought to be the main silver-bearing mineral at Outokumpu. Magnetite is a minor constituent of both pyrite- and pyrrhotite-dominant ores. Magnetite can be found in all three ore types. Vähätalo distinguishes two types of ore, based on mineral content; these are the normal paragenetic type and the cubanite type, with the latter being usually associated only with the massive (normal) type. The cubanite ore rarely is found with the other two structural ore types.

Two major problems remain to be solved concerning the Outokumpu ore: (1) how was the ore originally emplaced and (2) how much have the relations among sulfide minerals and of sulfide minerals with gangue been affected by the various tectonic episodes that followed the introduction of the ore. In considering the first problem, it is necessary to emphasize that Gaál and his colleagues believe that the primary ultramafic material was emplaced before the first deformational phase and that the production, probably by autometamorphism, of the serpentinite and its surrounding envelopes of carbonate rock, skarn, and quartzite occurred as one continuous process. The major fact favoring the development of the quartzite during this stage is the fact that the amount of quartzite in any given area is proportional to the amount of serpentinite in that area. Further, Peltola (at the 11th Nordic/Winter Geological Meeting at Oulu in 1974) says that presence of bent and broken fragments of sulfides in the breccia ores strongly suggests that the ore were emplaced before the breccia ore were formed. This concept is subject to some question because it is difficult to imagine a deformative process that always would break the ore material along the boundary between massive and disseminated ores. It seems at least equally probable that the breccia ore was emplaced by the replacement of fragments already broken before the sulfides were introduced. This explanation would require that the ore fluids entered along the zones of brecciation with the breccia being replaced in large part by sulfides but with more massive ore being formed inward from the broken zones toward the center of the ore plate and the disseminated ore formed outward (toward the margins of the ore plate) from the brecciated zones. Thus, unlayered disseminated ore and the low-grade layered ore into which it usually grades were produced outside the breccia zones and the high-grade layered ore and the massive ore into which it grades were emplaced inward from the breccia zones.

Even if the ore was introduced after the first deformative phase that produced the breccia zones that later were converted into breccia ore, the additional three phases of folding found by Gaál and his colleagues undoubtedly had a very appreciable effect on the already emplaced ore, probably producing many of the intermineral relationships that, in the early days of geological study of the Outokumpu ores, were ascribed to replacement by hydrothermal fluids. For example, the chalcopyrite, pyrrhotite, sphalerite, and quartz that occur in veinlets in fractured pyrite cubes may have flowed into these fractures under the influence of tectonic stresses and that the pyrite cubes themselves may have been developed as porphyroblasts rather than by metasomatic processes.

But, granting the importance of metamorphic effects on the intermineral relationships of the ores, this does not explain how the ores originally were introduced. The suggestion has been made that the original ores: (1) were deposited on the sea floor at much the same time as the ophiolite parent magma of the serpentinite and their surrounding envelopes, (2) were hydrothermally introduced after the magma and the sediments containing it had been lithified and, in the case of the ultramafic rock, autometamorphosed to the serpentinite and its surrounding rocks of the Outokumpu association, (3) were emplaced, so far as the disseminated ore (and low-grade layered ore) were concerned by hydrothermal replacement and that the massive ore was introduced as molten material rather than through the agency of hydrothermal fluids, and (4) were emplaced through the lateral secretion of sulfide material from the surrounding mica schist - black schist sedimentary sequence.

To complicate the problem further, Mikkola and Väisänen (1972) consider that the ores at Outokumpu (and at Vihanti) owe most of their relationships to the enclosing rocks to remobilization. They believe that the banded structures that are common in the Outokumpu ores are remnants of original sedimentary banding. These bands were distrubed by metamorphically induced plastic flow of sulfides and quartz, which were forced largely into crests and troughs of folds and into cross fractures along the fold limbs and into the pressure shadows created by boudinage and breccia structures. These authors also think that both the compact sulfide material in veins in the ore bodies or in the closely associated wall rocks were emplaced by such pressures. They also envision veins having been formed by what they call "metahydrothermal" solutions, solutions produced by pressure-induced (?) processes. They believe that the present structures of the ores and the epigenetic features (at least at first glance) of the ores all are the result of remobilization. They do not, however, think that remobilization changed the chemical composition of the ore minerals much but only increased the irregularity of their distribution.

That most, if not all, of the ore textures discussed in the preceeding paragraph could have been produced by replacement after the rocks of the Outokumpu association were developed does not simplify the problem of explaining the manner in which the ores actually were formed.

The high concentration of cobalt in the Outokumpu copper-cobalt ore is a most unusual circumstance. If the ore were nickel- and not cobalt-rich, it would be easy to agree with the concept that the massive ores were introduced in the molten state. So far as I know, however, no water-poor melt of magmatic origin that has formed a workable ore body (or any other mineralized deposit) has been low in nickel and high in cobalt. On the contrary, cobalt-copper deposits, such as those of the Zambian copper belt were formed either by hydrothermal fluids invading lithified rock or by solutions of the same type pouring out on the sea floor to deposit the copper-cobalt ores. On balance, then, it seems most reasonable to consider that the Outokumpu ores were introduced by hydrothermal fluids that either invaded the rocks of the Outokumpu association or deposited their load on the sea floor in close spatial relationships with the ultramafic bodies from which the serpentinites and the rocks associated with them were formed. Before a decision can be made as to which form of hydrothermal activity was responsible for the Outokumpu ores, it is necessary to consider the second question posed a few paragraphs previously - namely, how much have the relations among sulfide minerals and of sulfide minerals with gangue been affected by the various tactonic episodes that followed the introduction of the ore.

Although evidence exists that some of the ore minerals have been remobilized - the sulfides in cracks in pyrite, for example, none has been reported that indicates that the mineral species originally present have been changed by tectonic activity. From this it would seem to follow that change in position of certain rocks relative to others and of certain minerals relative to others have resulted from tectonism but neither the rocks nor the sulfide minerals have been so changed. Thus, the minerals now present can be considered to have been those originally formed in the Outokumpu association. These minerals include several - pyrrhotite, cubanite, cobalt, pentlandite, and the more sparse magnetite, plus exsolution textures in the pyrrhotite and the chalcopyrite - that indicates that the ore minerals were formed at high temperatures, temperatures higher than could have existed on the sea floor as hydrothermal fluids were extruded out on it. Thus, it seems more probable that the Outokumpu ores were formed by the replacement of the rocks of the Outokumpu association at temperatures that included the hypothermal range but that also were reduced during ore formation into the mesothermal. For these reasons, therefore, the Outokumpu ores are classified here as Hypothermal-1 to Mesothermal, the designation of Hypothermal-1 being used because the ores were mainly, if not almost entirely, emplaced in quartzite and not in carbonate rock.

The copper-cobalt ore at Vuonos is much the same as that at Outokumpu, but, in addition, enough low-grade nickel ore is present (separately from the copper-cobalt ore) to add an interesting complication to explain. The copper-cobalt ore is quite similar in size and shape to that at Outokumpu, as well as in geologic setting. The ore body itself is about 3500 m long, 50 to 200 m wide, and 50 to 60 m in average thickness; locally it is up to 20 m thick. The ore fades out gradually at both ends, with the southwestern end being 60 m below the surface and the northeastern about 200 m down; of course the ore body did not outcrop.

As is true at Outokumpu, the host rock is predominantly quartzite, with mica and black schists being the country rocks at the upper edge of the ore body and skarns, serpentinites, and talc schists are at the lower edge. The ore is banded, with the bands being caused by sulfide-rich layers alternating with ones of gangue; these bands appear to conform to the contacts that usually are quite sharp. Isokangas (1978) says that the country rock and the ore body enclosed in them have been folded, recrystallized, and internally mobilized. Along the lower edge of the ore body, one fold has turned into a fault; in the continuation of this fault, immediately beyond the main ore body, Isokangas reports that there is a heterogeneous dissemination of remobilized sulfides in which the sphalerite is unusually abundant. The ore area also contains a small satellite ore body that he assumes is a downfolded continuation of the upper edge of the ore body.

The three main sulfide minerals are pyrrhotite ($Fe_{15}S_{17}$), about 40 per cent of the ore mass, chalcopyrite about 7.3 per cent, and sphalerite about 3.2 per cent; copper grade is about half that at Outokumpu, but the zinc grade is about twice as high. The main gangue mineral in the ore is quartz (unreplaced quartzite most probably), with some of the skarn minerals in minor amounts, mainly tremolite and diopside. The most important accessory sulfide is cobalt pentlandite, with some cobalt also in the pyrite; Isokangas thinks this pyrite is "secondary." Minor sulfides include mackinawite, stannite, cubanite, marcasite, cobaltite-gersdorffite, siegenite, violarite, molybdenite, and argentian pentlandite. The principal, though minor, oxides or oxysalts are zincian chromite, magnetite, eskolaite, cassiterite, and uraninite. Small quantities of graphite (with uranium), bismuth, and gold are present.

The rocks of the Outokumpu association at Vuonos (not, of course, including the mica and black schist) contain an average of 0.10 to 0.20 per cent of nickel. One volume of this rock, in the hanging wall of the copper ore, averages 0.20 per cent nickel, 0.04 per cent copper, and 0.02 per cent cobalt. The better nickel ore (0.2 to 0.4 per cent) is in quartzite and skarn rocks and averages 0.35 per cent nickel in the best ore of that type. In the serpentinites, the nickel content ranges between 0.16 and 0.21 per cent; the dolomite has less than 0.15 per cent nickel.

The principal minerals in the nickel ore body are pyrrhotite and pentlandite (nickel- and not cobalt-bearing). Where the ore is rich in mica, pyrite locally is abundant. The accessories are essentially the same as those in the copper-cobalt ore; even cobaltite-gersdorffite is one of them. The chief gangue mineral is quartz (from quartzite), with much less diopside and tremolite. In the breccias in this ore type, chlorite, biotite, fuchsite, and talc fill fractures. Isokangas considers part of the pyrite in the nickel ore to be primary; it is a cobalt-poor type.

Although it is easy to assume that the copper-cobalt ore at Vuonos was formed in the same manner as that at Outokumpu, its relationship to the nickel ore is not clear from what Isokangas says. It would appear that the nickel ore does not overlap with the copper-cobalt ore, although this is not certain. Further, the nickel ore is, except for its essential lack of copper and cobalt minerals, much the same mineralogically as the copper-cobalt ore. Nothing is said by Isokangas that indicates if the copper-cobalt ore was introduced before or after the nickel ore or at the same time. This raises the question, that was not answered for Outokumpu, - from where did the ore fluids come that introduced both ore types? They almost certainly did not come from the Maarianvaara granite, so they must either have come from the serpentinite or from some other, probably less mafic source, no trace of which has been found as an igneous rock in the Outokumpu-Vuonos area. It seems improbable that the fluids came from the serpentinite; the volume of serpentinite, large though it is in comparison to the other rocks of the Outokumpu association, is far from large enough to have supplied a fluid sufficient in volume and varied enough in compositon to have produced both the rocks of the Outokumpu association and the ores as well. No other area of serpentinite in the world (to my knowledge) has deposits of the Outokumpu-type associated with it. Further, the production of ore fluids from serpentinite to form both the copper-cobalt ore and the nickel ore at Vuonos and the failure of the serpentinite at Outokumpu to do the same is a problem to explain. If the serpentinite is eliminated as a source of ore fluid for the ores at Outokumpu and Vuonos (Luikonlahti, as will be seen later, is similar to Outokumpu in lacking a nickel deposit), then appeal must be made to a hydrothermal fluid from an unknown source. This is true whether the fluid emplaced the ore minerals by replacement or by pouring out on the sea floor. It has been suggested that the cobalt-rich solutions were produced from a magma less mafic than the parent of the serpentinite and that the ore-forming fluid did not separate until quite late in the crystallization cycle, late enough that most of the nickel, in preference to cobalt, had been taken up by crystallizing silicate minerals. This would have resulted in an ore fluid rich in cobalt since the cobalt is less readily accepted in silicates forming at high temperatures than is nickel.

All this far from satisfactory, for, if the explanation given is correct, Outokumpu is the only pentlandite-bearing deposit (even if the pentlandite is cobalt rather than nickel pentlandite) I know of in which the ore-forming material was a water-rich fluid and not a metal-rich, water-poor melt. The problem of the origin of the ores of the Outokumpu-type, therefore, is far from solved. It would help to know the relative ages of the nickel and cobalt ores at Vuonos, but even if this were known, much would remain to be solved.

At Luikonlahti, the rocks and ores are analogous to those at Outokumpu. The Luikonlahti complex is 1.8 km long and 1.0 km wide and is embedded in a mica gneiss (not schist) that contains black schist; the complex trends N60°E and dips 60° to 75°NW. It has been considerably folded, faulted, and fractured. It also is cut by granite and pegmatite veins that appear to be related to the Maarianvaara granite. The mica gneiss, despite the different name, is similar to the mica schist-black schist complex at Outokumpu, with scales or aggregates of graphite giving the rocks a dark color. The black schist contains pyrite, pyrrhotite, chalcopyrite, and sphalerite in amounts normal at Outokumpu. The serpentinites form elongate lenses or irregular bodies that include, in addition to serpentine minerals, anthophyllite, actinolite, talc, relict olivine, chlorite, and carbonates; the rock is dark due to magnetite dust. Also present are minor chromite, pyrrhotite, and less pentlandite and chalcopyrite. The quartzites are adjacent to the serpentinites, and are coarse-grained and as much as several meters thick. No evidence that the quartz was the result of weathering has been seen. Carbonate rocks and tremolite and diopside skarns are intercalated in quartzite and appear on the contacts of the quartzite with serpentinite. As at Outokumpu and Vuonos, the skarns contain such chromium-bearing minerals as uvarovite, and chrome-bearing diopside, tremolite, and mica (fuchsite). Heterogeneously disseminated in the quartzite are pyrrhotite, plus rare chalcopyrite and sphalerite; pentlandite is exsolved in the pyrrhotite.

The Luikonlahti deposit is made up as several irregular lenses that have sharp contacts with the wall rock. Most of the ore is massive and two types are distinguished, those with quartzite host rocks and those with skarn. Locally, the ore is brecciated. As at Vuonos, the major mineral is pyrrhotite, with only occasional idiomorphic grains or grain aggregates of pyrite at the edges of the ore bodies. The other ore minerals are chalcopyrite, sphalerite, and cobalt pentlandite, with the latter being quite minor. Accessory minerals are mackinawite, cubanite, stannite, cobaltite, molybdenite, and magnetite, with some chromite and ilmenite. The cobalt content ranges between 0.15 and 0.10 per cent, and nickel is even less abundant. No suggestion has been made that the Luikonlahti ores formed in any way differently from those at Outokumpu or Vuonos.

A final judgement on the manner of formation of the ores of the Outokumpu region is difficult to make. The considerable presence of nickel and cobalt (uncommon in hydrothermal or volcanogenic copper-zinc deposits) argues against the ores having been deposited from such a fluid. The proportions of nickel to cobalt in the Outokumpu-type ores and the quantities of nickel as well, however, are so low as to suggest that the ores were not introduced through the medium of a water-poor melt. This leaves the alternative of a hydrothermal fluid pouring out on the sea floor during primary sedimentation or entering the solid rock after at least the first stage of metamorphism. Neither of these last two explanations is fully satisfactory, and the syngenetic one even less than the solid-rock replacement hypothesis. However, if such a choice is necessary here, I consider the ores as Hydrothermal-1 to Mesothermal. The ore minerals do not appear to have been changed radically as to species or quantities by metamorphism subsequent to their emplacement, but considerable change appears to have been effected in form and location by that process, so it seems advisable to add Metamorphic-C to the classification of deposits of the Outokumpu type.

Mäakelä (1974) studied 258 sulfur-isotope analyses from Outokumpu. In these, the range of delta-^{34}S results is from $+5.8^{\circ}/_{oo}$ to $-19.2^{\circ}/_{oo}$, and the arithmetic mean is $-3.5^{\circ}/_{oo}$. On the basis of these results and the structure of the ores, Mäkelä favors a volcanic-sedimentary mechanism for the formation of the Outokumpu ores. He does, however, believe that the ore-fluids came from the same magma that produced the Karelian volcanic rocks and that the bulk of sulfur came in with those fluids even though a small fraction of the sulfur might have come from sea-water sulfate though how this might have been accomplished, he does not say.

Recent work on the Vammala ores (cf. Häkli, and others, 1979, ref. under Ahlainen-Sääksmäki Nickel Belt), however, indicates that the sulfides in an ultramafic flow extruded on the sea floor are completely oxidized and removed from the flow rock. In the case of Vammala, a small fraction of these sulfides were later substituted for by a much smaller amount of secondary sulfides. Similar oxidation effects have been reported (Lalou, 26th IGC, 1980) on sulfide deposits now forming on the sea floor so that the possibility that sulfides so deposited can be incorporated as such in the sea-floor stratigraphic sequence seems remote. Thus, the possibility that the sulfides of the Outokumpu deposit were laid down in the sea floor in conjunction with the (now altered) ultramafics with which they are now stratigraphically associated is not good.

Mäkelä (1974) says that work with the delta-values of 59 pyrite-pyrrhotite pairs and 10 sphalerite-pyrite pairs convinces him that the equilibrium temperature for these minerals is about 350°C. He points out, however, that 350°C can be accepted only as a statistical value indicating the temperature at which isotopic equilibrium was reached. It seems doubtful that such temperatures could have obtained on the sea floor during even the outpouring of the hydrothermal fluids that Mäkelä assumes were the parents of the Outokumpu sulfides. Such temperatures can have been achieved only during subsequent metamorphism and, therefore, do not represent the temperature at which the sulfides were introduced into the Outokumpu system. Centainly, the sulfides were not developed by the autometamorphism that produced the serpentinite and other alteration products of the ultramafics that surround it. This means that the ore fluids must have come from some other source than that that supplied the ultramafic magma.

Age determinations for the lead in the Outokumpu galenas have suggested that the original ore-forming fluid arrived in the area about 2250 m.y. ago. The only igneous rocks that Mäkelä mentions as having about the same early Karelian age (2250 m.y.) are diabases in the radiometric study of which the U, Th-Pb method was used. This information indicates that the Outokumpu ore-forming fluids well may have come from the same magma chamber as the diabases and deposited their load of ore minerals either on the sea floor or by reactions with the rocks of the Outokumpu association. Such ore fluid-solid rock reactions could have taken place, of course, only if the ore fluids had reached the Outokumpu area after the ultramafic parents of the serpentinites had been solidified and autometamorphosed . These events may have occurred after the first of the four stages of metamorphism to affect the rocks of the Outokumpu association.

Mäkelä is convinced that all, or most,of the apparently epigenetic features shown by the Outokumpu ores probably were produced by a redistribution of the primary sulfides by one of the three mechanisms suggested by Mikkola and Väisänen (1972), that is: (1) plastic flow, (2) solid diffusion, and (3) metahydrothermal solution and redeposition. Mäkelä believes that plastic flow of the sulfides was the most probable method for redistribution of sulfides in the initial stages of metamorphism. He, however, considers that solid-state diffusion was the main mechanism for the transportation of the minerals now filling tension joints. He thinks that the mechanism for producing the sulfide veins that normally contain idiomorphic quartz cannot have been plastic flow and probably not solid-state diffusion; instead it must have been the activity of metahydrothermal solutions.

Mäkelä points out that both plastic flow and solid-state diffusion move in the same way so far as isotope fractionation is concerned. Actually, diffusion is the major factor in both processes of kinetic separation with the molecules of the light-sulfur isotope moving more rapidly than those of the heavy one. Further, in plastic flow, the recrystallization that is associated with plastic flow aids the movement of the lighter isotope. Also the free energy increase in material undergoing plastic deformation, preliminary to plastic flow, also favors the greater movement of the lighter isotope, insuring that ^{34}S is additionally depleted in the redistributed phase. [This follows because the free energies of isotopic molecules are inversely proportional to the square roots of their masses (as Rankama demonstrated in 1963), thus slightly favoring the movement of the ^{32}S out of its initial position and into another at a distance removed.]

When quartz is involved in what Mäkelä considers to be redistributed material, appeal must be made to metahydrothermal solutions; dependence cannot be placed entirely on relative movement of sulfur isotopes as is apparently achieved in plastic flow or solid-state diffusion.

Mäkelä lists the sequence for remobilized sulfides as (in order of least ease of mobilization): pyrite > sphalerite ≥ chalcopyrite ≈ pyrrhotite, that is, the content of ^{34}S is greatest in pyrite down to least in pyrrhotite, although in a few places, pyrrhotite and chalcopyrite reverse positions.

Mäkelä's study shows that much of the movement of the Outokumpu sulfides under the influence of the various states of metamorphism well may have been due to the three forces listed by Mikkola and Väisänen, but it does not tell how the sulfides initially entered the Outokumpu association. The sulfur-isotope ratios in the ore sulfides that show no evidence of redistribution certainly are close enough to the meteoritic standard that they probably came from a magmatic source. But this does not tell how they originally became associated with the serpentinite - dolomite-skarn - quartz of the Outokumpu association. They either poured out on the sea floor to interleave with the ultramafic rocks or they were introduced somewhat later, after the ultramafic lavas had crystallized as hydrothermal fluids that entered fractures generated by, probably, the first stage of metamorphism in the region. If the ore fluids poured out on the sea floor, the classification of the deposit must by deposits concentrated in quiet waters by volcanics exhalations, that is - Sedimentary-A3, plus Metamorphic-C. If, the ore fluids entered the system after the ultramafics had solidified and had been auto-metamorphosed, they would be categorized as Hypothermal-1 to Mesothermal, plus Metamorphic-C. I prefer the latter because, had the ore fluids been poured out on the sea floor, they would have been sufficiently oxidized thereby that the effects of this process still would be recognizable even after the passage of 2.2 x 10^{-9} years.

PYHÄSALMI

Middle Precambrian	Zinc, Copper Pyrites	Hypothermal-1 Metamorphic-C (?)

Helovuori, O., 1979, Geology of the Pyhäsalmi ore deposit, Finland: Econ. Geol., v. 74, p. 1084-1101

Huhtala, T., 1979, The geology and zinc-copper deposits of the Pyhäsalmi-Pielavesi district, Finland: Econ. Geol., v. 74, p. 1069-1083

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Kurppa, R. O., and others, 1964, Outokumpu Oy., Pyhäsalmen kaivos, Vuorit. Bergs., p. 18-35

Mikkola, A. K., 1969, Aspects of wallrock alteration associated with some Finnish sulphide deposits: a review: Inst. Min. and Met. Tr., v. 78, Sec. B (Bull. no. 750), p. B65-B71; disc. 1970, v. 79, p. B97-B101; particularly v. 78, p. B69-B71

Notes

The Pyhäsalmi ore body is in the commune of Pyhäjärvi, in Oulu County; the coordinates of the mine are 63°39'33"N and 25°02'55"E. The deposit did not outcrop, being covered completely by a few meters of till. It was discovered by a farmer who tried to sink a well on his property and, instead, struck massive pyrite ore that contained considerable sphalerite and chalcopyrite. No float that could be traced to the Pyhäsalmi ore body has yet been found, although float sent in by local residents induced the beginning of exploration at Säviä and Vuohtajoki.

Although the Pyhäsalmi zinc content is less than half of that in Vihanti ore and the copper grade only about 1.5 times that of Vihanti, the total annual tonnage produced from the Pyhäsalmi open pit and mine over the years 1970 through 1974 was 4.0 million tons compared with 2.5 million from Vihanti over the same time period. At the beginning of 1975, the reserves at Vihanti and Pyhäsalmi, respectively, were about the same, 21 and 22 million tons. The present production (1979) is about 1.1 million tons annually; this includes 200,000 tons of oxidized ore mined earlier and stored for future use. The concentrator was completed in 1962 and reached 600,000 tons by 1967 by which time, the concentrator had been enlarged and automated. Mining began in the open pit before the original concentrator was completed; mining both on the surface and underground continued until 1976, when all ore recovered came from below the surface. The Pyhäsalmi ore is estimated (Helovuori, 1979) to average 0.85 per cent copper; 2.8 per cent zinc; 37 per cent sulfur; 33 per cent iron; 0.2 gm/ton gold; and 14 gm/ton silver, plus 4.9 per cent BaO; 0.06 per cent lead; and 15 ppm selenium.

The Pyhäsalmi mine is about 100 kilometers south-southeast of Vihanti. The Pyhäsalmi deposit (as is true of those at Vihanti and of other, smaller, compact, strata-bound and unexploited deposits in the Pyhäsalmi-Pielavesi district, such as Vuohtojaki, Hallaperä, and Säviä) is in a Precambrian mobile zone that borders on an old craton in (west) central Finland (Huhtala, and others, 1979). The craton area, close to which the Pyhäsalmi ores occur, is designated as a "Presvecokarelidic Complex" that is 2600 to 2800 million years old, and the rocks of the craton are reported to be granitoids that differ in compositon and structure and include banded migmatic granites and granodiorites plus homogeneous granites. Other rocks present in the basal complex are mica and hornblende gneiss of sedimentary origin. All of the rocks of this complex are cut by gabbros, diorites, and granites.

The boundary zone that separates this ancient craton from the Svecofennian rocks that contain the Pyhäsalmi ores is made up of large blocks of cratonal rocks divided by fracture zones, the schistosity of which has the same strike as the fractures. The ore-bearing Svecofennian formations are located in the wide-spread Savo schist area in that portion of the Savo known as the Ruotanen schist zone. The Ruotanen zone is about two kiometers thick in the ore deposit area and trends about north-south. The dips are between steep eastwardly to vertical, and the fold axes and lineation trend 40°-60°S. To the east, the schist zone abuts against a granodiorite mass that is a highly altered dome of the Presvecokarelian gneiss. To the west, the schist zone is cut by an obviously younger porphyritic granite (Isokangas, 1978). The location of the ore body in the Ruotanen schist zone is controlled by a sericitic quartzite with which leptite (quartz-feldspar schist) is closely associated. Other prominent rocks in the Pyhäsalmi ore zone are amphibolites and mica gneisses. The leptites are fine-grained, slightly foliated rocks in which the feldspar is oligoclase and the mica content is low; quartz phenocrysts are characteristic. Locally, a vaguely developed banding is seen, but definite layering is not. A few amphibolites layers occur in the sequence.

The ore-host, the quartz-sericite schist (or sericitic quartzite) is more intensely foliated than the leptites, but a portion of this rock contains enough sericite to be called sericite schist. In these rocks, sparse amounts of pyrite are disseminated; this pyrite increases in abundance as the contact with massive ore is approached. Pyrrhotite, chalcopyrite, sphalerite, and, in a few places, tourmaline are other accessory minerals.

Certain mica-rich rocks, with a composition near to that of the leptites, are known to be in contact with the leptites. Associated with these mica-rich rocks are schists that contain cordierite or cordierite and anthophyllite. Zones of schists several meters in thickness occur in the leptites; the leptites also may include pinitized porphyroblasts of cordierite. These cordierite rocks normally are in the hanging- or footwalls of the ore body.

Although the mica-rich schists control the location of the ore, the most abundant rock in the Ruotanen schist zone is amphibolite; it is mainly in the footwall west of the ore deposit. These amphibolites include intercalations rich in diopside or quartz or both. On the most westerly edge of the schist zone are dolomite and diopside skarn and graphite beds, interlayered with the schist and in many places impregnated with iron sulfides (Isokangas, 1978).

A microcline pegmatite, at the south end of the ore contact, is younger than the schist formation.

This metasedimentary sequence was intruded by plutonic rocks that range in composition from ultramafics to granites; their mutual age relations have not yet been determined. With the ultramafic rocks are associated such deposits of the nickel belt as Hitura and Makola (see Hitura-Makola - Kotalahti, this volume) that lie in a far different trend (nortwest-southeast). These ultramafics have been given dates from 1950 to 1900 million years old. The silicic rocks of the Vihanti-Pyhäsalmi zone range from 1900 million years to 1750 million years in age. This age-range also is that of the regional metamorphism and of the main phases of deformation (Isokangas, 1978).

Isokangas (1978) is of the opinion that the sulfide deposits of Pyhäsalmi (and Vihanti as well) are located in the lower portion of the geosynclinal sediments, that is, in the epicontinental metasediments or in the rocks stratigraphically immediately above them. The country rocks of the Vihanti-Pyhäsalmi zone differ markedly from one mine to the other; at Pyhäsalmi, they are pelitic schists, amphibolites, calc-silicate rocks, and cordierite-anthophyllite rocks. The ore bodies in many places follow the boundaries between the geologic formations, although, where the ores are predominantly pyritic, small-scale discordant structures are present.

Most of Pyhäsalmi ore is contained in a single massive lens with, at deeper levels, a small, separate lens lying east of, and parallel to, the southern tip of the main body. The main deposit is located in a shear zone along the axial plane of a fold on the western flank of a syncline, but the parallel lens is on the eastern flank of the syncline. These structures parallel the local north-south strike of the schists (Mikkola, 1969). The ore body is about 650 meters long (north to south), has a maximum width of 80 meters, and tapers gradually toward both ends; it is known to extend to a depth of 600 meters. Starting at its northern end, the ore body dips flatly to gently southward but gradually steepens; at its southern tip, its dip is nearly vertical. The horizontal shape of the ore body is that of a stretched-out "S".

The country rock of the ore body (the sericite quartzite) has sharp conformable contacts with the ore. In most instances, the ore is a massive, coarse-grained material that is rich in pyrite and contains various amounts of sphalerite and chalcopyrite. The gangue minerals, about 25 per cent of the rock, are mainly quartz and micas, although locally cordierite and tourmaline are present. Where the ore is zinc-rich, barite and calcite occur. Randomly distributed accessory sulfides include pyrrhotite, arsenopyrite, and tetrahedrite, plus magnetite. The structure of the ore is due mainly to the accumulation of sphalerite in bands and layers. Some of the ore is brecciated, with fragments that range in size from a few cubic centimeters to thousands of cubic meters. Fragments made up of sericite quartzite or sericite schist are rare in the breccia; most of such fragments are leptites or plagioclase- or quartz-porphyrites. Near the footwall contact, fragments of amphibolites and carbonate rocks are present; in such places, the zinc content of the ore is higher than is normal. Ore in this situation also is unusual in that chalcopyrite, galena, arsenopyrite, and native gold and silver are more than normally abundant. Beyond the ore, in the sericite-quartzite country rock, these minerals occur in disseminated form; further, veins of chalcopyrite, containing idiomorphic tourmaline, scapolite, diopside, and tremolite are present in the sericite quartzite near the ore. In the southern part of the ore body, and along the eastern and southern contacts of the minor ore mass associated with pegmatite, pyrrhotite-rich ore is met. The zinc content is highest in the center of the ore body and becomes less toward either end. In the contact zone, some ore contains round and fractured phenocrysts of pyrite (Isokangas, 1978). The ore in the eastern parallel body is a pyrrhotite-rich one.

Some near-surface alteration has affected the ore with chalcopyrite partly converted to chalcocite, covellite, and bornite; pyrrhotite in part has been converted to melnikovite and marcasite; these minerals make flotations of the ore appreciably more difficult.

The ore in the eastern parallel body is pyrrhotite-rich, and the shape of the body is most complex. Helovuori (1979) thinks that this parallel ore body is in its present shape through polyphase deformation, plus local remobilization of sulfides. This intricacy of form is further promoted by the late intrusion of the coarse microcline pegmatite that cuts the ore zone at its southern end.

Normally, the contacts of the host rock and the massive ore are sharp, but, in some places, as the mineralization is followed outward from the massive bodies, pyrite gradually decreases in the sericite schist, but any veins or veinlets have sharp contacts with the sericite schist. The pyrite in the schist, moreover, is not in sufficient quantity to blur the contact between the main mass of disseminated and massive sulfides.

In part, the varied- sized fragments (cubic centimenters to several thousands of cubic meters) are in random orientation, but others are concentrated in a definite zone parallel to the longitudinal axis of the ore body. Helovuori (1979) recognized boudinage structures in certain layers. Most of the fragments in the massive ore are silicic volcanics, some of which showed clear quartz-porphyry structures. Some fragments are composed of amphibolites and plagioclase porphyries. Near the footwall contact fragments of limestone, somewhat or considerably rounded, are contained in a zone in which the zinc content of the ore is higher than average. Thus, it appears, even in rock volumes in which carbonate sediments or skarns are present in modest amounts, that the original ore fluids were much more active in their attack on the carbonates and their derivatives than in the schistose beds. Essentially none of the fragments are sericite quartzites or sericite schists.

Helovuori (1979) recognizes five classes of ore. The first of these is normal or massive ore in which the pyrite is coarse and the quantities of sphalerite and chalcopyrite are appreciably varied. Galena, arsenopyrite, magnetite, and tetrahedrite-tennantite are present in accessory amounts. Barite and calcite are present with this ore type, but mainly these gangue minerals are in zinc-rich ore only. The sphalerite ore in many places is in bands or layers; the parallelism of what Helovuori calls "relict sulfide bands" in the mineralized sericite schist host rock he thinks are suggestive of primary layering. The alternative, of course, is that primary replacement was much more readily accomplished along original bedding planes. The breccia structure appears prominently displayed in parts of the massive ore.

The second (disseminated) type of ore is found in a zone that is transitional between massive ore and unmineralized country rock. These are, Helovuori says, replacement ores with the sulfides replacing schist minerals. If schist minerals can be replaced by sulfides outside massive ore, why cannot schist minerals have been replaced in fragments (showing bedding) within massive ore?

The third type is pyrrhotite-bearing ores; it generally is concentrated in the southern part of the deposit. In the northern end of the ore, pyrrhotite ore is randomly distributed as numerous pyrrhotite veins in massive ore or as irregular accumulations or disseminations. Helovuori (1979) believes that this pyrrhotite filled fractures after the massive sulfides had been deposited.

The fourth type is pyrrhotite-selvage ore. Such ores are in the southeastern part of the ore, localized where pegmatite contacts massive ore. Some chalcopyrite and less sphalerite and pyrrhotite occur in this situation; pyrite is present only as euhedral crystals. Helovuori believes that this type of ore was produced by thermal metamorphism induced by the heat of the intruded pegmatite.

The fifth type is pyrite ore that occurs in rounded or fractured phenocrysts in a fine-grained matrix of pyrite, sphalerite, and some chalcopyrite. The gangue is mainly quartz and oligoclase. Also included under this heading are porphyritic ores in which the matrix is well-supplied with chalcopyrite. Helovuori believes that these ores were formed from normal massive pyrite ore by movements induced by recrystallization. These zones of movement are normally between 10 to 50 centimeters wide and may be rich in sphalerite as well as chalcopyrite. These zones are close to the contacts between ore and country rock in volumes in which large rock fragments are common.

Helovuori also describes what probably should be a sixth ore type in which ore minerals have accumulated at boundaries between ore and country rock and at the contacts of fragments of country rock and ore. Their main difference from the other ore types is their markedly different composition. The main minerals in this ore are accessories in the other ore types, for example, arsenopyrite, galena, different arsenosulfides, tellurides, molybdenite, and native silver and gold. These minerals also are present as veinlets and irregular accumulations in the country rock close to ore (?) contacts. The close relationship of these "accumulations" and veins near normal massive pyrite ore suggests to Helovuori that they were formed in the final stages of the ore-forming process. He believes, however, that he can demonstrate (he does not say how) that some of the chalcopyrite, where it is abundant in this ore type, was produced by remobilization caused by tectonic movements after the emplacement of the main ore.

Helovuori thinks that the country rock fragments with disseminated pyrite that are enclosed in massive ore indicate that the pyrite in these country rocks was deposited well before the massive pyrite ore was formed. Again, this relationship could have resulted from various degrees of replacement in various portions of the country rocks in which the reaction occurred.

Tourmaline, in accessory amounts, has been reported from throughout the ore body as well as concentrated in veins with quartz.

Helovuori (1979) reports that the composition of the Pyhäsalmi ore averages: barite - 7.42%, pyrite - 63.48%, pyrrhotite - 4.44%, sphalerite - 7.04%, chalcopyrite - 2.04%. The total of this list is 84.92 which indicates that the remaining 15.08% is composed of non-metallic gangue minerals. The sample from which these data were obtained was taken during the exploration stage and probably is no more than an approximation of the mineral content of the ore now being mined. The content of the ore in lead and zinc differs considerably from one place to another in the mine; the highest contents of sphalerite are directly associated with those of barite. The contents of Au, Ag, and Hg in the ore are low (e.g., Au = 0.2 ppm). What concentration there is of mercury places the highest value (75 ppm) in the zinc concentrate.

The delta-^{34}S values are concentrated rather closely around +7.5 per mil; the distribution of sulfur isotope ratios in the "eastern parallel ore body" is slightly less positive than in the main ore. The amphibolite west of the ore contains a minor amount of disseminated pyrrhotite in which the delta-^{34}S values are about -9.5 per mil; this strongly suggests a different history for this pyrrhotite than for the sulfides in the ore. This concentration of delta-^{34}S values of the ore closely around +7.5 per mil indicates (to me) that the ore-sulfur probably had a magmatic source, but where the magma originated and where the ore fluid escaped from it are uncertain.

Lead-isotope ratios from the Pyhäsalmi ores indicate (to Helovuori, 1979) that the lead therein contained has an age of 1970 million years and that the lead-isotope date suggest a mantle source for that lead. He also ways, however, that (according to the Stacey-Kramers two-stage model) the somewhat higher values for the ^{207}Pb/^{204}Pb ratios than those at Outokumpu indicate that, despite the source just stated, the Pyhäsalmi deposits have a somewhat more continental character than the leads at Outokumpu. He goes on to say that the leads in Finnish "syngenetic sulfide deposits," including Pyhäsalmi, do not fit the primary growth curve for terrestrial leads. This statement that the deposits are syngenetic may be correct, but saying so alone does not make it so. He remarks, however, that the ore conforms roughly with the sericite schist zone and that "the complicated structures and forms of the ore proper and its environment can be attributed to polyphase deformation." His analysis of the deformation indicates to him that it consisted of three phases that resulted in the present structures exhibited by the Pyhäsalmi rocks and ores. This structure analysis seems to be correct, but his attribution of the structures of the ore to this deformation assumes that the ores were introduced before the deformation began. At the same time as the brecciation of the carbonate and siliceous rocks that were components of the original stratigraphic sequence, the ore sulfides (once interbedded with the associated rocks) were remobilized and flowed into fractures in these rocks (more competent than the ores according to Helovuoir) to produce the structures now found there. Of course, the sericite-schist zone around the ore almost certainly is connected with the processes that produced and affected the ore bodies. It may have been the result of reactions promoted by the deformation of the ores or they may have been caused by the ore fluids that deposited the ores and also affected the rocks surrounding them. When this development of sericite schist took place relative to the deformation of the rocks of the ore district is assumed by Helovuori to have been fairly late in the metamorphic process, just as it would have been if the sericite envelope had been developed by ore fluids that deposited the ores late in the deformation process. Further work is necessary before a final verdict can be reached on the stage, in relation to the phases of deformation, when the mineralization took place. The automatic assumption, so common these days, that a strata-bound deposit must be syngenetic with the rocks containing it, is not always certainly correct as can be seen from the numerous epigenetic characteristics exhibited by a variety of strata-bound deposits in rocks that are essentially unmetamorphosed, such as the Kuroko deposits.

The lack of fluid inclusions in the Pyhäsalmi sphalerites prevents any real estimate of the temperature of the deposition of these sphalerites, although Helovuori thinks that this lack may indicate that the ores were essentially dry during their last remobilization and recrystallization.

Isokangas (1978), as opposed to Helovuori, says that the ore at Pyhäsalmi was introduced at about the same time as that at Vihanti, about 2,055-2,075 million years ago. Although the host rock of the zinc-rich ore of Vihanti is metamorphosed carbonate rocks, the pyritic ore in that mine is mainly in quartzitic rock, but some is contained in skarn or altered mica-rich rock. Thus, the pyritic ore at Vihanti is more similar to that at Pyhäsalmi than is the Vihanti-zinc-rich ore.

Isokangas (1978) quotes the resident geologist at Pyhäsalmi (O. Helovuori) to the effect that that ore is syngenetic in its manner of emplacement (as Helovuori confirms in his 1979 paper). His observations indicate that the intense sericitization of the host rock is confined to the immediate vicinity of the ore body. This sericitization, plus the accompanying disseminated pyrite in the sericitized rock, probably occurred during the first stage of mineralization. After the massive ore was developed, the altered host rocks were appreciably deformed, with the ore being folded conformably with the fold structures of the host rocks. Boudinage structures developed in the country rock intercalated in the ore indicate to Helovuori that the deformation stage continued appreciably after the ore had been emplaced. The ore body dies out abruptly at its southern end against a fold that has an axial plunge of 70°S and that probably was pre-ore; to the north, no such fold was present to act as a barrier to the ore fluids, so minor impregnations of ore minerals can be followed for several kilometers north of the massive ore.

Isokangas considers that the ores were remobilized in several stages; he thinks that this is shown by veins and fracture fillings of chalcopyrite and fracture fillings of various ages of pyrite that cut the ore and, in the case of the chalcopyrite, the country rock adjacent to the contact of ore and country rock. He also thinks that some small lenses of pyrite were formed by remobilization of ore induced by tectonic movements. The possibility, of course, exists that these later ore veins and lenses were, instead, the result of deposition from further introductions of ore fluids into rocks broken by earth movements.

The massive pyritic ore in contact with the pegmatite, southeast of the main ore body may have resulted from the contact effects of the pegmatitic magma on pyritic ore already emplaced.

The ore of the Pyhäsalmi ore body is typical of that of deposits produced by the reaction of high-temperature ore fluids with mainly non-carbonate rocks. Almost all the minerals present, both sulfides and gangue, are compatible with formation at high temperature and pressure. Only the very minor quantity of tetrahedrite does not fit in this category, but it is not enough in amount to place any of the ore in the mesothermal or leptothermal range. Some of the chalcopyrite and sphalerite may have been introduced under mesothermal conditions, but no work has been done (or at least reported) to indicate whether exsolution textures were developed in either of these minerals or not. The low content of galena in the ore at Pyhäsalmi also decreases the chances of any of the ore having deposited in the mesothermal range. So far as currently available information goes, therefore, it appears best to categorize these deposits as Hypothermal-1, plus probably Metamorphic-C.

If it turns out that the Pyhäsalmi deposit was deposited syngenetically with the surrounding rocks, as may be the case, it will have to be classified as Sedimentary-A3, Metamorphic-C. From what I have seen of Vihanti (I have not seen Pyhäsalmi) I cannot believe that the ores were introduced until appreciably after the enclosing rocks had been lithified and somewhat metamorphosed. The ores certainly were affected by the later phases of metamorphism, but I doubt if the mineral content or mineral arrangement was greatly changed by the metamorphic process.

VIHANTI

Middle Precambrian	Zinc, Lead,	Hypothermal-2 (Zn-Pb)
	Copper, Silver,	Hypothermal-1 (pyrite)
	Gold, Pyrites	Metamorphic-C (minor)

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Notes

Vihanti (64°29'N, 25°00'E) is one of the three largest sulfide deposits in Finland (the others are Outokumpu and Pyhäsalmi) that lie in a zone that extends from Raahe (64°42'N, 24°30'E) on the Gulf of Bothnia to Lake Ladoga. Along the same line on the west side of the Gulf, in Sweden are the deposits of the Skellefte area. Many other deposits still in the prospecting stage (Rouhunkoski, 1968) and similar to Vihanti and Pyhäsalmi are in the Finnish part of this belt between Raahe and Kuopio (62°54'N, 27°40'E) and Rouhunkoski has designated this area as the Raahe-Kuopio metallogenic province. The Vihanti mine is slightly over 30 km southeast of Raahe. Zinc-lead ore at Vihanti is mined mainly from the Ristonaho shaft and the pyrites from that of Lampinsaari, but an appreciable and additional body of zinc ore is in the Lampinsaari section. The first real indication of ore in the Vihanti area was in a sphalerite-rich skarn boulder found in 1929, 12 km southeast of the present mine by a man from the area; exploration was carried out to some extent during World War II, and several parallel zones of ore minerals were found by the Geological Survey of Finland. In 1951, ownership of the property was transferred to the Outokumpu Company by which organization it still is operated. From 1954 through 1966, nearly 5 million tons of ore, averaging 10.78% Zn, 0.75% Cu, 0.51% Pb, and 0.36 gm/ton Au and 25.4 gm/ton Ag were recovered from the Vihanti mine, this being prior to the full-scale recovery of pyrites from the Lampinsaari shaft area. From 1965 through 1969, 2,250,000 tons of ore were mined, with an average grade of 8.04% zinc, 0.52% copper, 0.48% lead; during that same period, 622,000 tons of pyrite ore were milled, with a sulfur content of 19.8%. From 1970 through 1974, 2,450,000 tons of ore were mined, with an average grade of 6.96% zinc, 0.47% copper, and 0.45% lead; during the same period, 1,300,000 tons of pyrite ore were milled, with a sulfur content of 16.1%.

The rocks of the Vihanti district lie near to the boundary between the areas of terrane that once were known as Svecofennian (southwest) and Karelian (northeast) and are now designated as Svecokarelian because, despite an appreciable difference in the degree to which they were metamorphosed, they almost certainly are of essentially the same age [see Finland (General) in this volume].

The latest division of these Svecokarelian rocks in the neighborhood of the Vihanti mine is given by Rauhamäki, and others (1980) as, from top to bottom:

Rock Designation		Age in m.y.
Silicic and intermediate intrusives		1700 to 1900
Mafic and Ultramafic plutonic rocks		1900 to 1950
Upper geosynclinal unit	metaturbidites	
	quartzite meta arkose	
	conglomerate	
------------discordance------------		
Lower geosynclinal unit	mafic and silicic volcanics	
	mica gneisses (meta-graywackes)	
	mafic volcanics epi-continental meta-sediments	Vihanti area proper
---------great discordance---------		ca. 2500
Presevcokarelian basement complex		ca. 2800

In the Vihanti area proper, the supracrustal rocks form the lower portion of the Lower Geosynclinal unit. Originally, they were poorly sorted pelitic and graywackes sediments; now they are heterogeneous, partly migmatitic biotite gneisses and biotite hornblende gneisses. In addition to these metamorphosed products of clastic weathered materials, these supracrustal rocks include ever-present volcanic material both as the actual volcanics (metamorphosed) and beds composed of their (metamorphosed) fine-grained weathered products.

This volcanic activity was developed in two phases: (1) the older of the two was mainly basaltic to rhyolitc with dacites being the most common; the silicic volcanics of this phase, together with dolomites and skarns, a constitute zone designated as "Lampinsaari-type rocks" that can be followed, with some breaks, for several tens of kilometers and contain traces of copper-zinc sulfides and a zone of uraninite- and apatite-bearing rocks that have a "distinct regional distribution" and (2) a younger mafic and silicic volcanic series that is not known in the Vihanti district and is apparently unmineralized where it is.

The district has been intruded by three syntectonic infracrustal rocks - gabbro, granodiorite, and granite, to which the Vihanti ores may, if they are hydrothermally deposited in solid rocks, be genetically related. If the Vihanti ores are syngenetic (which majority opinion favors), then the ore fluids that deposited them in the predominantly mica schist beds of which the Lower Geosynclinal unit in the Vihanti district is now composed derived from the same igneous sources as the earlier volcanics.

The Vihanti ores are contained in an area of predominantly mica schists about 10 km long and 2 to 3 km wide that is surrounded in the main by granodiorite and microcline granite. The schist zone was affected by multiphase folding in which the two trends that have been recognized. A southwest-trending synform-antiform structure in which the synforms of the supracrustal rocks are separated by antiforms of gabbro and provide the more intense folding of the two. The axes of the structures plunge gently to the northeast. The less intense folding produced northwest-trending structures. This folding differs from that in the central Bothnia area in that northwest-trending folding clearly is more important south and southwest of Vihanti.

In the southern part of the district, a late-orogenic microcline granite occurs as cross-cutting veins.

The mica schist zone at Vihanti has a boomerang shape with the open end to the north. The rocks of the mine area are divided into those of silicic volcanic parentage and calcareous varieties; the former group contains mica schist, cordierite gneiss, silicic volcanics, and schist; the latter is composed of dolomitic rocks and skarns derived from them. This scheme is complicated by gradations between rock types and mixtures of them. At Vihanti, the calcareous rocks occupy a central position with the volcanic rocks lying around the edges of the ore-bearing complex, but intercalations of various thicknesses of one type-group occur in the other. Tops and bottoms of the beds cannot be determined because of the metamorphism that they have undergone, so it cannot be established if the rocks are right-side-up or upside-down.

The ore complex (Rauhamäki, and others, 1980) is made up of silicic volcanics, dolomites, skarns, and cordierite gneisses (that occur mainly in the western part of the district); all of these rocks are crosscut by granitic and mafic dikes. The silicic volcanics include homogeneous quartz porphyries, banded tuffs, and tuffites, and metamorphism has complicated the problem by producing skarn-banded silicic volcanics. The ore complex is surrounded by intensely metamorphosed mica gneiss with amphibolite intercalations. Movements of the rocks during metamorphism have caused appreciable brecciation. The complex strikes about east-west, and the dip is toward the south.

The intercalated skarns and dolomites may be as much as several tens of meters in thickness and normally have silicic volcanic rocks and graphitic tuff associated with them. Outside the ore-containing area, the carbonate and silicated skarn rocks are thinner and less well-defined than in the actual mine surroundings. In the ore zone, the ore is largely confined to a calcareous layer in mica schist.

Although the Vihanti structure has not been thoroughly worked out, Mikkola's description (1963) seems quite reasonable. The main ore bodies are on the northern limb of one of the synforms; drilling has developed pyritic mineralization on the southern limb in the same structural position that the ore bodies occupy on the northern one, but these south-limb bodies are largely on the quite straight hanging wall; the axial plane of the main synform has a dip of 70° to 80°S at the east end of the ore area, but the dip has dropped to 40° to 50°S at the west end. The axis of the fold plunges 10° to 20°W. The main synform is overturned to the north. The northern (actually northwestern) limb (Mikkola, 1963) is a simple structure, whereas the southern (southeastern) one has numerous small folds and dragfolds. The rock cleavage is parallel to the axial plane, its strike being generally N75°E, and bedding parallels the cleavage except in the crests and troughs of folds. Faulting is quite common in the Vihanti mine, being both of the strike and cross varieties. The strike faults are pre-ore; in the Ristinaho ore body a strike fault adjoining the footwall of the ore is mineralized in the same way as the ore, except that the grain size is much finer than in the ore. The displacement along the strike faults is negligible. The cross faults are more common than the strike variety, and they had considerable effect on the geology of the ore bodies. These faults strike about N40°W and dip steeply to the southwest; all show some displacement that ranges from a few centimeters to several tens of meters. This displacement is such that the southwestern sides of the faults have dropped down and to the southeast.

These faults are filled with post-ore porphyrytic plagioclase-hornblende dikes, and these dikes were fractured by later earth movements; these later fractures are filled by such mineral as calcite, zeolites, quartz, apophyllite, barite, and iron sulfides. Where these dikes cut the ore, they usually are broken. The fragments of dike rocks in such instances are cemented by ore minerals, whereas fractures in fragments are filled with serpentine or the minerals mentioned in the preceeding sentence.

Mikkola states that the ore bodies are located in a definite horizon in the mica-schist series, and the most favorable location has been the base of the dolomite bed contained in the schist. Most of the ore is in tremolite skarn, but unaltered dolomite and the narrow quartz bands also are mineralized. It would appear that, since the major portion of the dolomite bed has been converted to skarn, the bulk of the ore is in that rock type, not so much because it was more favorable for attack by the ore-forming fluid, but because there simply was much more of it than of unaltered dolomite and quartz bands. A more important factor in controlling the localization of the ore than rock type was the minor folding on the limbs of the main syncline.

The flanks of the complex were affected differently by the folding. On the southern limb, which contains the largest of the massive pyritic sulfide ore bodies, is quite straight, but the northern limb where the major zinc ore bodies are found is drag folded and the ores are localized by these folds. Välisaari (the most south-westerly of the zinc ore bodies) and Ristonaho (more north-easterly) are in the largest and uppermost drag fold. The middle drag fold contains the Lampinsaari ore body, and this and a small ore lens that lies near the lower edge of that ore body are connected to the middle fold. The lower fold hosts the B1 and B2 ore bodies. In Figure 2 in the paper by Rauhamäki, and others (1980) various ore bodies are shown (but not discussed) that do not appear in earlier papers. These are: (1) the Ramesaari ore body that appears to be pyritic ore that underlies the eastern end of the Välisaari-Ristonaho ore body; this ore body in its eastern section (above the Ramesaari) also is pyritic. It is possible that the Ramesaari ore body has been projected into the plane of the section and may not truly underlie the Ristonaho ore; (2) another pyritic ore body (not described), the E ore body appears to underlie the Hautaräme ore body at the western end of that body. The same caveat applies here - the E ore body may have been projected into the plane of the section; and (3) a third undiscussed ore body - the Isoaho - lies deeper than any of the other Vihanti ores, extending downward to below the 700 level; except for the E ore body, which goes down to about the 450 level, the Isoaho is the only body known below the 400 level. Again the position of the Isoaho relative to the plane of the section in Fig. 2 is not clear.

In addition to the zinc ore masses, there are two pyritic ore bodies, the Hautakangas and the Hautaräme; the former is located almost due south of the Lampinsaari ore (on the 200 level). On that level, the Hautakangas is a narrow body, over 400 m long with its long dimension extending from east to west. The Hautaräme ore body is southeast of the western portion of the Välisaari mass; it is not much more than 100 m long (on the 200 level) and is appreciably wider than the Hautakangas. Both of the pyritic ore bodies include some zinc ore, but the proportion of zinc ore to pyrite is much lower than in those designated as zinc ore bodies.

The zinc ore bodies normally are located at the contact between the calcareous beds at the center of the mica schist-skarn complex and the northern quartzitic bed. The Välisaari-Ristonaho combined ore body is about 900 m long and 5 to 50 m wide; it is in the upper part of the favorable zone. The Lampinsaari ore body is about 400 m long and 19 to 40 m wide and is the lower part of this favorable zone. The main zinc ore bodies and their minor satellites commonly are arranged in echelon, with the largest ore body to the west and the others becoming progressively smaller to the east.

The Välisaari-Ristonaho ore body is cut into separate fractions by several faults, and the Välisaari fault divides the ore into the Ristonaho to the east (actually east-northeast) and the Välisaari to the west. This fault has moved the western (Välisaari) part southward for about 40 m. The other major fault in this area is the Hautärame, and it also has displaced the ore to the south on the western side and dropped it down a few meters. The Ristonaho fault cuts the ore body of that name still farther to the east; on this fault, the western side has moved the northward and down for more than 30 m. All of these three faults strike more or less northwest to southeast. The net result is that the western half of the Ristonaho ore body has been displaced northwestward in relation to the eastern Ristonaho and the Välisaari. The Ristonaho has been mined out; the Välisaari was intact in 1968. No ore is known to lie beyond the eastern end of the Lampinsaari ore body, unless the B3 and B4 ore bodies (Rouhunkoski, 1968, Fig. 5) can be considered as northeasterly continuations of the Lampinsaari ore. The Lampinsaari fault also strikes about northwest-southeast, but it does not cut the Lampinsaari ore body, and only a small bulge at the east end of the mica schist-skarn complex lies east of it.

Rouhunkoski (1968) believes that the ores were formed by hydrothermal solutions rather than by sedimentary, plus metamorphic, processes. The dating of the activity that provided the solutions cannot be certainly done; the lead in potassium feldspar of the granite dike that cuts the ore has appreciably different isotopic ratios from that of the lead in the ores, and this surely debars the chamber from which the magma of the granite dikes came from being the source of the Vihanti ore fluids, granted the ores were hydrothermal and not volcanic-exhalative. The ages Rouhunkoski derives from these two leads are 1860 m.y. for the lead-in-feldspar in granite and 2075 m.y. for the lead-in-ore. He goes on to say that the 2075 m.y.-lead is Karelian and the 1860 m.y.-lead is Svecofennian. In this context, these last two terms are confusing because of the present-day agreement that the difference between Svecofennian and Karelian rocks is not in their age but in the degree of metamorphism that affected them. Thus, the age of both leads would currently be said to be Svecokarelian with the lead-in-ore being appreciably older than the lead-in-granite. This does not change the assignment of a middle Precambrian age to the Vihanti ore. In the Otanmäki area (see Otanmäki in this volume), a granite (the Honamäki) is dated at 2100 m.y., suggesting strongly that appropriate magmatism may have occurred at about the time when the Vihanti ore fluids should have been generated. Obviously, this is not proof that such activity took place at the appropriate time in the Vihanti area, but it certainly does not argue against it. Rouhunkoski believes that rocks of the mica schist-skarn complex differed so much from those around them, both in rock type and degree of folding and fracturing, that they attracted the ore fluids to them as a site for ore deposition. Why the zinc ore deposited largely in the skarns and the pyritic ores in the silicic volcanic rocks probably can be explained by the different ways in which the one ore fluid reacted with the two rock types. Rouhunkoski's explanation that the ore fluid fractionated to give a zinc-rich solution to deposit the zinc ore and an iron-rich one to deposit the pyritic ones seems most unlikely. Rouhunkoski thinks that the ore fluids came upward from the west as this is the natural direction for upward flow. On the basis of this discussion, here are classified as late middle Precambrian.

The main portion of the zinc ore bodies is made up of massive layers of sulfides with a zinc content that, in many places, is greater than 20 per cent. These ore layers extend for tens of meters along strike but are of a shorter dimension down dip. The layers pinch for short distances but are mainly rather constant in size, ranging between a few centimeters to two meters in thickness. The different proportions of the sulfide minerals in the various bands of which the ore layers are composed gives the ore a banded character that is parallel to the bedding of the enclosing country rocks. This banding is further emphasized by intercalations of quartzite and skarn that are less well mineralized than the bulk of the beds that are the main host of the ore. In places, barite is an abundant constituent of the ore and adds to its banded aspect. In a few localities, the ore extends outward into the quartzites; in this rock type, the ore generally follows the bands in the quartzite but narrow sulfide veins and streaks may cut across the rock structure.

Where skarn is intimately associated with ore sulfides, it normally contains a high proportion of diopside, but pure tremolite skarns are not uncommon. Where ores have dolomite or skarn matrix, intercalated bands of graywacke, quartz-feldspar rock, or quartzite may be present; these intercalations usually are no more than weakly mineralized. Frequently, ore beds have been broken and the fractures have been filled by ore minerals; Rouhunkoski considers this ralationship was produced by remobilization of already present ore sulfides, but it well may have been caused by fracture filling at the same time that the ore entered into the rock system as banded replacements. Ore normally follows bedding, but in places, where schistosity cuts the bedding, the ore also follows the schistosity, suggesting that folding and schistosty preceded the introduction of the sulfides.

In one ore body, the B3, the ores are not bedded; instead, sphalerite and abundant pyrite form varied disseminations and patches or schlieren that parallel the major axes of the ores but are no more than a few meters in length.

Joints that run in several directions have been found in the ore with the most prominent being parallel to the major transverse faults; less important jointing occurs in longitudinal and horizontal directions in the ore and in places is diagonal to that direction.

The sphalerite ore at Vihanti (Rouhunkoski, 1968) averages sphalerite 20 per cent, chalcopyrite 2 per cent, galena 0.8 per cent, pyrite 1.0 per cent, and pyrrhotite 5.0 per cent; gold runs 0.4 gms per ton and silver 30 gms per ton. In addition to antimony-bearing sulfides (see below), the ore includes such minerals as cubanite, valleriite, arsenopyrite, stannite, molybenite, and native gold and silver. Magnetite, rutile, uraninite, and graphite are present, but these may have been inherited from the host rock. Rouhunkowski believes that the ore was introduced in two generations, with the banded ore being the first and the fillings of transverse joints being the second. The second generation was produced, he thinks, by preferential remobilization of minerals of the first. Appreciable differences exist between the first and second generations; the second contains pyrrhotite, chalcopyrite, and sphalerite in that order of abundance with local concentrations of galena and sulfosalts. This second generation is quantitatively minor in comparison with the first, but he considers that the remobilization affected the "most mobile elements of the ore." The average iron content of sphalerite is 7.2 per cent, with the highest iron being in coarse sphalerite in transverse fractures. The sphalerite replaces carbonates and silicates and, along with other sulfides, fills narrow fractures in such non-sulfide minerals. Some sphalerite has lamellar twining that Rouhunkoski considers resulted from deformation during or after deposition.

Chalcopyrite is quite uniform through the ore bodies but has a tendency to be concentrated, with pyrrhotite and pyrite, along the wall-rock borders; there the copper content may be between 2 and 3 per cent. In places, particularly in the hanging wall of the Ristonaho ore body outside the zinc ore proper, chalcopyrite and iron-sulfide mineralization gradually fades out over several meters. Copper in this disseminated ore is 1.5 to 2 times richer than in the massive type. On the other hand, such disseminations in the hanging wall of Välisaari and Lampinsaari ores are less enriched than around the Ristonaho ore. Where disseminated sulfides are in the intercalated skarn, the copper content is higher than in the main ore bands. Minor inclusions, probably exsolution bodies, of sphalerite (as stars) and pyrrhotite occur in the chalcopyrite. Most commonly chalcopyrite appears to replace sphalerite, but locally mutual replacement of one sulfide by the others has been seen by Rouhunkoski. Chalcopyrite also is known to contain cubanite lamellae, and lamellae may be cut by pyrrhotite stringers that derive from pyrrhotite rims around chalcopyrite.

Galena in the ore is varied in amount; in many places, it occupies the farthest extremities of the ore bodies and favors coarse-grained skarn on the flanks of the ore bodies and has its highest concentrations at contacts and in fractures in fragments of quartzite and porphyrite in the ore. Galena is unusually rich in the Välisaari ore body and in the lower portion of the Lampinsaari.

Pyrrhotite, in almost all instances, is more common than pyrite and is concentrated in brecciated ore-body contacts and in disseminations in skarns. The percentage of pyrrhotite (as does that of pyrite) differs from one layer to another in the ore.

The most common members of the antimony-arsenic sulfosalts at Vihanti are tetrahedrite, tennantite, gudmundite, and boulangerite; pyrargyrite and niccolite have been identified, jamesonite, bournonite, freibergite, and berthierite may be present. Gold occurs in cracks in tourmaline and in association with galena and the sulfosalts.

A small amount of supergene alteration has been preserved in the upper parts of the ore bodies by which pyrrhotite has been converted to pyrite, marcasite, and limonite; chalcocite is rare in this alteration.

Barite averages 5 per cent but ranges up to over 30 per cent in the ore and occurs either as streaks parallel to the banding or as almost monomineralic rock, with only small quantities of sphalerite. In a few places, fluorite is present in altered tremolite skarn, and brownish tourmaline is somewhat more abundant; all other minerals in the ore, Rouhunkoski believes were inherited from the host rock.

The pyritic ores at Vihanti are those portions of the complex where pyrite and pyrrhotite are dominant and where the zinc content is low. These pyritic deposits are of two types; (1) those at the southern edge of the Lampinsaari ore and, (2) the iron sulfide-dominant masses closely associated with the zinc ores. Type (1) bodies are economically far more important; the pyritic ores of type (2) make up less than 5 per cent of the known (Rouhunkoski, 1968) pyritic ores.

The largest concentrations of pyritic ore is in two successive lenses on the southern edge of the ore complex. One of these is the Hautakangas body that is located almsot entirely between the 200 m level and the surface. This ore body is over 700 m long and ranges in width from a few meters to over 20; it dips about 45°S. The thickness and grade of this generally incoherant ore body differ considerably over short distances; locally narrow ore bands may be separated by sulfide-poor country rock layers that are several meters thick. At its eastern end, the ore body grades from pyritic to pyrrhotitic and then into a narrow zinc ore body that dies out in a few tens of meters. The host rock is mainly quartzite but locally it is calcareous material.

The second of these ore bodies, the Hautaräme, is in the same horizon as the first but lies farther down, between the 150 m and 500 m levels. Although this ore body plunges 25°SW, the internal swell and pinches in ore body occur in a mainly horizontal direction. At its edges, the ore body generally dies out by pinching into numerous short tongues. This second pyritic ore body is considerably more regular than the first and is higher in grade.

Although the second pyritic ore body is mainly in quartz-rich rock, it penetrates the altered skarns below and is localized by the contact between the quartzitic and calcareous skarn rocks. The hanging wall host rock ranges from quartz-feldspar rock at the eastern end, through a mixture of quartz-feldspar and cordierite gneiss on the west. The country rock of the footwall is talcous and sheared cordierite-bearing rock. The lower portion of this ore body lies about 15 to 20 m above the contact of the talcose and sheared cordierite-bearing material (above) and the mica schist (below). As the ore is followed toward the surface, the distance between ore and mica schist increases to between 30 and 40 m. The ore body generally conforms to the dip of the complex; this dip is about 40°S.

In the calcareous center of the complex are small pyrrhotite-rich bodies that are in oblong lenses parallel to the strike and dip of the complex, although locally the bodies cut across these. Similar iron sulfide-rich bodies occur at contacts between zinc ore and wall rock both in the foot- and hanging walls of the Välisaari and Ristonaho ore bodies; the amount of such sulfide bodies is small in comparison with the mass of zinc ore. Somewhat larger iron-sulfide bodies (B_3 and B_4) are found at the NE ends of the Lampinsaari ore body. The host rock of this iron-sulfide ore differs considerably from one of these bodies to another; the host rock is mainly heterogeneous skarn-dolomite, with partly brecciated intercalations of quartzite. The iron-sulfide bodies are rather high in chalcopyrite (0.5-1.5 per cent Cu).

As is to be expected, the main mineral of the pyritic bodies is pyrite, its content ranging from 30 to 69 per cent, with the Hautaräme ore being the richest. The pyrrhotite content normally is 10 to 20 per cent of that of pyrite, although the proportions locally may be reversed. The pyrrhotite is both hexagonal and monoclinic and microscopically shows a lamellar texture. The host rock appears not to influence the pyrrhotite-pyrite ratios; the contacts between the pyrrhotite-rich outer margins and the pyrite-rich cores is generally clear but not always sharp, and not all pyrite-rich bodies have pyrrhotite-rich margins.

In addition to the iron sulfides, the pyritic ore bodies normally include small amounts of chalcopyrite and sphalerite, with a little galena. These minerals, plus pyrrhotite, occur as a fine-grained matrix material in the spaces between pyrite grains or in cracks within them. Since chalcopyrite is almost universally associated with pyrrhotite in the pyritic ore, the copper content is highest where pyrrhotite is most abundant. Pyritic ore in the eastern end of the Hautaräme ore body grades into rich pyrrhotitic material and, this, in turn, grades into a narrow band of zinc ore.

Mafic dikes occur in the rocks around the ore complex and in the ore itself where they are generally parallel to the bedding. The width of the dikes normally is measured in decimeters, and their main minerals are plagioclase and hornblende with some minor, but varied, amounts of biotite and quartz. Dikes in the ore are locally, at least, rich in sulfides, with chalcopyrite and pyrrhotite being the most abundant. Except for these places where they cut the enclosing beds at low angles, they might be considered as conformable beds of amphibolite. These dikes almost certainly are older than the ore. All other dikes in the area cut, and are younger than, the mafic dikes; these younger include plagioclase-porphyrite, granite, and pegmatite dikes. These silicic dikes probably all are younger than the ore, and strike generally NE-SW and NW-SE.

The dolomite skarn rocks of the area, which form the central core of the ore-containing syncline, are surrounded on both foot- and hanging wall by cordierite gneiss, silicic volcanics and graphitic tuff. These latter rocks are, in turn, surrounded by mica schist that provides the outer shell of the rock complex that contains the ore. The ore is located in the footwall of the dolomite-skarn body. The mica schist on the north (hanging wall) side of the ore is a rather fine-grained biotite schist, with local and abundant garnet in what Mikkola (1963) calls porphyroblasts and considerable quartz and plagioclase. On the south side, the porphyroblasts (?) are of cordierite, staurolite, and andalusite that have been converted by retrograde metamorphism to pinite and sericite. Above (south of) the footwall mica schist, the rock is mainly quartzite, and partly replaces the entire quartzite thickness. The cordierite gneiss is high in magnesium and, in addition to cordierite, contains phlogopite (partly altered to chlorite) and talc, plus minor fine-grained corundum. Downdip, the quartzite grades into black schist; the color is due to graphite, biotite flakes, and finely disseminated iron sulfides; it also contains quartz and some plagioclase. Such primary calcareous bands as the schist contains are altered to tremolite-bearing schist. All of these three rocks belong to the same stratigraphic horizon with the contacts being gradational from quartzite into the rocks above and below it.

The skarn economically is the most important rock in the mine. Mikkola (1963) considers the skarn to have been developed on a regional scale from the reaction between dolomite and rocks rich in silica. The most common skarn minerals are tremolite and diopside; small amounts of quartz, plagioclase, and calcite are associated with these two silicates. Accessory minerals include tourmaline, apatite, corundum, and sulfides; talc and serpentine were produced as alteration products of the two silicates and probably were formed at the same time as the ore was introduced. [This last assumption agrees with Rouhunkoski but not with Mikkola.] In the skarn-dolomite association, the dolomite is on the hanging wall; it contains narrow quartzite bands and is partly altered to skarn. Forsterite was presented in minor amounts, and it is now altered largely to serpentine.

Rouhunkoski reports (1968) that the average delta-^{34}S value of the ore sulfides at Vihanti is +7.5 per mil; this, of course, means that the ore sulfides have been enriched in ^{34}S in comparison with the meteoritic standard. A similar shift can be observed in the sulfides disseminated in the skarn and in the cordierite gneiss. On the contrary, the sulfides in the black schists have an average delta-^{34}S value of -5.6 per mil, showing enrichment in ^{32}S in comparison with the meteoritic standard. These opposing sulfide ratios suggest that the schist sulfides and the ore sulfides were formed by different processes, and that the sulfur may have been derived from different sources.

Assuming, as appears probable, that the initial skarn formation was a regional process, the alteration associated with the Vihanti ores that can be considered as hydrothermal is minor. On the footwall side of the zinc ore, a narrow, weakly bleached quartzitic layer was seen locally; in this place, the plagioclase was strongly sericitized. The development of talc and serpentine in the mafic dike rocks also may have been a hydrothermal (or possibly an autometamorphic) phenomenon. The dolomite in the area almost certainly was regionally produced. The major alteration that can be considered as hydrothermal is the retrogressive conversion of the diopside to tremolite. Rouhunkoski considers that this alteration is closely connected with the ore in time and space. In places, this alteration of diopside to tremolite has proceeded farther, converting the tremolite to mica, particularly in the neighborhood of the iron-sulfide ore bodies. These alterations have caused a strong enrichment of the original skarn rocks in Al_2O_3 and MgO. Such alterations, however, are marginal to the ore rather than appreciably affecting the skarn remnants within the zinc ore bodies.

The main problem in the genesis of the Vihanti ores is to explain why the zinc ores occur in calcareous rocks and the pyritic ones mainly in quartzitic rocks. The fact that the two ore types are near together suggests that they must have been closely related genetically, with the difference in host rock probably explaining the difference in mineral composition of the two ore types. Rouhunkoski considers (1968) that the structure, mode of occurrence, and contact relations of the pyritic ore suggest that they were epignetically formed. To some extent, however, it is possible to argue that the ores were formed in the same manner as the rocks that contain them. This is because some individual layers in the zinc ores can be followed for tens of meters; however, most layers normally pinch out after only short distances. Further, where zinc mineralization is in quartizitic rock, it occurs as coarse-grained and quartz-rich veins or veinlets (Rouhunkoski says streaks) that cut the bedding of the rock. In addition, he points to the obvious (to him) differences between the character of the syngenetic sulfides in the black schists and the sulfides in the zinc ores and to the lack of conformity of ore sulfides with the primary structures of the original sediments. He points out that it is unlikely that the metamorphism the rocks have undergone was sufficiently drastic to have destroyed all syngenetic zinc ore structures (granted there ever were any). The lack of any volcanic activity in the area, he considers, argues against a volcanic-exhalative introduction of the ore elements into the present ore-containing rocks volumes. He also thinks that the gangue minerals, barite, and quartz, indicate that the sulfides (and the gangue) were deposited from hydrothermal solutions. [Considering the syngenetic origins that have been proposed for such barite-containing ores as Rammelsberg, Meggen, Buchans, and Walton not all economic geologists would recognize the validity of this criterion.] Although much of the skarn was produced on a regional scale, the enrichment of Al_2O_3 and MgO in skarns adjacent to ore is considered by Rouhunkoski to indicate a spatial link to ore formation that could hardly have been produced by any solutions other than those that brought in the ore elements. Similarly, the sulfur in the cordierite gneiss is enriched in ^{34}S, as is the ore-mineral sulfur, and differs markedly in isotopic composition from the syngenetic sulfur in the black schists.

In the pyritic ore bodies, the pyrrhotite-pyrite relationship presents certain genetic problems. The pyrite-rich and pyrrhotite-rich portions of the pyritic ore bodies show no characteristics that help to differentiate genetically the one type from the other. The concentration of the pyrrhotite (and chalcopyrite) at the margins of the pyritic ore bodies, is a problem particular to Vihanti. Rouhunkoski believes that the pyrrhotite rims are primary features and were not produced by metamorphic action. He suggests that the ore fluid moved most easily through the central portions of the sites of pyritic ore deposition, and that, therefore, the partial pressure of sulfur was higher than in the margins. Thus, the marginal environment was the only one in which pyrrhotite could remain stable; this seems a reasonable explanation to me and one that eliminates the need to appeal to metamorphic processes to produce the pyrrhotite-pyrite relationship.

None of the discussion in the last two paragraphs argues for a syngenetic manner of formation of the Vihanti ores. Mikkola (1963) said that many features of the deposit point to an epigenetic formation of the ores at Vihanti: (1) the 10 cm wide zone of weakly bleached quartzite in the footwall of the ore in which plagioclase is partly altered to sericite and the micas completely so [he does not, however, think this alteration is too weak for the size of the ore bodies]; (2) he mentions the skarn and the serpentinization of olivine in the skarn as suggesting a hydrothermal formation of the ores but says that the skarn is regional and so even is the serpentinization; (3) he even thinks the cordierite gneiss also is a regional metamorphic effect; (4) certain minerals in the ore might be hydrothermal, the tourmaline in the ore (and in the adjacent skarn and granite) and barite and quartz in the ore. He believes, however, that the totality of these hydrothermal arguments is weak. On the contrary, he considers possible that the Vihanti ore were introduced at the time of carbonate and quartz-grain sedimentation by volcanic-exhalative processes and that the ores are quite similar to those of Rammelsberg and Meggen [although they are much younger than Vihanti and contain much more barite]. He admits, however, that volcanics are lacking in the Vihanti mine area and that any primary sedimentary textures have long since been removed by metamorphism, making a primary sedimentary ore formation difficult to prove [or disprove]. In short, Mikkola, while favoring a syngenetic method of formation does not yet think he can prove that the Vihanti ores are such.

Rouhunkoski, in opposition, believes that the information he has provided, and the interpretations he has put on it, are in agreement with the concept of a hydrothermal manner of ore formation. He suggests that the ore fluids came from the differentiation of a magma produced by a partial melting of basement rocks as (he says) the result of a crustal downbuckling and other processes. This differentiation accumulated and set free magnesium- and sulfur-rich solutions that contained the ore elements as well; he adds that some of the iron may have been removed from the mica schist during the ore fluids upward movement although he cites no evidence of mica schists showing such iron removal.

The presence of very considerable amounts of pyrrhotite and of exsolution textures of chalcopyrite in sphalerite and sphalerite in chalcopyrite strongly suggest a high temperature of ore mineral deposition. The minor amounts of cubanite, valleriite, arsenopyrite, stannite, and molybdenite in the ores are in agreement with this conclusion. The wide-spread, though minor, sulfosalts in the ore suggest that ore deposition may have extended down through the mesothermal into the leptothermal range, but the amounts of these minerals is small and it is not certain that even the galena was deposited in any great part in the mesothermal range, so it is best here to categorize the Vihanti zinc-lead ores as hypothermal in calcareous rocks; that is, Hypothermal-2, and the pyrite ores as Hypothermal-1.

YLÖJÄRVI

Middle Precambrian	Copper, Tungsten Arsenic, Silver, Gold	Hypothermal-1

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Notes

The copper-tungsten deposit of Ylöjärvi (61°37'N, 23°30'E) lies 12 km northwest of the major industrial center of Tampere. Exploratory activities were carried out in the area by the Geological Survery of Finland in the 1930s. In 1937, attention was brought to the Ylöjärvi area by the finding, by a local "rock hunter", of a pyrite sample, which lead to the discovery of a pyrite-bearing outcrop and boulders of the ore-bearing tourmaline breccia. Rocks and ore, similar to those found at and near the outcrop, were discovered on the shores of Paroistenjärvi (lake), the lake giving an alternate name to the deposit. In 1940, the Survey transferred the rights to the deposit to the Outokumpu Oy, which company brought the mine into production in 1942. Despite the low grade of the ore (about 1 per cent Cu), wartime shortages directed that the deposit be mined. The first mining was by open-pit, but underground mining was commenced in 1945. By 1966, the deposit had been exhausted. During the life of the mine, just over 4 million tons of ore were extracted; on benefication, 128,343 tons of copper concentrates were produced that contained 28,322 tons of copper or about 1.6 pounds of copper per ton mined. In addition, the copper concentrate contained 49,500 kg of silver and 270 kg of gold. Further, the rare, unevenly distributed scheelite provided 894 tons of concentrates of that mineral containing 427 tons of tungsten; 2100 tons of arsenpyrite concentrates also were made that had a content of 567 tons of arsenic. The underground mine operated by shrinkage and sublevel stoping.

The Ylöjärvi deposit lies at the west end of the geologically famous Tampere schist belt of Svecokareldic age and is part of the Bothnian group of Svecofennian formations that were deposited under eugeosynclinal conditions. The northern border of this belt is the granite batholith of central Finland; to the south the belt grades into Svionian rocks with an increase in metamorphic grade. These schists and their crosscutting granodiorite intrusion and the Ylöjärvi and Haveri ore deposits have been assigned an age of about 1900 m.y. The oldest work of importance on this schist belt was done by Sederholm (1899).

The Tampere schist belt is made up of volcanic and sedimentary rocks in a state of excellent preservation. These volcanics range in composition from silicic porphyries to mafic porphyrites and include major amounts of tuffites, agglomerates, volcanic breccias, and various types of conglomerates; quartz-feldspar schists and arkosites occur as intercalations among these formations. The conglomerates are rudaceous varieties of the tuffites, arkosites, and graywackes. The basement on which the rocks of this belt were depositedis unknown. The principal igneous rocks are batholithic granodiorites and quartz diorites around which the schists are curved (Isokangas, 1978).

Simonen (1952) says that the most important bedded rocks in the belt range, from bottom to top, from turbidites, through quartz-feldspar schists, arkosites, tuffites, and porphyrites to tuffites and conglomerates with the entire sequence having been deposited during the orogenic revolution. He believes that these sediments may have been derived from areas in which widespread masses of plutonic rocks now exist. During vertical movements associated with the orogeny, both the eruption of volcanic rocks and processes of erosion and sedimentation took place simultaneously. Considerable differences in composition and facies of sedimentation are seen in the basin in both time and space.

Such regional metamorphism as took place produced conditions of the amphibolite facies with retrograde metamorphism, caused by tectonic processes, producing lower-intensity conditions, the end product being sericite schists.

The intrusion of the batholiths caused isoclinal folding in the schists; the depressions and culminations so formed, plus the presence of volcanic rocks, controlled the rock volumes into which the various ores and areas of mineralization were introduced.

The quartz-diorite and granodioritic igneous mass, known as the Hämeenkyrö, was intruded into the schist belt during the Svecokarelian orogeny.

Around the ore bodies themselves, the host rock is tuffites and intercalated porphyrites. Although the bulk of the isoclinally folded schists trend east-west in the general Ylöjärvi area, at the mine they trend southwest and the axial planes of the synclinal folds dip steeply northwest. The tourmaline-quartz rock that contains the copper-tungsten ore is present in the mine area in two nearly vertical zones separated by about 150 m of tuffite and porphyrite. The tourmaline-quartz zones have the same trend as the local sediments and volcanics, but they cut across the dip of the host rocks. The mine was located at the northeastern end of the southeastern (main) breccia zone that has been followed from the shore of Paroistenjärvi to the marginal variety of the granodiorite. Although the main breccia narrows markedly to the southwest, it is about 80 m wide at the mine site. The northwestern breccia zone is appreciably narrower than the main zone, being only a few meters wide in the southwest, but it is at least 700 m long. The brecciation occurs in the intermediate and mafic volcanics of the upper Bothnia group; these rocks are principally tuffites with intercalations of thick beds and lenses of porphyrites. This formation is cut by the granodiorite, resulting in brecciation as far as 1 km from the contact and with the degree of brecciation ranging from incipient to total fragmentation. The fragments of the tuffites and porphyrites range from small pieces to huge masses that weigh several tons; all fragments appear to have sharp edges.

In addition to the two major breccias bodies, tourmaline breccias also occur in smaller, discontinuous lenses and patches that may connect with each other or with the two major zones. In some areas where the tuffite is intensely folded and foliated, tourmaline, of the type found in the breccias, makes up bands and isolated augen that include coarse quartz.

In the mine area, the main breccia zone is displaced by two major faults; these divide the ore into three distinct blocks. The more easterly fault in the south part of the area strikes about N10°E and dips 57°W. The more westerly fault strikes about N22°E and dips essentially vertically. The subhorizontal displacements of the two faults are 170 m for the 57°W dipping fault and 65 m for the one that dips vertically.

The major rock type around the ore deposit is an intermediate tuffite with agglomerate interlayers; variants that are more mafic or less silicic are present locally. Chlorite and epidote and, in places, garnets are the constituents of the thin layers that are intensely folded. Thin and conformable zones of biotite bring out the banded character of the rock. Interbedded agglomerate may be from a few to several 10s of meters in thickness; the fragments in the agglomerate are composed of porphyrite ejecta of varied composition and size. The plagioclase porphyrites are in thick beds and tongues near the marginal phase of the granodiorite; they show sharp contacts with the tuffites and have flow structures and some amygdules. These rocks are more silicic than the tuffites, but their composition ranges from dioritic to quartz-dioritic. The Hämeenkyrö igneous body is about 200 km^2 in area and cuts off the schists to the west of the belt; it is equigranular and slightly foliated and ranges from quartz-diorite to granodiorite. Its main minerals are plagioclase (An$_{30-35}$), quartz, microcline, hornblende, and biotite. The massive pink marginal variety contains euhedral, zoned albite crystals in a fine matrix of microcline and quartz that locally forms a micropegmatite; its chemical composition, nevertheless, is granodioritic. Simonen (1962) considers that the magma, carrying albite crystals in major numbers, consolidated rapidly when it came in contact with the cold host rocks.

The tourmaline breccia in the mine area is a pipe-shaped, roundish body about 80 m in diameter and 500 m in depth. The unmineralized breccias, however, extend some 900 m southwest from Paroistenjärvi. Tourmaline breccias and tourmaline-bearing veins, moreover, are present at several places near the border of the granodiorite with breccia tongues radiating outward from the igneous body and cross-cutting the marginal granodiorite and the volcanites. At its southwestern end, the breccia zone pinches out into vertical tongues. The breccia occurrence near Paroistenjärvi is the largest and the only one containing economically viable mineralization. Himmi and his colleagues think that breccia is of large enough volume to be designated as a distinct rock type. The matrix of the breccia is deep black, setting this rock apart from all others in the area. The breccia fragments range from small pebbles to boulders that are from 20 to 50 cm across; the greater part of them are of tuffite. The various types of breccia are determined by the size of their fragments and may grade gradually from one type to another or may change in a short distance from highly brecciated to only slightly fractured wall rock. Some fragments are porphyrite, these being most abundant when porphyrite is close by. Whereas the porphyrite fragments essentially equal in dimensions, the tuffites are elongated parallel to the bedding. Fragments normally are surrounded by thin alteration bands, mainly quartz with coarser grains on the outside and finer ones on the inner; some chlorite, epidote, hornblende, and plagioclase may be present. A thin, dark-green chlorite seam may lie between the quartz and the fragment. Some smaller fragments may be completely altered (Himmi, and others, 1979).

The matrix of the breccias is composed of tourmaline with various quantities of ore and other minerals; the tourmaline needles are in small radial clusters, and coarse grains are uncommon. In the core of a tourmaline "sun", there may be a grain of quartz or apatite; apatite in small grains also is sparsely included in the tourmaline. Some grains of uraninite and thucolite (having radiation haloes) are found. In the mill feed, tourmaline ran 13 to 16 per cent of the ore, whereas the P_2O_5 content was from 0.13 to 0.25 per cent; this is much less than the P_2O_5 content in the tuffites and porphyrites. Clark (1965) believed that the hydroxyl-rich apatites crystallized between 600° and 500°C and under pressures of 2000 to 3000 bars

Although large quartz augen are rare in the tourmaline matrix, quartz is quite common in tourmaline veins beyond the breccia bodies; the small nests of chlorite in the matrix near its margins probably are altered fragments of mafic tuffite.

Minor dikes and veins cut the tourmaline breccia, but they are most uncommon; some of them contain chalcopyrite, pyrite, galena, and sphalerite.

The sulfide deposits of ore grade are at the northeastern end of the main breccia zone where they form four subparallel, vertical ore zones. The boundaries of these zones are gradational with the uneconomic breccias; the four zones are quite similar, the one to the others. Ore zone I was the most important economically, following the southeastern boundary of the breccia and extending from the surface to 330 m below it; it had a horizontal length of 190 m and its average width was 14 m. In this zone, the breccia fragments were fine-grained and much silicified, and the tourmaline matrix was the more abundant than that in the other three ore zones. This zone also had the highest scheelite content. Zone II was more irregular, pinching out entirely in several places between the 65 and 285 m levels. Zone II joins zone I in several places, and the two zones are quite similar in composition. Zone III reaches from the surface down to 365 m below it; it has an average length of 95 m and its width averages 12 m. Zone IV follows the northwestern border of the breccia and plunges down from the surface at 75° to 250 m where it joins zone III. Zone IV is 110 m long and 8 m wide. Between 285 and 365 m, there is another ore body that is in a corresponding position to that of IV; this zone is 85 m long and 26 m wide. This lower zone is located in a swell in the tourmaline breccia. Zone IV is not as much brecciated as the other three. It contains much arsenopyrite, especially along the northwestern margin of the zone (Himmi, and others).

No breccia ore is known below 380 m; at this depth, the rock grades into an agglomerated tuffite that is only slightly brecciated and includes ore minerals only in local spots. The average composition of 2.5 million tons of ore mined between 1958 and 1965 was as follows (Himmi and colleagues, 1979)

Constitutent	Percentage	Constitutent	Percentage
SiO_2	51.54	S	1.34
TiO_2	0.96	F	0.09
		As	0.46
Al_2O_3	14.77	Cu	0.68
Fe	13.17	WO_3	0.04
		Co	0.02
MnO	0.17	Zn	0.017
MgO	2.91	Sn	0.013
		Mo	0.012
CaO	1.76	Ni	0.004
B_2O_3	1.34	Pb	0.0036
		Bi	0.001
P_2O_5	0.19	Ag	17.0 g/metric ton
		Se	12.0 g/metric ton
Total	86.81	Au	0.13 g/metric ton

The Ylöjärvi ores had two distinct mineral assemblages: (1) arsenopyrite that was found in the chloritized portions of the wall rocks and in silicified and chloritized breccia fragments and (2) the chalcopyrite-arsenopyrite-pyrrhotite assemblage localized in breccia matrix. Seven stages of fracturing separated the main (Cu-As) mineralization into eight stages. The scheelite was confined to the first of these stages; arsenopyrite was present only in the first stage. Minor magnetite, pyrite, molybdenite, and rutile accompanied this first stage as probably did minor cassiterite and native gold. The second stage contained magnetite, pyrite, ilmenite, and monlybdenite. All of these but the pyrite extended into the third stage where they were joined by tourmaline, minor apatite, and vanishingly minor uraninite. The fourth stage included chalcopyrite, pyrrhotite, and minor quartz, cubanite, sphalerite, and stannite. Stage 5 contained quartz that continued on through the sixth stage; 5 also included galena, matildite, native bismuth, and native silver and perhaps tetrahedrite, bournonite, boulangerite, bismuthinite, and löllingite as very minor minerals. The sixth stage added more, but minor, native silver and considerable pyrite, marcasite, troilite, and pyrrhotite; it also included minor siderite and ankerite, and possibly mackinawite. The sixth stage was confined to minor pyrite, siderite, and ankerite, and the seventh and eighth stages had only pyrite.

The arsenopyrite is the principal sulfide mineral in the tourmaline breccias; the other sulfides are confined to the northeastern part of the main breccia zone but FeAsS is throughout the breccia zone as well as in tourmaline veins and patches outside it. Arsenopyrite rarely has inclusions; if it does they are silicates, ilmenite, magnetite, or molybdenite; arsenopyrite is almost never altered but may be fractured or slightly corroded.

Chalcopyrite is economically the most important mineral in the breccia zones. Fine-grained chalcopyrite occurs as small grains in small-fragment breccia or it fills spaces between, or cracks in, tourmaline needles. Coarse chalcopyrite in places shows lamellar twinning and usually encloses inclusions and exsolution bodies. Sphalerite is minor in size and amount; sphalerite inclusions normally contain tiny exsolutiion bodies of chalcopyrite. Pyrrhotite is closely associated with chalcopyrite in space, but it is found only in a few places as inclusions in the copper mineral. Cubanite is present in chalcopyrite only as exsolution lamellae or irregular bodies enclosed in the chalcopyrite. Pyrrhotite may contain idiomorphic inclusions of arsenopyrite and locally exsolution bodies of pyrite. Native bismuth occurs in droplet-shaped grains that may result from the bismuth having been separated from the ore fluid in the molten state.

Although pyrrhotite is not as common as chalcopyrite, it is present in much the same places as the copper mineral although its distribution is more irregular, and it appears most likely to be in tourmaline-poor matrix of unsilicified breccia. Mackinawite, although rare, is rather widely distributed in the ore; it may be in cracks in, and around the margins of, grains of magnetite where the magnetite grains are enclosed by younger chalcopyrite, cubanite, and pyrrhotite. Pyrite is rare but is commonly present both by itself and with other minerals; where it is in separate grains, these normally are octahedra, but it also can be anhedral. Marcasite usually is an alteration product of pyrrhotite.

Sphalerite, stannite, galena, and molybdenite are scattered throughout the deposit, in places alone, in others closely associated together. Sphalerite is, in the main, in, or close to, chalcopyrite, and most of the ZnS is as inclusions in the copper mineral. These inclusions of ZnS in chalcopyrite exhibit tiny inclusions of chalcopyrite; this entire association of the two sulfides indicates that the chalcopyrite was deposited with a modest amount of ZnS is solid solution in it and that much of it came out of solution in the $FeCuS_2$at a time when some $FeCuS_2$ still could be held in solution in the exsolved sphalerite. On the basis of Clark's work, Himmi and his colleagues suggest that the tin content of the copper concentrate, low though it is, indicates that some tin is in solid solution in the chalcopyrite.

Scheelite is with molybdenite but with some of the molybdenum being incorporated in (MoO_4^{-2}) radical and that radical being present in the scheelite as powellite. The distribution of the scheelite in the ore is random with the crystals of that mineral being grouped as bands or by forming larger groups of small patches of $CaWO_4$ crystals. Scheelite is concentrated along the margins of the breccia bodies; this occurs mainly in the southeastern corner of zone I, and the scheelite normally occurs as idiomorphic crystals. The molybdenum content of the scheelite averages 0.04 per cent. Ilmenite, magnetite, and cassiterite are present in the tourmaline matrix as small grains that, in many places, are as inclusions in chalcopyrite and pyrrhotite. Uraninite occurs as inclusions in tourmaline, apatite, molybdenite, and even in arsenopyrite. In tourmaline, the uraninite inclusions are enclosed in an alteration zone that contains torbernite and parsonsite, radially oriented. In turn, this alteration halo is surrounded by a distinctly marked radiation halo that includes locally tiny crystals of galena and native selenium.

Silver is present in the entire mineralized area of the breccia but is minor in amount everywhere but with its greatest abundance between the 150 and 300 m levels. Gold is enriched above normal background in zone IV along its northwest boundary where the breccia is rich in arsenopyrite. This is simply another example of ores in which arsenopyrite and gold have a direct relationship.

The silver inclusions almost everywhere are associated with chalcopyrite, but most of them are too small to be seen under the optical microscope. The content of silver in the copper concentrate is 10 times greater than in that of arsenopyrite, but the gold is about equally divided between the two.

Himmi and his colleagues (1979) studied a total of 378 analyses taken along a NE-SW section that cuts representative parts of the ore bodies and the breccia. Their results show that the entire breccia is anomalously high in copper as is breccia plus tuffite and agglomerate (5,462 ppm and 4,210 ppm, respectively); where lower quantities are present in the breccia, this normally is due to the large size of the fragments. The unbrecciated porphyrite and agglomerate, however, are very low in copper. The tuffite contains about 35 per cent of the copper of the breccia but about 45 per cent of that in the breccia, plus agglomerate and tuff. The copper content is not uniform in the breccia since the western part of the breccia is low in copper even though also low in fragments. The maximum silver content is centrally located in the breccia along its eastern contact; this maximum decreases both upward and particularly downward. The high-arsenic grades are confined to the breccia with its maxima in the upper parts, in the east, in the western margins of the breccia, and in the base tongues. Not surprisingly, the distribution of sulfur parallels that of copper and silver; the sulfur maximum coincides with that of silver. Tungsten distribution is quite different from that of the copper (and all other metals), being most abundant along the eastern contact of the breccia, plus other minor highs in the narrow two-branched zone in the center of the breccia.

Factor analysis (R-mode) indicates, among other things, that the mineralization began with the crystallization of arsenopyrite and scheelite, followed by the deposition of molybdenite and, slightly later, by apatite. After the early mineralization, that of copper and its associated minor and trace elements commenced. This sequence is much the same as that described previously and is based on Clark's (1965) microscopic work.

Radiometric dating on various minerals gives age dates for the Ylöjärvi ores as ranging between 1810 and 1890 m.y.; an average of about 1850 m.y. seems close enough for dating purposes in this discussion. Zircons from the Hämeenkyrö granodiorite average about 1840 m.y. This agreement in dates would seem to establish not only the time at which the ores were emplaced in the tourmaline breccias but also that a genetic connection probably existed between the granodiorite magma and the Ylöjärvi ore deposit.

Himmi and his colleagues (1979) summarize the formative processes of the Ylöjärvi ore deposit as follows. The breccia was produced at the margins of the Hämeenkyrö granodiorite massif in folded volcanics (tuffs and agglomerates) that had been at that time as much metamorphosed as they ever were to be. Additional though small, bodies of breccia also are associated with the granodiorite. The breccia is bordered, adjacent to the wall rock, by an irregular fracture surface, and only the southeastern margin is emphasized by a shear zone; the walls of the breccias are not mylonized or sheared except for the shear just mentioned. The mineralization of the breccia was almost entirely by the filling of open space; fragments were not appreciably replaced. The close association in time and space of the breccia, its ore mineralization, and the granodiorite indicate that all are of essentially the same age. These authors say that the maximum temperature of the ore-forming process cannot now be estimated but that it probably was rather high and lasted only a short time; this is suggested by the medium thickness of the sharply defined and totally silicified layer on fragments and by the lack of alteration of the cores of the fragments and of the walls of the breccias.

They suggest that brecciation, ore formation, and minor alteration all might be explained by an explosive discharge of hot gases. After the discharge had ceased, they believe, the feeder channel was blocked, new pressures built up and a second explosion occurred; it produced less fragmentation nor did silicification take place. [This explanation does not stand rigorous examination as the tourmaline, the ore minerals, and such alteration as did take place, must have been produced by reactions involving water-rich solutions. These solutions almost certainly were above the critical temperature of water and perhaps even above the critical temperature of the ore fluid. How much higher than the critical temperature of water that of the fluid may have been is, of course, unknown. At any event, the paragenetic sequence almost certainly could not have been produced by a rapid precipitation of materials from a tenous gas.]

The Ylöjärvi deposit is unique, not only in Finland but also in Fennoscandia. Himmi and his colleagues point to certain resemblances to such Tertiary tourmaline-breccia pipes as Braden (El Teniente) in Chile and Ilkwang in Korea. The major difference between Ylöjärvi and Braden and Ilkwang is in age, Precambrian as opposed to Tertiary; also Ylöjärvi is less round and more plate-like in form and does not have sheeted fractures along the wall rocks. They think, however, that the mechanism of formation of all three ore bodies mentioned could have been essentially the same.

Anyone reading this summarized discussion will immediately recognize that it is a shortened version of the 1979 paper by Himmi and his colleagues. The long association of such a distinguished geologist as Dr Himmi with Ylöjärvi guarantees the accuracy and completeness of this report of this most interesting and unusual deposit.

The deposit almost certainly was emplaced at high temperatures and at least high enough pressures to be classified as hypothermal or xenothermal. I would suspect that the resealing of the brecciated area was so swift and strong that the pressure that could build up during ore deposition was sufficiently high for hypothermal to be the better choice of the two categories that might be applicable. As no carbonate rocks are present in the area, except for the few and minor carbonate veinlets produced as part of the ore-forming process, the classification of the deposit must be Hypothermal-1 (this is, in non-carbonate rocks).

POLAND

FORE-SUDETIC COPPER

Middle Mesozoic	Copper, Lead, Zinc Silver	Leptothermal to Telethermal

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Notes

The Fore-Sudetic copper district has its administrative and operational center in the town of Lubin (51°23'N,16°10'E) that is about 60 km northwest of Wroclaw (formerly known as Breslau). The deposit is the richest that has, to date, been found in the Permian Zechstein formation; its outcrop length is at least 15 km and the mineralized area encompasses no less than 400 km^2. At the present time, three mines are in operation and the ground for a fourth is being explored. The oldest of these mines is at Lubin; the second is at Polkowice, about 16 km northwest of Lubin; the third is Rudna, some 14 km ENE of Polkowice and about the same distance NNE of Lubin; and the fourth area is in the vicinity of Sieroszowice and is approximately 9 km WNW of Polkowice and 22 km almost due west of Rudna. The width of the mineralized belt averages about 14 km, although, of course, not all of it contains ore-grade material. Although all the present mines produce copper (and perhaps some silver) only, a considerable portion of the belt contains lead-zinc-bearing shale; this lead-zinc area is in an irregularly shaped area between Lubin and Rudna with a minor extension northwest almost as far as Zofiowka, about 5 km northeast of Sieroszowice. The copper ore ranges in grade from 12 per cent in copper-bearing shale to 1 to 2 per cent in sandstone. The ore is sandstone, however, is much more amenable to concentration than the ore in shale, so a much lower grade can be mined from the sandstone. The ore is appreciably different than that in the Mansfeld area of East Germany (DRD); the old mines in the North-Sudetic area are more like those of Mansfeld than are the Lubin mines.

The deposits of Lubin district are located in the Fore-Sudetic monocline; this major structure and the Fore-Sudetic trough to the southwest of it form the northern foreland of the Sudety Mountains (the Fore-Sudetic ridge). Further southwest of the ridge is the North-Sudetic trough. All of these structures are underlain by a crystalline basement composed of Precambrian and lower Paleozoic schists, amphibolites, and gneisses with the Precambrian portion having been intensely folded and metamorphosed in the Assyntian orogeny at the end of the Precambrian; the Paleozoic rocks were intensely folded and faulted in the Sudetic orogeny although upper Westphalian sediments (at least farther west in Europe) were not affected by these movements. The basement may have been affected by minor movements during the Asturian orogeny late in the Carboniferous to produce some thrust faulting. Overlying the basement is the monocline; the first rocks are those of the probably lower Permian Rotliegende (sandstone), followed by the upper Permian Zechstein and by the Triassic Bundsandstein, Muschelkalk, and Keuper and then by Upper Cretaceous and Cenozoic beds. The ore bodies in the Fore-Sudetic monocline in the Lubin region are confined to the uppermost of the Rotliegende (perhaps bleached, reworked, and redeposited to produce the Grauliegende or Weissligende sandstones) and to the basal Zechstein. The southwest border of the ore coincides with the sub-Miocene outcrops of the Zechstein.

The Rotliegende is a thick (250-300 m) series of continental quartz sandstones (with some feldspar fragments) with a clay or clay-carbonate matrix and contains intercalations of conglomerates and slates and some examples of volcanic rocks. The Grauliegende ranges in thickness from a few to 40 m with its composition, grain size, and matrix generally the same as those of the red sandstone. The change in color is thought by Banas (and predecessors) to have resulted from in situ leaching of the Rotliegende by reducing solutions percolating downward from the Zechstein beds. Thus, this process must have occurred at a considerable time after the red sandstone had been deposited. On the contrary, the suggestion also has been made that the reworking and redeposition took place in a reducing environment prior to Zechstein deposition. The color of the gray sandstone changes at depth gradually into that of the red. Banas reports (1980) that recent investigations of the gray sandstone show that the grain size is fine to medium, that is has a psammitic texture and a random structure. Carbonates are present in the upper part of the gray sandstone, and the uppermost layer (about 10 to 20 cm thick) of that sandstone is cemented by dolomite and clay minerals and minor calcite. The main minerals of this matrix are illite and dolomite with minor quantities of montmorillonite, kaolinite, calcite, and chlorite; locally anhydrite, quartz, chalcedony, gypsum, and sulfides, plus clay-organic-matter cement are present. The sedimentary structures present in the Grauliegende are typical of shallow-water deposition. Rotliegende is the oldest of the rocks of the Permo-Triassic sequence in the Fore-Sudetic copper district.

Above the Grauliegende is the Zechstein, and this is followed by the usual order of Triassic sediments - Muschelkalk, Keuper, and Rhaetic. Unconformably above the Triassic beds are (perhaps) some Lias and then Cretaceous rocks. No middle or upper Jurassic rocks are known in the area. Thus, the Cretaceous is unconformable on the Triassic or (where it may be present) on the Lias. In turn, the Cenozoic beds are unconformable on the Cretaceous.

The ore mineralization in the Lubin district begins in the Grauliegende of the Upper Rotliegende. The thickness of the mineralized portion of the Grauliegende ranges rather widely as does the sulfide content of that sandstone. Certainly not all of the Grauliegende is mineable ore. The upper 0.5 m or so of the Grauligende is a dolomitic sandstone (also gray). Immediately above the dolomitic sandstone, in most places, is a thin layer (up to 20 cm) of dolomite, known as the Boundary dolomite. It is considered to be the first of the Zechstein evaporite beds; although it is mainly dolomite, it also contains minor calcite, illite, terrigenous and authigenic quartz, sulfides, organic matter, and gypsum. Above the Boundary dolomite is the Copper-Bearing shale, a continuous bed with an average thickness of 30 to 50 cm. The basal contact of this shale is either with the Gray sandstone or with the Boundary dolomite (if this latter rock is present). The Copper-Bearing shale is made up mainly of clay minerals (illite, mixed-layer illite-montmorillonite, montmorillonite, and minor kaolinite and chlorite), carbonates (dolomite and locally calcite), organic matter (about 90 per cent kerogen and some 10 per cent soluble hydrocarbons), terrigenous and authigenic quartz, sulfides, and locally gypsum. Bedding and lamination are common in this rock with microlamination being a characteristic feature. These microlaminae contain rhythmically repeated dolomite and clay microlayers. The lower the position of the Copper-Bearing shale, the higher the content of clay and organic material; this part of the shale is known as "smearing shale" for obvious reasons. The upper part of this shale passes gradually into clayey dolomite; the lamination in this clayey dolomite is emphasized by linearly arranged sulfide grains and aggregates and by calcite and authigenic quartz.

In a considerable portion of the ore-containing rectangle, the Copper-Bearing shale may be Lead-Bearing, Zinc-Bearing, or Polymetallic shale (mainly a mixture of galena and sphalerite in which the galena is about three times as abundant as the zinc sulfide). These varieties of shale are confined to the area already mentioned as being concentrated in the southeast portion of the ore rectangle and lying largely between the Lubin and Rudna mines.

The best ore mineralization is concentrated, on the average, in the upper meter or less of the Grauliegende and in less than a meter of shale ore immediately above the Gray sandstone. In places, however, lower-grade ore extends downward into the Gray sandstone about 2.5 m, and less than ore-grade material is even wider-spread than actual ore. Above the shale ore in the lowermost Zechstein, the next 2.0 to 2.5 m is known as dolomite ore. The lowest of the rock types containing this dolomite ore is about 0.5 m of dark-gray clayey dolomite; this is followed by less than a meter of gray streaky dolomite. The next 0.75 m of the overlying dolomite and limestone also normally is mineralized. This dolomite ore may be thicker or thinner than the 2.0 to 2.5 m mentioned above. The average total thickness of the mineralized zone, however, is some 5.0 to 5.5 m.

The total thickness of the carbonate rocks in the Fore-Sudetic copper district ranges from 7 to 10 m in the northwest part of the district but may be as much as 60 to 80 m in the southeast portion. The lower part of this carbonate sequence is dolomite and the upper limestone. Fossils are abundant with foraminifera and molusca being most common; these fossils in places are replaced by sulfides and/or anhydrite. Some joints and bedding planes in these carbonate rocks are filled with calcite, gypsum, barite, anhydrite, or sulfides. Nodules in these rocks are widely distributed and are filled with coarsely crystalline anhydrite plus some calcite, barite, and sulfides. Fluorite has been found in stylolites.

The boundaries between ore and waste are not sharp, the contacts being gradational both at the top and bottom of the ore horizon in any area. The highest concentrations of ore minerals occur in the Copper-Bearing shale, but locally ore is quite high grade in the uppermost part of the Grauliegende, and locally farther down in that rock.

Banas (1980) reports that microscopic studies show that the sulfides have a variety of textures. Typical in the sandstones are impregnations and disseminations. The uppermost of the Grauligende (dolomitic) sandstone contains it sulfides as spots and grain cements. In the shale streaks, the dominant textures are lenses and veinlets, whereas in the dolomite ore (above the shale ore) disseminations and nests are most common.

In 1980, 70 minerals (mainly sulfides and sulfosalts) are found in the ore. The most abundant sulfides are those that contain more or less Cu^{+1}; these are chalcocite, digenite [$Cu^1_8Cu^2S_5$], djurleite [$Cu^1_{30}Cu^2S_{16}$], and anilite [$Cu^1_6Cu^2S_4$]. Covellite, which of course does not contain any Cu^{+1}, also is present in the ore, but it is of minor importance, as are djurelite and anilite. The iron-copper sulfides present are bornite [$FeCu^1_4Cu^2S_5$] and chalcopyrite [$FeCuS_2$ (approx.)]; pyrite also is present and normally is accompanied by galena and sphalerite, particularly in the Lead-Zinc-Bearing shales. Essentially all of the sulfides are xenomorphic against the sedimentary minerals by which they are enclosed. The sulfides, therefore, were introduced after the gangue minerals and thereby were forced to accomodate themselves to their surroundings and could not develop rational faces. On the contrary, the rare nickel and cobalt minerals and the uncommon sulfosalts are idiomorphic. In some instances, digenite, chalcocite, digenite, bornite, galena, and stromeyerite exhibit framboidal textures. These are considered by all Polish geologists to indicate that they were precipitated in the colloidal state. It is possible, however, that these textures simply mean that the minerals were deposited rapidly around relatively numerous centers of nucleation. Banas reports that recrystallization textures are shown typically by bornite, chalcocite, galena, sphalerite, and chalcopyrite. Just how textures are identified, Banas does not say. He further states that replacement textures are exhibited by the major sulfides in connection with zonal, cementing, atol, and cataclastic textures.

Chalcocite is the most abundant of the ore minerals, making up about 60 per cent of the total sulfide content; it may be digenite (Oberc and Serkies, 1968) but this is uncertain. In the carbonate-rich shales, bornite is the most common sulfide, and it is next in total importance to chalcocite; chalcopyrite is present in all rock types in amounts ranging from 10 to 2 per cent of all ore minerals. Covellite and digenite are no more than trace minerals (Oberc and Serkies, 1968). Pyrite is present with the copper minerals but in no more than minor amounts. Except for the northeast area of lead and zinc ore, sphalerite and galena are rare minerals, although locally lead-zinc mineralization lies above the copper ore. Average grade of copper in shales is 11 per cent (6% Cu is shales is regarded as too low-grade to mine), grade in sandstone 1 to 2.

The economically valuable ore ranges considerably in vertical thickness, in some areas being too thin to mine economically.

Banas (1980) sees the distribution of ore minerals somewhat differently. He recognized horizontal zoning with the southeast portion of the copper-ore-bearing section containing a bornite-chalcopyrite association; in the central part, the main minerals are chalcocite with some digenite, covellite, and djurleite. Banas describes vertical zoning with, in both the top and bottom parts of the ore zone, chalcopyrite and pyrite being high and minor marcasite being present. In the middle of the ore zone, the dominant minerals copper-sulfur and copper-iron-sulfur with copper increasing as iron becomes less abundant. Above the copper horizons, lead-zinc ore occurs with galena being more abundant in the lower part of this ore and zinc being so in the higher portions. In places, lead-zinc ore and copper ore are intermixed, and in such places, tennantite is a common mineral together with sphalerite, chalcopyrite, and pyrite.

Haranczyk (1970) envisions the depositional environment of the Fore-Sudetic deposits to have consisted largely of an isolated lagoon to the northeast (Rudna lagoon and an euxinic sea to the southwest. The barrier between euxinic sea and lagoon) was composed of sand and ran irregularly northwest to southeast and was divided by sub-barriers, particularly in the southeast portion of the ore-bearing area. The major overflow path from lagoon to sea was located toward the northwest end of the lagoon which accounts for the location of the ore-bearing shales in greater quantity in the southeast portion of the Lubin district.

Haranczyk prepared a "lithologic redox map" of the Lubin area on the basis of core samples from some 300 drill holes drilled with a 1.5 km^2 grid with two of its sides oriented about parallel to the strike of the Zechstein beds, that is northwest-southeast. Haranczyk chose these samples so that they would represent the whole of the lithologically homogeneous part of the lower Zechstein. The ratios he used in contouring this map were: (1) $S^{+6}/S^{-2} > 1$: (2) $S^{+6}/S^{-2} > 0.25$; and (3) $S^{+6}/S^{-2} < 0.25$. The sediments were divided into three classes by using the coefficients 1 and 0.25 as limiting values. The value of "1" corresponds to conditions under which sediments, showing equal activity of the various oxidation grades of sulfur, give potentials equal to "1". The value of "0.25" was chosen as a model value to distinguish euxinic sediments from carbonate rocks. The redox map in Haranczyk's 1970 paper shows two compact areas in which the redox potentials are much higher than 1. One of these areas irregularly encircles Sieroszowice where the unmineralized sediments are red-strained. The second of these areas extends northeast of a line lying north of the lagoonal barrier and is of major proportions. In this second area, despite the high redox coefficient, no red oxidation stains are observed. In addition to these two major areas of high coefficients, three small isolated areas showing similar high values are present; (1) just south of Zofiowka (a location 4.5 km northeast of Sieroszowice); (2) an even smaller area less than 2 km southwest of Polkowice; and (3) the largest of the three areas that is slightly over 3 km south of Polkowice. Lead-Bearing shales are concentrated largely northeast of the sand-barrier between euxinic sea and lagoon but southwest of the sediments of the Rudna lagoon. The Copper-Bearing shales are located irregularly but mainly in the region from northwest of Zofiaka and Polkowice, that is, northwest and west of the Lead-Bearing shales. The boundary of the Lead-Bearing shales is located where drill-holes did not intersect either Zechstein Copper- or Lead-Bearing shales. The belt of metal-bearing shales delimits the zone of maximum thickness of mineralized sandstone underlying the Zechstein carbonate rocks.

At the time Haranczyk's paper was written (1970), the northeast extension of the lagoonal beds had not been explored by drill-holes since, in that area, the bottom of the Zechstein beds is below a depth of 1200 m. Several drill-holes to the northwest of the lead-bearing area, however, encountered typical Copper-Bearing shales, so the northwest boundary of the lagoon may be farther in that direction than Haranczyk's map indicates.

Stratigraphically, Copper-Bearing shales are of two varieties: (1) a bottom layer of typical, black foliated bituminous shale, nearly completely lacking in carbonate - it is normal "smearing" shale and is richly mineralized with microscopic sulfides and (2) a poorly mineralized black, platy material that is highly fissile because it consists of alternating layers of carbonate and clay. These Copper-Bearing shales grade upward into black limestones and dolomites that contain scarse fossil detritus; above these carbonate beds are middle Zechstein stylolithic limestones and dolomites, several meters in thickness. No Copper-Bearing shales have been found in the barrier belt; they always are on the lagoonal side of that belt. In the lagoon, near the barrier belt, slightly mineralized sandstone is overlain by up to a few centimeters of Lead-Bearing shales. Farther away from the barrier, the Lead-Bearing shales are underlain by a thin bed of Copper-Bearing shales. The upper part of the Lead-Bearing shales also is rich in zinc; upward, these layers pass into dense, dark-gray, lead-bearing marl and, above these, are crystalline or dense dolomite that, at higher levels, contain nodules of calcium sulfate.

In the central part of the lagoon, basal Zechstein rocks are cryptocrystalline dolomite with numerous nodules of calcium sulfate that, in places, are marginally encrusted by sulfides. At the contact of these rocks with the Rotliegende beneath is a thin layer of Copper-Bearing shales that contain fossils of the same genus as are present in the organodetrital dolomite of the barrier belt.

That the epicontinental Zechstein sea entered the area quietly is shown by the lack of interformational conglomerates, which, however, are known farther to the southeast. The development of a stagnant bottom zone in this Zechstein sea produced a euxinic basin, and deposition of Copper-Bearing shale began throughout that area. A partial regression of this Zechstein sea resulted in the emergence of a bottom ridge to form the barrier-belt between euxinic sea and lagoon. Although the metal-bearing shales formed along the marginal and more elevated part of the lagoon nearest the barrier, calcium sulfates formed in the more saline, slightly deeper part of the lagoon. Periodic or continuous outflow from the lagoon appears to have taken place at least from an outlet in the northwest part of the barrier. Later, the barrier-bar was submerged and unified the salinity and chemistry of the euxinic sea and the lagoonal basin. Around this submarine ridge, organic life of the biohermal type developed; in this area, organogenic dolomites were formed, but, in the adjacent sea on either side of the ridge, normal dolomites were developed that contained, however, smaller forms of those that formed the fossils of the biohermal submarine ridge.

Considerable attention has been given to the presence of anhydrite sandstone in the Grauliegende where it is developed locally in the upper portions of that gray sandstone. These calcium-sulfate irregular lenses or layers are antipathetic to the Copper-Bearing shales, but typically are present with the anhydrite organic layers at the top of the sandstone and in the bottom of the overlying dolomite. The anhydrite-sandstone bodies are bordered by broad transitional zones that consist of gray, copper-bearing sandstone with spots of anhydrite with the latter diminishing in quantity away from the massive anhydrite sandstone.

The anhydrite sandstone petrographically is a light-gray, compact hard rock, composed of terrigenous grains of quartz and feldspar, cemented by anhydrite. In the centers of the bodies, the anhydrite makes up 60 per cent of the total volume of the rock; toward the margins, this content decreases to 50 to 30 per cent in the transitional zones. In addition to acting as a cement, the anhydrite also replaces terrigenous grains. The cement also includes dolomite, calcite, and clay minerals. In the typical gray sandstone (without anhydrite), the sulfide mineralization is of the copper-sulfur and copper-iron-sulfur types, but the anhydrite sandstone includes pyrite, galena, and sphalerite with minor quantities of marcasite, the Cu-S minerals, and the Cu-Fe-S minerals.

Banas (1980) considers that the relation of the anhydrite cement to the terrigenous material points to its late diagenetic development which he thinks possibly is connected with the dolomitization of the lower Zechstein carbonates and the infiltration of connate brines into the porous, underlying sandstone. In places where metal-bearing shales beds lie at the base of the Zechstein, the descent of such brines would be prevented or inhibited, leaving the Grauliegende beneath free of anhydrite.

In the Fore-Sudetic monocline, the most important structure is, of course, the monocline itself. It appears that this structure was formed during the Laramide orogeny and was accompanied by appreciable faulting, the most important direction of which was WNW-ESE but to which east-west and north-south fractures also belong. The major result of this faulting was the production of horst and graben structures which are characteristics of Fore-Sudetic tectonics. The Triassic series of beds dips northeast under the younger Mesozoic formations of the Szczecin-Miechow synclinorium (also produced during the Laramide orogeny). The Fore-Sudetic monocline, therefore, belongs to the Larmaide stage and is an area of intense Larmaide movements. The Laramide phase produced shallow brachyanticlines and brachsynclines. Some megascopic drag folds were formed on northwest-striking axes within the Copper-Bearing shales, particularly in the "smearing" shales as were some tectonic grooves on bedding-plane surfaces in a NNE lineation (Oberc and Serkies, 1968).

The major fault developed in the Laramide stage was the Middle Odra fault that separates the rocks of the crystalline basement from those of the Fore-Sudetic monocline; the Middle Odra belongs to the WNW-ESE system, a system that is intersected by an east-west-striking system. The third fault system, the north-south one, does not intersect the two just mentioned. The effect of these intersecting faults was to produce the block (horst and graben) structure mentioned previously.

The joints and fissures in the area are principally pre-Laramide and form two main joint systems: (1) WNW-ESE and NNE-SSW and (2) SE-NW and ENE-WSW. Additional joint and fissure systems are present in the carbonate rocks, and ore veins are associated with these; these ore veins are intersected by Laramide slip zones that are parallel with the bedding plane surfaces.

The youngest structures were produced by the Alpine orogeny and affect the dislocated Laramide rocks horizontally; the units of this group have been folded locally by the movement of the continental ice sheet during the Pleistocene (Oberc and Serkies, 1968).

The copper mineralization of ore grade is located mainly at the bottom of the Zechstein and the top of the Grauliegende. The principal host rocks are: (1) the upper part of the Gray sandstone, (2) the Boundary dolomite, (3) the Copper-Bearing shale, and (4) the bottom of the immediately overlying Zechstein dolomite beds. The copper-bearing horizon does not occupy the same stratigraphic position throughout the district; instead, in the Lubin region (SE portion), it is mainly in sandstone and shale; in the Polkowice region (middle portion) it is contained in both shale and dolomite, and in the Rudna region (to the northeast), the most productive layer is the Gray sandstone. Thus, the ore horizon is stratigraphically lowest in the Rudna region, somewhat higher in the Lubin area, and highest of the three in the Polkowice district, that is, the ore horizon rises from southeast to northwest.

Presumably, though this has not yet been (1980) thoroughly tested, the ore-bearing horizon should be highest in the Sieroszowice area.

The only other element of possibly economic value in the copper ores of the Lubin district is silver. This element is concentrated principally in the shale where it is present in amounts up to 2000 ppm. In the other host rocks of copper ore it amounts to no more than background quantities. Most of it is contained in the Cu-Fe-S minerals with as much as 15 per cent of the bornite being composed of silver although 2 per cent is much more common. Smaller amounts of silver are found in chalcocite, djurleite, digenite, and galena (0.x per cent to several per cent). Silver also occurs in electrum, the palladium arsenides, and thucolite. Of course, most of the silver, granted it is being recovered, must come from the bornite and chalcocite as these are the only silver-containing minerals present is large enough quantities to be potentially economic sources of the silver.

Lead and zinc (almost entirely in galena and sphalerite) occur in the copper-bearing ores, but the quantities are so small of these two minerals in this ore-type that it does not appear to be economic to try to recover them in the processing of copper ore. In time (if not at present), the Lead-Bearing shales probably will be mined, at which time the availability of lead and zinc in economically recoverable amounts will rise dramatically.

At least 20 other elements are present locally in more than background amounts, but none of these can be considered as of economic value. For example, nickel and cobalt are present in pyrite (2% Ni and 7% Co) and, to a lesser extent in chalcocite, covellite, and sphalerite - again of no value.

Organic substances are a main constitutent of shale, ranging up to 15 per cent; in other host rocks of ore, the organic matter is negligible. Banas (1980) reports that Haranczyk (1972) has suggested that the material was of major importance in reducing sulfate-sulfur to S^{-2}. The reduction process may utilize the hydrogen from hydrocarbons or oxidation may convert the hydrocarbons to carbonyl groups or to CO_2. Just how these processes worked is not made clear.

The concensus of opinion among the ore geologists who have studied the copper deposits of the Lubin district is that they were formed in three stages. The first of these was the provision on the heavy metals contained in these deposits from the surrounding land surfaces to the Zechstein sea in notable abundance over a geologically short period of time. Haranczyk (1972) considers that these metals were extracted from this sea water by biogenic processes as is evidenced by the strong correlation between organic carbon (in the Copper-Bearing shales primarily) and sulfide-sulfur content in these same sediments. The occurrence of billions of examples of mineralized bacteria of the genus Desulfovibrio desulfuricans in the shales allows, Haranczyk thinks, the assumption to be made that processes on which these bacteria depended for their sources of oxygen provided the S^{-2} needed for the formation of the various sulfides of the heavy metals. This precipitation was the second stage in the formation of the Fore-Sudetic sulfide ore. Obviously, the dominant minerals of the deposit indicate that the temperature of their formation was low and that the species present severely limited the possible variations in Eh and pH of the waters from which they were precipitated. He also points out that the bacterial reduction process can go forward so far as the sulfide ions produced by them are immediately used up in the precipitation of heavy-metal sulfides. He claims that it is only if the development of sulfide ions appreciably exceeds the amount that can be used in forming metal sulfides that the bacteria will be killed by their own wastes. The development of sulfide ions, he says, could not have involved the oxidation of carbon or the direct correlation between carbon and sulfide content in the shales would not be possible.

This explanation may serve to produce the sulfides in the Copper- (and Lead-)Bearing shales but it does not provide for the presence of appreciable amounts of copper minerals in the sandstones under the shales nor in the dolomites over them. Haranczyk recognizes this and considers that the ore minerals in these underlying and overlying rocks must have been moved there by processes secondary to their original deposition. He interprets the textures exhibited by these "secondary" ores as having been produced at an early stage, probably before the end of consolidation and lithification. These early stage secondary sulfides are in horizontally elongated clusters and stringers. Following these, some generations of vertical veins, composed of anhydrite or barite with some ore minerals were developed. Rarely diagonal veins and stringers intersect the vertical veins and the horizontal stringers and clusters; the vertical and diagonal ore veins cut consolidated and lithified sediments. The horizontal ores consist of simple copper sulfides and calcite. The later-stage fissures (vertical and diagonal) were the result of tectonic processes through which Haranczyk believes that waters from overlying aquifers moved downward into the Copper-Bearing shales and there leached out sulfides and redistributed them into the surrounding formations (sandstone below and dolomite above). The anhydrite in the overlying Zechstein beds might, Haranczyk believes, have been leached from still higher Zechstein beds that contained appreciable amounts of that mineral. This leaching would have produced solutions, in his opinion, capable of dissolving, transporting, and depositing in the lower fissures the minerals now found there. He further thinks that these solutions had much the same oxidation-reduction regimen as prevailed during the deposition of the horizontally arranged clusters and stringers. The downward moving solutions were not rainwater but probably were intrastratal connate waters of high salinity derived from that portion of the Zechstein beds in which anhydrite and various salts were available. He does not think this explanation valid for the secondary copper minerals (such as digenite, covellite, stromeyerite, native silver, and cuprite) that do not occur in association with barite and probably were developed by the action of meteoric waters rich in oxygen.

Several difficulties are apparent in these rather complex explanations for the variety of minerals contained in various positions and structures in the Fore-Sudetic ores. It is not clear why the descending intratratal waters would have been able to extract sulfides from the Copper-Bearing shales and then have moved back with them into the dolomite overlying those shales. Certainly, the dolomite did not originally contain sulfides that could have been redistributed within those beds, so those minerals must have come from below.

It seems much more probable to me that, if the ores are essentially syngenetic in the Gray sandstone, the metallic elements, if not the sulfur, would more logically have been introduced from hydrothermal solutions locally reaching the sea floor during the sedimentation process of the sandstone and Copper-Bearing shale. If this is correct, then the sulfur needed for the sulfides in the sandstone of the upper Rotliegende must have come in the metal-bearing ore fluids though that for the sulfides in the Copper-Bearing shale may have been locally produced by the reduction of sulfate sulfur.

It is, however, also possible that hydrothermal fluids of magmatic origin (as would have been those solutions mentioned in the preceding paragraph) may have entered the volume of now mineralized rock (after all the rocks had been solidified) in various stages in relation to the tectonic fractures to produce the sequence of mineralization now known.

A major argument put forward in favor of the biogenic origin of the sulfur contained in the Fore-Sudetic sulfides is that they are notably depleted in ^{34}S. The delta per mil values of some of the major sulfides show -28.4°/₀₀ for galena; -21.3°/₀₀ for bornite; and -13.7°/₀₀ for chalcocite.

Certainly such delta values are far more negative than those produced from solutions of magmatic genesis that have not had their delta values changed by reactions that took place between the ascending solutions and the rocks that they encountered on their upward journeys. Do these highly negative values indicate without doubt that the sulfide-sulfur in the ore minerals was produced by bacterial reduction of sulfate-sulfur? The delta values of syngenetic pyrite very commonly are similar to those of the Fore-Sudetic ore sulfides; thus, if the ore sulfides were emplaced by the replacement of pyrite, these reactions normally would derive their sulfur from the pyrite they replaced without appreciable change in such values. In the Fore-Sudetic ores, however, although replacement of pyrite does appear to have occurred, most of the sulfides were not emplaced in that manner.

It would be a remarkable coincidence if the bleaching of the upper portion of the Rotliegende (to give the Grauliegende) did not have some connection with the deposition of the sulfides in that rock. Whether the sediments from which the Grauliegende were produced had been lithified when the disseminated sulfides were introduced or not does not matter. Under either variety of condition of the sediments, the original iron material that gave them their red color would be available to react with the ore fluids to oxidize excess sulfur to S^{+6}, a process that would cause the sulfide ions in the disseminated sulfides to be greatly reduced in ^{34}S over what they contained prior to their reaction with iron in the sand sediments or lithified sandstone took place. Similar reactions between sulfide ions and cupric copper also would tend to reduce the proportion of ^{34}S in the sulfides formed in the Grauliegende and also would account for the dominance of cuprous over cupric ions in the sulfides in the Grauliegende.

It also would be a remarkable coincidence if the necessary high concentrations of copper sulfides in the Copper-Bearing shales did not have some genetic connection with the organic content in those shales, organic material that certainly was there prior to the deposition of the sulfides. Thus, any explanation for the presence of abundant copper in the shales and moderate quantities of sulfides in the Gray sandstone beneath them must incorporate the effects of the iron in the sandstone and the organic matter in the shale in the solution to the problem.

Haranczyk (1972) reports that the sulfur content of the Copper-Bearing shales is no more than 0.1 per cent but that the organic-matter content is as much as 15 per cent. This amount of sulfur certainly does not seem enough, unless it was aided by much sulfur brought in from an outside source, to have supplied all the sulfur (as sulfide ion) now contained in the ore sulfides of the Copper-Bearing shales. The suggestion already has been made here that the needed sulfur came down from above, but this would need the help of simultaneously rising metal-rich fluids from below, a coincidence in timing that is unlikely to be acceptable in any model for the formation of these ores. Surely, the metal-poor Zechstein beds could not have been the source of the metals in the Copper-Bearing shales.

Haranczyk (1972) also demonstrated that a positive correlation exists between organic carbon and sulfide ion in this shale. He claims that the presence of hydrocarbons in the shale acted to reduce the sulfur in sulfate ions (brought into the shale in solutions that picked up their sulfur in sulfate ions in the overlying calcium-sulfate-bearing Zechstein evaporites). As evidence of this he shows that kerogen (fossilized insoluble organic material found is sedimentary rocks, usually shales) lenses in some places in the Copper-Bearing shale are covered by sulfide aggregates and carbonates. If Haranczyk's concept is correct, then not only must the sulfide in these shales have been deposited from descending solutions but so must the sulfides in the underlying sandstones. Probably the most reasonable way to apply his suggestion would be that sulfate-rich waters entered the metal-rich carbon layer and there reacted with carbon to produce metal sulfides as the sulfate sulfur was reduced to sulfide-sulfur. Further, it is necessary to assume that appreciable amounts of metal and sulfide ions were moved downward to the Grauliegende and there deposited as disseminated sulfides in that rock. The major problem with this tenative hypothesis why the disseminated sulfides to the Gray sandstone are confined to that body and in no places extend on downward into the Rotliegende.

If, on the contrary, the solutions are thought to have risen through the sandstone underlying the Zechstein beds, carrying their sulfur as sulfide ions, these would react with the iron in the Rotliegende to convert (bleach) the iron from red to gray and to deposit disseminated sulfides there. As these solutions rose into the basal Zechstein (the Boundary dolomite) and on into the ore-bearing shales and as these solutions passed through the carbon-rich shales, some of the sulfate-sulfur produced in the gray sandstone would be reduced to sulfide ion. Further, these solutions also would be adding metal ions to those already present in the shales, thus making possible precipitation of sulfides in much greater amounts than had been possible in the Gray sandstone. Also, the rather wide variety, though small, quantities, of sulfides and sulfoslats not characteristic of the mineralization in the Gray sandstone would have resulted from the impressive variety of trace metals in the shale reacting with sulfide ion produced by the reduction of sulfate ions already present in the rising solutions. As these solutions moved upward, out of the ore-bearing shale, the mineralogy of the sulfides deposited would become quite simple again. This follows because trace metals were not present in the dolomites overlying the shales in anything like their abundance in the ore-shales. The rising ore fluids, however, normally would contain enough metal and sulfur to account for the degree to which the Zechstein dolomites contain copper mineralization.

This proposed mechanism for the emplacement of the ore minerals presumes that the host rocks were lithified at the time when the ore fluids entered them and that the vertical and horizontal fractures now filled with various sulfides, sulfosalts, and gangue minerals existed then. These fractures probably, then, would not be chargeable to Laramide orogeny but to earth movements more nearly the age of the Traissic rocks in which they are contained. Perhaps they should be thought to be of about the same age as those the solutions that entered the Upper Silesian lead-zinc district, that is Middle Mesozoic. No matter from which direction the ore-forming fluids came, they cannot have been at temperatures above the low-moderate range. Thus, in the depositional process, these fluids probably would not have been above the leptothermal range and most of the deposition from them almost certainly was within telethermal limits.

UPPER SILESIA

Middle Mesozoic Zinc, Lead Telethermal

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Notes

The zinc-lead deposits of the Upper Silesian-Cracovian region are located in what is now south-central Poland, with the center of mineralization being about 60 km west-northwest of Krakow (50°03'N,19°55'E). Mining has been carried out from the vicinity of Tarnowskie Gory (50°28'N,18°40'E) and Bytom (50°21'N,18°51 E) at the northwest end of the major syncline to the Trzebionka-Matylda area on the southwest margin of that structure and to the Olkusz-Bolesław area on its northeastern margin. Recently major ore bodies have been found in the Zawierce (50°30'N,19°24'E) and Siewierz areas that are some 20 km north of the Olkusz-Bolesław area but are in the same Triassic Muschelkalk beds as are the other ore bodies of the district.

In the U.S. Bureau of Mines "Mineral Facts and Problems", 1980 edition, the reserves of zinc and lead in Poland as of that date were estimated as 2 million metric tons of contained lead (plus other far less certainly estimated lead-containing material that may include as much as 3 million tons of contained lead) and as 3 million metric tons of contained zinc (plus lead-containing material that may hold as much as 1 million metric tons of contained zinc). Production figures for current operations are, of course, not easy to obtain from or about Polish mine production, but before World War I, 192,000 tons of metallic zinc and 45,000 tons of zinc metal were obtained from the entire district that was then essentially all in Germany. Between the two world wars, production ranged between 85,000 and 170,000 tons of zinc metal per year and between 10,000 and 37,000 tons of metallic lead in the then Polish section of the district and between 42,500 and 95,000 of zinc metal and between just over 8,000 and nearly 19,000 tons of lead metal from the German section. In 1946, less than a year after World War II ceased, the production in the entire district had returned to over 61,000 tons of metallic zinc and to nearly 9,000 tons of metallic lead. Beyond this date, no more than approximate tonnages are available, but, until 1980, the production almost certainly has reached appreciably higher tonnages than in 1946. If the estimates of zinc and lead metal reserves are of the right order of magnitude, these reserves should last well into the 21st century.

The first discoveries of zinc and lead ore (then only of interest for the lead) were made in the Tarnowskie Gory area in 1517 at which time the region was under Polish sovereignty. Between that date and the end of World War II the district changed its political affiliation many times, but after World War II and much movement of peoples in and out of the area, the entire ore-bearing area belongs to Poland. Although the earliest announced purposes of prospecting were for lead and silver, the amount of silver recovered over more than 460 years must have been negligible. At least most samples I have tested ran less than 0.01 per cent in the galena. According to Zwierzycki (1950), total production from the Olkusz area between 1549 and 1668 was 17,000 tons of lead, 10,000 tons of lead oxide, and about 1500 kg of silver. How it was possible to recover this silver economically can only be explained by remembering that the mining industry of those days was labor intensive and that the wages paid to such labor were unbeleivably small by modern standards. That the smelting process was crude is testified to by 20th century analyses showing that the lead-smelter slag contained 9 per cent lead. At the end of the 16th century, it became profitable to produce galman (hemimorphite) for export to the brass-making establishments of western Germany and Belgium, but production amounted to no more than 100 to 200 tons a year. A galman monopoly was granted to a Polish estate-owner; this monopoly remained in the original family until 1802 and served as the basis of one of the largest mining and smelting enterprises in Silesia. After the first World War, the American Harriman interests bought into this firm, one of the by products of this association being the Childrens' Hospital in Krakow. The mining of sphalerite, to replace the galman that was becoming nearer and nearer to exhaustion,was begun in the third quarter of the 19th century; by 1930, 90 per cent of the zinc came from ZnS. During World War II, production of both lead and zinc greatly increased but only at the cost of damage to the mines and mining out of essentially all high-grade ore. Zwierzycki (1950) estimates the Bytom syncline reserves were about 18 million tons of 11 per cent zinc and 1.5 per cent lead, 11 million tons being primary zinc ore and 7,000,000 of galman ore. Those in the Olkusz area, he thought were about 7.8 million tons of zinc and lead ore combined with a grade of about 11 per cent zinc and 2 per cent lead. He reported that a minor amount of pyrite (37% S) also was available.

This total of 25.8 million tons total Upper Silesian ore estimated by Zwierzycki with 11 per cent zinc is not too far from the Bureau of Mines estimate of 3 million tons of contained zinc now available in Poland, particularly making allowances for additional ore discovered between the end of World War II and today. The 2 million tons of contained lead, however, may be rather high.

The oldest known rocks in the Upper Silesian area are Paleozoic, but they lie almost entirely outside or below the mineralized Triassic rocks volumes. In places these older rocks are appreciably, though not markedly, affected by ore-mineralizing processes; these probably are related to the Jurassic ore mineralization, but this is not certain and remains to be demonstrated fully (if this is possible). The Triassic rocks in the Upper Silesian ore area are divided stratigraphically into Bunter, Muschelkalk, and Keuper epochs. The Bunter is mainly of continental origin and is largely sandstones, although the upper members of the Bunter are marine dolomites, limestones, and marls. The Bunter is overlain by the middle Triassic Muschelkalk, the lowest layer of which is the 30- to 40- meter-thick alternation of wavy limestone and marl, known as the Gogolin beds. Above the Gogolin are 40 to 50 meters of carbonate sediments that are divided into the Gorazdze, Terebratula, and Karchowice limestones. The middle Muschelkalk consists of a single bed, the Diplopora dolomite, that is about 10 meters thick. The upper Muschelkalk is made up of limestones, dolomites, marls, and conglomerates with a total thickness of about 30 m. Above the Muschelkalk is the upper Triassic Keuper, and above the Keuper the beds are of Jurassic and Cretaceous age; although, to the northeast, Jurassic and Cretaceous formations lie directly on the Muschelkalk. The mineralization is confined almost entirely to the carbonate rocks that directly overlie the Gogolin beds. Throughout the mineralized rock volumes, these carbonate rocks are dolomites; outside the mineralized area, however, they are limestones. It appears probable that all this dolomite was developed after the rocks had been lithified, although some geologists think that the dolomite was produced diagenetically. Dolomites in other Triassic beds have much the same content of magnesium but are mineralized only to a very minor extent.

In much of the Upper-Silesian region, the Triassic strata overlie slightly disturbed upper-Carboniferous Coal Measures. Along the northern and northeastern margin of the Silesian Basin, the Triassic rocks transgressively overlap middle and lower Paleozoic beds. These Paleozoic strata are strongly disturbed and compose a positive structural element - the Myszkow-Cracow elevation - this structural element is part of an extensive orogenic belt that includes remnants of the Caledonian structures, but, owes its present shape to Variscan earth movements. The rocks of this elevation include abundant volcanic and intrusive rocks that were introduced during the middle and late Paleozoic.

On the pre-Triassic land surface, resistant Devonian carbonate rocks formed hills that appeared as islands in the Muschelkalk sea and provide the "buried hill" topography that is among the most notable features of the upper-Silesian ore district.

The Triassic rocks are tectonically quite simple; over large areas, these beds are almost flat-lying and their dips seldom are more than five degrees. The first of the tectonic events to affect these rocks were the early Cimmerian movements that closed the Triassic and were completed before Jurassic sedimentation began. These movements produced broad undulations and were accompanied by appreciable faulting. Late Cimmerian movements (post-Jurassic and pre-middle Cretaceous) developed slight uplifts and minor unconformities. Major tectonic faulting occurred in the early Tertiary and during the Miocene. These last faults produced the horst and graben structures that characterize the rocks of the region as it is now known. The grabens were filled during their production and later by Miocene marine clays.

During the early Jurassic, the Triassic rocks were subjected to intense karstification; a second phase of karstification and weathering took place in early Tertiary time and before the middle-Miocene marine transgression.

The ore-bearing dolomite (OBD) is a distinct unit of the Triassic strata in the Upper-Silesian district, and it is the host of the sulfide ores. Most students of the problem consider that the OBD is epigenetic, and this concept appears to have been proved by the work of Bogacz and others, 1972.

The typical mass of OBD is roughly tabular with the horizontal dimension greatly exceeding those in the other two directions; these bodies show metasomatic and cross-cutting contacts with the surrounding limestones and also with early diagenetic dolostones. Thus, the OBD must be younger than these other rocks, and its distribution is far more limited than theirs. The majority of geologists who have studied the OBD are convinced that it and the ores were formed as part of the same general process. Only disseminated zinc and iron sulfides, however, are of exactly the same time of formation as the OBD. The more massive sulfides are distinctly later features of this replacement process. The bulk of the OBD and its sulfides appears to have been introduced after the formation of the Muschelkalk but before the beginning of Jurassic sedimentation. The OBD, probably because of stress distribution developed during its formation, is highly cracked and very susceptible to brittle failure. This contributes much to the further fracturing of the OBD as dissolution occurred as solutions were forced through the formation.

The OBD and its accompanying mineralization are best developed in the fault bounded troughs (grabens) produced during the minor Cimmerian folding and faulting. The grabens originally were synclinal structures but now their margins are fault-bounded. These troughs include: (1) the Bytom (Beuthen) trough, (2) the Wilkosyn-Trzebinia trough, (3) the Tarnowskie Gory (Tarnowitz) trough, and the Krzeszowice-Siewierz monocline. To the north of the Byton and Wilkoszyn structures, the Triassic rocks have been eroded away, exposing unmineralized Carboniferous rocks; still farther north, the Triassic rocks reappear as the Krzszowice-Siewierz monocline, and even farther in that direction pass under the Jurassic rocks of the large Nida trough.

To some extent, the fracturing and fissuring in the OBD and the adjoining Triassic rocks was produced by the Cimmerian earth movements, but these structures were enlarged by hydrothermal karstification processes before lower Jurassic sedimentation began. In many instances, the karst structures so developed became large enough for roof collapse to occur and major masses of breccia to have been formed. Although some breccia may have been tectonic, by far the greater bulk of it seems to have been produced by the failure of the roofs over karst cavities of a wide range of sizes.

Small quantities of ore sulfides are disseminated through much of the OBD. Such sulfides are considered (Mochnacka and Sass-Gustkiewicz, 1980) to be early in the paragenetic sequence. The major ore bodies, however, may assume various positions within the OBD and consist of cavity-filling ores metasomatic ores, and nest-like ores. They are largely concentrated along three levels of "ore horizons" and are separated by barren rock several meters thick. In general, these ores are strata-bound but not stratiform.

The general ore-forming sequence was: (1) formation of OBD, (2) repalcement of OBD by sulfides, and (3) dissolution of much OBD, the production of open-space, and the filling, in whole or more often in part, of these cavities by sulfides. The processes were largely contemporaneous, recurrent, and overlapping, but normally they took place in the order listed above and form integral parts of a single ore-forming process.

A minor, but significant amount of sulfide ore occurs in disaggregated dolomite, that is dolomite that is disaggregated into a friable and porous mass through recrystallization and subsequent dissolution along crystal edges. This process is effected chiefly through the non-integrated movement of aqueous solutions. The final product of disaggregation is a semi-coherent or incoherent mass of fine, silt-sized particles; these masses yield by plastic flow and behave in the same manner as unconsolidated carbonate sediments. Most of these disaggregated dolomites are recemented but prior to that, may have been corroded by mechanical means.

In the disaggregated rock, the galena occurs as euhedral crystals or clusters of crystals or along the boundaries of disaggregated masses or around relics of rocks that have not been so affected. The galena marks the outline of disaggregated dolomite and may be straight or irregular, depending on the shape of the boundaries in question.

In contrast to the ore in disaggregated dolomite, metasomatic sphalerite and galena make up an important part of the Upper Silesian ore deposits. The metasomatic sphalerite ores occur as irregular patches and nest-like accumulation that are roughly tabular bodies that generally parallel the bedding, although these in many places show cross-cutting contacts with the enclosing dolomite. The boundaries of this type of mineralization are gradational over millimeters or centimeters.

The tabular masses of ore largely are in ribbon-like structures that are intermediate between metasomatic ores proper and cavity-filling ores. The patterns of these tabular ores may reflect the fracture textures in the host rock or the bedding planes as the case may be. Whatever the origin of the ribbons, they follow the directions of solution movement through the rocks being mineralized.

The metasomatic galena ores are massive and show sharp, cross-cutting contacts with enclosing dolomite; but, in all cases, replace sphalerite preferentially over dolomite. Sphalerite ore, therefore, definitely predates galena ore in the large majority of instances, although exceptions do occur.

Of equal, or even greater, importance is the cavity-filling ore of which there are two types. The first is sulfide precipitates that exhibit drusy, colloform, mammilary, and grape-like shapes that indicate precipitation under pheatic conditions and occur on the walls of openings. Shapes that form under vadose conditions - stalactites, stalagmites, and dripstones - are rare to lacking. The second type is internal clastic sediments that fill rock openings and consist of solutional residues, disaggregated dolomite grains, clastic ore fragments and authigenic sulfide crystals. These accumulations show sedimentary structures and soft-rock deformations but clearly are not external sediments laid down on the sea floor; their contents of ore fragments and authigenic sulfides is, in many instances, so high that they can be referred to as "sedimentary ores", but they are definitely sediments internal to the cavities in which they occur. They make up only a tiny fraction of the total amount of Upper-Silesian ore. They are significant only in that they have been considered as external, and therefore, syngenetic sediments which they certainly are not, either here or in other Mississippi-Valley-type deposits.

The large amounts of Upper-Silesian ores in cavities of a wide variety of sizes and shapes were produced by deposition from aqueous solutions that also, at a slightly earlier time, were responsible for the formation of these caverns. Most of the caverns contain major masses of dolomite breccia, formed largely by roof collapse produced by the solution-generated removal of support from below. Ore was emplaced in these breccias in much the same time span as breccia formation, although initial breccia development appears to have preceded massive ore deposition and ore deposition to have continued for some time after breccia formation had tapered off. Thus, some breccia fragments contain ore that was emplaced metasomatically before brecciation affected that portion of the rock to any great extent. Then, these ore-containing breccia fragments were cemented by later ore masses deposited in the open space around the fragments. Karst cavities may, however, develop without the formation of major caves. Solution along bedding planes, for example, though not of vertical importance, may produce enough area of open space that fracturing and collapse may occur even though the vertical distance the broken fragments move may be small. Breccia development is not, in many instances a continuous process; the cessation of solution flow stops the development, but a resurgance of solution traverse through the rock volume in question will reinitiate the production of further breccias and additional open space. At any time during breccia formation, the solutions passing through the volumes of broken rock can reinstitute the deposition of ore minerals.

Sass-Gustkiewicz, and others (1982) list the following features as demonstrating the karstic origin of the Upper-Silesian breccias: (1) the highly irregular, tabular, or nest-like form of the breccia bodies; these are commonly branching and bedding controlled; (2) lowermost boundaries are solution surfaces that are the former floors of karst cavities; (3) solution surfaces are covered with internal, in some places stratified, sediments that grade upward into chaotic masses of angular blocks; (4) upper and lateral boundaries of breccia bodies are normally gradational and these pass upward and outward through crackle breccias into networks of cracks and fractures, which, in some places approximate the form of tension domes.

As ore mineralization of the breccias continues, the larger rock fragments become partly to entirely encrusted by sulfides and, locally, the fragments themselves are replaced to greater or lesser degrees of completeness.

The fractures and faults associated with the breccias seldom pass upward beyond the limits of the enclosing OBD.

Most, if not all, of the breccia bodies are the result of karst and not of tectonic processes. Such faults as exist in the OBD are nearly always post-ore, but a few faults that lack vertical displacement may be filled with ore, so some of such openings were available when the ore-forming fluids still were passing through the brecciated rocks after at least some of these faults had been formed.

The zinc minerals in the Upper-Silesian deposits are sphalerite, brunckite, and wurtzite; the lead minerals are galena and boleslawite (a mineral not recognized by Fleischer but considered in Upper Silesia to be a form of lead sulfide). These minerals are accompanied by tiny amounts of rare lead sulfosalts: jordanite [$Pb_4As_2S_7$], gratonite [$Pb_9As_4S_{15}$], and meneghinite [$CuPb_{13}SB_5S_{24}$]. The iron sulfides are marcasite, pyrite, and melnikovite (which Fleischer says is a form of greigite [$Fe^2Fe_2^3S_4$] and which locally means black, finely divided, colloidal sized particles of pyrite). In the main areas of mineralization, zinc sulfides are several times as abundant as galena, but, in the marginal portions of the various structural features of the district, galena is the more abundant of the two. Cadmium usually is present in amounts of less than 25 ppm, silver always is less than 200 ppm and, in most instances is well below 100 ppm. Arsenic particularly and antimony are quite common. Northwest-southeast-trending lines, however, divide the district into high- and low-arsenic and high- and low-antimony; the high areas lie to the northeast of the two lines (though with the high-arsenic line being some 10 km northeast of the high-antimony line). These lines are essentially perpendicular to the direction in which, it is believed, the solutions bringing in the ore sulfides moved through the district, that is northeast-southwest. This matter needs further study.

Prior to World War II and for some years thereafter, the ore-sulfides were considered by Polish geologists to have been formed syngenetically as disseminations in the rocks that now contain them. This viewpoint has been contested since about 1970 but some recent work (Smolarska, 1968; Pawlowska and Wedow, 1980; Zartman, 1980) continues to advance this concept. These latter authors believe that the present epigenetic ore bodies were developed by remobilization from the disseminated syngenetically formed sulfides. Such remobilization would have had to have been achieved by solutions of probably local origin (although this is not an essental requirement) that would be capable of dissolving sulfides disseminated widely, but sparsely, through the OBD and transporting them into the favorable structures (such as solution caverns, karstic breccias, and more porous portions of the OBD). Within such structures, the physical and chemical conditions of the solutions might have been changed sufficientely that they could precipitate the sulfides. Even granting that such marked changes in solution and environmental characteristics could have occurred, thorough examination of the rocks surrounding the present-day ore bodies does not show barren zones around the ores. Instead, a gradual diminution in sulfides is found as the rocks are followed away from the ore bodies as would be expected if the ore fluids had entered the ore-bearing rock volumes through the most easy avenues in ingress and, in those open spaces, had reacted with the wall rocks or within themselves to precipitate their ores.

Another possibility for the emplacement of the sulfides is that they were deposited from descending solutions. Against this possibility is the fact that the ores were, almost in their entirety, deposited before the middle Jurassic (mineralized boulders are incorporated in mid-Jurassic sediments). Since these descending solutions would have had to have accumulated their load of sulfides from rocks formed between Muschelkalk and mid-Jurassic time and since these rocks are notable for their lack of, sulfide mineralization, the concept of these rocks as source beds is eliminated.

Thus, it is necessary to return to the idea of solutions rose from unknown depths and (as is indicated from the position of the ore bodies) and spread out through the Triassic rocks that to convert them to OBD and to deposit sulfide ores by space filling and replacement. Since the massive ores are surrounded by haloes of disseminated ore that become less and less intense as distance from the massive ore increases, this explanation appears to be the most acceptable one in comparison with the other two. The presence of similar (though somewhat higher temperature) ores in the Paleozoic rocks beneath the Upper-Silesian area seems to add appreciable to the validity of this concept. This does not say where or how the ore-forming fluids originated, but it does emphasize that they came from appreciable depths.

It should be pointed out that the total volume of dolomite, covering some 3000 km^2 in the ore district would be far more than would be expected from ore solutions that mineralized only a small fraction of the Upper-Silesian Triassic rocks. This, however, appears to be explained by pointing out that this wide-spread dolomitization is quite different from that of the OBD and almost certainly was produced by diagenesis of the original limey sediments. The OBD, on detailed study, is found to be quite different in character from the large-scale dolomitization that took place before the Muschelkalk beds were lithified, so this objection is resolved. This agrees with Smolarska's 1968 opinion that the dolomitization was diagenetic and not produced by descending solutions. When the work of this distinguised ore geologist was done, she did not have the benefit of more recent studies that show clearly the marked differences between the OBD and the much larger volume of diagenetic dolomite.

Under much of the OBD are beds of what have been known as vitriolic clays; these are argillaceous beds covered with crystals of iron sulfate, hence their name. These have been considered as ordinary sediments interbedded with the carbonate rocks, but it appears more probable that they are residual clays, left behind when carbonate material was removed during the dolomitization of the OBD. These clays were moved by ground water into depressions in the floor of the cavities, there to settle out and form the vitriolic material. The lack of great areal continuity from one area of vitriolic clay to another probably is due to this fact.

Appreciable work has been done on determining lead isotope ratios in the Upper-Silesian ores. The very close agreement in such ratios for $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ strongly suggests that the lead came from a single homogeneous source. This source may have been a magma chamber or sea water; in the latter case, however, the difficulty of getting leads with essentially the same isotopic ratios from a wide variety of surface sources into a major ocean area would seem to bar using that source and makes it much more possible that the source was a major magma chamber at appreciable depth beneath the Triassic rock surface of the district.

The conclusions reached by Sass-Gustkiewicz and others (1982) appear to be pertinent to the solution of the problem of the source of the ore-forming solutions of the Upper Silesian district. These are:

(1) penetration of hot ore-bearing solutions up into and then through the Triassic aquifer, the solutions moving from northeast to southwest;

(2) conversion of a small fraction of the Triassic carbonates to dolomite that became the host rock for the emplacement of disseminated sulfides;

(3) large-scale massive replacement of dolomite by sulfides;

(4) development of major volumes of karst openings, during which process considerable volumes of collapse breccias were produced and the open space so created largely filled by ore-solution-deposited sulfides;

(5) disaggregation of appreciable volumes of already lithified dolomite into some voids in which ore sulfides were deposited.

These processes were essentially contemporaneous, recurrent, and overlapping but, in the larger sense, took place in the order (1-5) just given. All of these processes were parts of one major formative event.

Although Sass-Gustkeiwicz and others (1982) do not speculate as to how and where the ore-forming fluids obtained their heat, metals, and sulfur, some remarks on the subject should be made here.

The higher-temperature, largely fracture-filling ore mineralization in the Paleozoic rocks underlying the area of the Upper-Silesian ore district can reasonably be considered as being related in their processes of formation to the lower-temperature, largely open-space filling deposits in the Triassic OBD. Even if the mineralization in the Paleozoic rocks is found to have been introduced prior to the deposition of the Triassic sediments, the ore fluids that deposited the Triassic ores may have come from a magma that was produced by rejuvenation of that type of molten material in Triassic time.

How the ore elements were transported and why they deposited ore minerals where they did is still moot. Experimental work indicates that the ore metals and sulfur are not likely to have been transported together as metal sulfur complexes unless the ore fluids were far higher in alkalinity than is thought practically possible. If this is correct, then a fall-back position is to have the ore metals transported in metal-chloride complexes with the sulfur traveling with these complexes with the complexes breaking down and to have been produced along with sulfides minor amounts of sulfosalts. It appears almost certainly impossible that the sulfide ions were waiting for the metal-chloride complexes in the sites of Triassic deposition.

At any event, the Triassic deposits in the Upper-Silesian district were laid down almost in their entirety in early to middle Jurassic time with minor amounts of the same sulfides having been introduced during the Cretaceous and at the opening of the Cenozoic. Whether these later mineralizations were the result of further and minor rejuvenation of magmatic activity at those times are were produced by remobilization, it is impossible at the present time to say. Thus, the low-temperature character of the dominant ore minerals indicates that the deposits should be classified as telethermal and the time of (their major) deposition should be designated as middle Mesozoic.

UNION OF SOVIET SOCIALIST REPUBLICS

INTRODUCTION

The total land area of the Soviet Union is about 2.5 times that of the 50 United States, and a very brief and general statistical study suggests that the Soviet Union should contain about 2.5 times as many ore deposits, discovered and yet to be found in that country, as does the United States. In the first Volume of this series, that dealing with the Western Hemisphere, 129 U.S deposits were discussed. If an approximately equal areal coverage were to be given to the Soviet Union, 323 deposits should be included in a comparable volume on the Soviet Union. In Smirnov's three-volume work on "Ore Deposits of the USSR", approximately 325 deposits are considered. On first glance, the near-correspondence of 323 and 325 might suggest that the number of deposits in each country, per any given unit of land measure, is essentially the same. The difference between the degree of exploration and exploitation in the two countries, however, suggests that far more deposits remain to be found in the Soviet Union or to be described in the Soviet literature than is true of the United States.

From the huge number of Soviet deposits available for inclusion in this Volume, I have selected only 30. Certainly no publisher would provide space for me to get more than 30 discussions into print, nor would I have the ability, geological or linguistic, to increase the chosen 30 to any appreciable extent. What I have tried to do is to select these 30 so that they would give as wide a coverage of Soviet-deposit types as possible.

I have divided the 30 deposits selected into two categories: (1) those deposits in, or west of, the Urals and west of the Caspian Sea and (2) those east of the Urals and east of the Caspian Sea. Of the 30 chosen, 16 are in category (1) and 14 are in category (2). A complete list of these deposits is as follows:

DEPOSITS IN, OR WEST OF THE URALS AND WEST OF THE CASPIAN SEA

POLITICAL SUBDIVISION	DEPOSIT NAME	LATITUDE AND LONGITUDE	METALS PRODUCED
Armeniyan SSR	Zod	40°12'N,45°52'E	Gold, Silver, Tellurium
Bashikirian SSR	Sibai	52°43'N,58°39'E	Copper, Zinc, Pyrites
*Chelyabinsk Oblast	Kusinsk	55°21'N,59°28'E	Gold
Georgian SSR	Chiatura	42°19'N,43°18'E	Manganese
*Kabardino-Balkarskaya ASSR	Tyrnyauz	43°25'N,42°55'E	Molybdenum, Tungsten
*Karelian ASSR	Pitkaryanta	61°31'N,31°30'E	Tin, Copper, Zinc, Iron
Kazakhstan SSR	Kempirsai	50°36'N,58°15'E	Chromium
*Kurgan Oblast (Eastern Urals)	Berezovo	?54°40'N,64°26'E	Gold
*Murmansk Oblast	Pechenga	69°28'N,31°04'E	Nickel, Copper, Cobalt, Silver, minor Pt metals
*North Osetian ASSR	Sadon	42°50'N,44°03'E	Lead, Zinc

POLITICAL SUBDIVISION	DEPOSIT NAME	LATITUDE AND LONGITUDE	METALS PRODUCED
*Sverdlovsk Oblast (Middle Urals)	Kachkanar	58°45'N,59°23'E	Iron, Vanadium, Titanium
	Nizhniytagil'	57°55'N,59°58'E	Platinum Metals, Gold
Ukraine SSR	Kerch	45°20'N,36°27'E	Iron, minor Manganese
	Krivoy Rog	47°55'N,33°24'E	Iron
	Nikopol'	47°34'N,35°25'E	Manganese
	Nikitovsk	48°20'N,38°03'E	Mercury

Note: Deposits marked with an * are in political subdivisions of the Russian Soviet Federated Socialist Republic

DEPOSITS EAST OF THE URALS AND EAST OF THE CASPIAN SEA

POLITICAL SUBDIVISION	DEPOSIT NAME	LATITUDE AND LONGITUDE	METALS PRODUCED
*Amur Oblast	Solnechnoe-Festival'noe	ca50°32'N,137°00'E	Tin, Tungsten, Copper(?)
*Buryat ASSR	Dzhida	50°20'N,6°59'E	Tungsten, Molybdenum
	Ozernoe	ca53°00',114°00'E(?)	Pyritic-Polymetallic
Kazakhstan SSR	Dzhezkazgan	47°48'N,67°24'E	Copper, Lead, Zinc, Silver, Rhenium
	Kounrad	46°58'N,75°00'E	Copper, Molybdenum
	Mirgalimsai	43°32'N,68°36'E	Lead, Zinc, Barium
	Ridder-Sokol'noe	50°23'N,83°32'E	Pyritic-Polymetallic
	Zhairem	ca48°20'N,70°30'E	Pyritic-Polymetallic
Kirghiz SSR	Kadamdzhai	40°07'N,71°44'E	Antimony
	Khaidarkan	39°57'N,71°20'E	Mercury, Antimony
*Krasnoyarsk Kray	Noril'sk	69°20'N,80°06'E	Nickel, Copper, Cobalt, Pt Metals, Gold, Silver
*Primor'ye Kray	Tetyukhe	44°35'N,135°35'E	Zinc, Lead
Uzbekistan SSR	Kal'makyr	40°50'N,69°40'E	Copper, Molybdenum, Gold
	Muruntau	41°23'N,64°23'E	Gold, minor Silver

The deposits in the Russian Soviet Federated Socialist Republic (RSFSR) number 11, although the RSFSR occupies about three-quarters of the total land mass of the USSR. Of course, more than half of the RSFSR is in Siberia and in northern Siberia at that, so exploration in that portion of the USSR has proceeded at a much slower rate than in middle or central Asia. Nevertheless, the completion of the BAM rail line, much farther north than the Trans-Siberian, will provide much more ready access to areas to be prospected in the far north or to ore bodies that already have been discovered there, deposits that await only better means of transport to be put into development and exploitation.

Thus, the deposits I have selected, a selection which was made only on their interest for non-Soviet ore geologists, seem to be well distributed throughout the USSR with what I think is a sound explanation for the relatively low number in the RSFSR and its subordinate political entities.

The age of the deposits chosen ranges from middle Precambrian to Quaternary with the distribution being as follows: (1) middle Precambrian - 2; (2) late Precambrian - 2; (3) early Paleozoic - 1; (4) middle Paleozoic - 2; (5) late Paleozoic - 12; (6) early Mesozoic - 3; (7) middle Mesozoic - 2; (8) late Mesosoic - 3; (9) early Tertiary - 2; (10) middle Tertiary - 1; (11) late Tertiary - 0; (12) Quaternary - 1. This is not, of course, an even distribution, but its range is sufficient to give some idea of the variety of ages in which ore deposits were developed in the USSR. If my sample had been truly random (and the number of deposits chosen larger), it might be safe to say that the late Paleozoic (Hercynian) orogeny was the most productive time in geologic history for ore deposits to form in the USSR, but this conclusion cannot be reached on the limited data here provided.

As for the conditions under which these deposits were formed, the same diversity of types can be seen as was true of age of ore-deposit formation. The summary of conditions of formation is as follows: (1) Magmatic deposits (of all types) - 6; (2) Hypothermal deposits in non-calcareous rocks - 7; (3) Hypothermal in carbonate rocks - 4; (4) Mesothermal - 11; (5) Leptothermal - 4; (6) Telethermal - 2; (7) Xenothermal - 1; (8) Kryptothermal - 2; (9) Epithermal - 4; (10) Metamorphic (normally only being part of the ore-forming process in any given instance) - 3; (11) Sedimentary (in its various forms) - 5; (12) Ground-Water Circulation - 4. This list totals 50, which is explained by many individual deposits having been formed over a series of conditions, such as a deposit having been produced in the range from hypothermal to mesothermal, for example - Sadon or a deposit having been developed under both types of hypothermal deposition, for example - Tyrynauz or a deposit having been formed under magmatic conditions of one variety or another and then so strongly affected by metamorphism that that effect must be included in the classification, for example - Noril'sk. Again, however, the 30 deposits, in their 50 varieties of temperature and pressure control, are quite widely distributed among the various classification categories.

I must emphasize, and probably should have done so sooner, that I did not select these 30 Soviet deposits on the basis of my personal knowledge of the ore geology of the USSR. The main source for the study of the deposits, selected particularly for someone whose knowledge of the Russian language is as limited as mine, is V.I. Smironov's "Ore Deposits of the USSR", in three volumes, the English translation having been published by Pitman Publishing Company of London in 1977 and having been translated by D.A. Brown. The original Russian edition was published by the Nedra Press of Moscow (an arm of the Soviet Ministry of Geology) in 1974. Smirnov's volumes are the source of most of the bibliographic material I have presented here, up to 1972, the last publication year that Smirnov was able to include in the Russian edition. For references later than 1972, I am indebted to Academician V.I. Smirnov himself, to Academician N.P. Laverov, the Editor in Chief of *Soviet Geology*, and to V.A. Baskina of the Institute of Ore Deposit Geology. The translations of the references so supplied were prepared by Dr. Baskina and by my wife. Recently assistance in these matters was given by Dr. A.D. Sokolov, also of the Institute of Ore Deposit Geology, normally known by a few of its initials as EGEM. Professor A.D. Genkin and Dr. Ruslan Volkov also have been most helpful. In this country, Professor E.A. Alexandrov has been of great assistance.

For copies of the Russian papers that I particularly wished to have translated, I am greatly indebted to the Library of the U.S. Geological Survey in the National Center at Reston, Virginia and, in particular to Barbara Chappell, Chief of the Reference Division of that Library. Without her help, this project never would have got off the ground.

The logistical and technical help provided by the Department of Geology at the University of Florida, where I have the honor and pleasure to be a visiting professor has been consistently helpful.

What is good in these bibliographies and discussions accompanying them is due to the help so generously given me by these Soviet ore geologists that I have mentioned; what errors of omission and comission this section of my Volume III yet contains must be charged entirely to me.

Because of the great increase in the cost of this volume that would have resulted if the titles of the references and of the journals from which they were taken had been given in the Russian language, I have provided these only in translations.

Thus, those using the Soviet section of this Volume and wishing to study the references cited in their original Russian can, I am sure, retranslate my translations, if necessary consulting the introductory pages of the Bibliography and Index of Geology of the Geological Society of America, to find the full name of most of the publications mentioned here.

The Russian section is included in this Volume in the hope that it will give the non-Russian, but English, speaking geologist some concept of the ideas of Soviet ore geologists and some understanding of the varied geology of the ore deposits of that huge country.

Except for the rare instances in which production, grade, and reserve figures of given in Smirnov's three-volume work of the Ore Deposits of the USSR, I have had to depend on material from sources outside the USSR. The main one of these is the annual publication - "Gold" of Consolidated Gold Fields PLC (formerly Consolidated Gold Fields Ltd). Since, quite reasonably, this publication is concerned with Soviet gold deposits, information about any other type of deposit is given only if appreciable gold is a by-product of gold production as is the case in such deposits as Kal'makyr and Kounrad. The figures given for production grade, and reserves by Consolidated Gold Fields is, of course, as accurate as someone observing Soviet mining operations from outside that country can be, but it must be remembered that none of the figures given from this source can be guaranteed absolutely for accuracy, nor does Consolidated Gold Fields claim them to be. Nevertheless, they are by far the best figures available and are given with confidence that they are close to being right.

ZOD, ARMENIYAN SSR

Late Mesozoic	Gold, Silver Tellurium	Krypothermal to Epithermal

Amiryan, Sh. O., 1972, The Zod deposit: Akad. Nauk Armeniya, Publications, p. 85-92

Borodaevskaya, M. B. and Rozhkov, I. S., 1977, The Zod deposit, in Smirnov, V. I., Editor, Ore Deposits of the USSR: Vol. III, Pitman Pub. Co. London, p. 19-21

Glynn, Ch., 1979, The Zod Combine, in Gold 1979: Consolidated Gold Fields Limited, London, p. 48-50

Konstantinov, M. M. and Gurshkin, V. A., 1972, The structural conditions of localization of the Zod gold-ore deposit: Institute of Higher Learning, Tr., Geol. and Exploration Repts. (Izvestia), no. 5, p. 60-67

Lindgren, W., 1933, Epithermal deposits, in Mineral Deposits: 4th ed., McGraw-Hill, N.Y., p. 444-455

Zil'man, F. Ye., 1972, Certain characteristics of the morphology and structure of mineralized veins in massive gabbro at the Zod gold deposit: Akad. Nauk Armeniyam Repts. (Izvestia), Earth Sci. ser., v. 25, no. 4, p. 82-85

Notes

The Zod gold deposit is one of several gold deposits that recently (relative to 1979) were begun to be exploited in Armenia. In contrast to the gold deposit of Murantau, Zod is small but it is the largest of these Armenian deposits; Zod, although much smaller than Muruntau, is much higher in grade. So favorable did the prospects for this area seem that soon the Zod Combine was formed, and only a few years ago, a major mining-milling-smelting complex was constructed by the Combine at a cost reported to have been more than 150 million roubles (Glynn, 1979).

The gold deposit at Zod is located about 17 km immediately east of the southeast corner of Lake Sevan, 120 km east of Yerevan, the capital of the Armenian SSR, and about 100 km northeast of the Turkish border. The elevation of the mine workings is between 2200 and 2500 m above sea level. In the far-off Neolithic times of the late Stone Age, placer deposits were worked in nearby rivers; by 1200 BC the outcrops of the Zod deposit had been found and were worked by trenches that were 1 m wide and 5-6 m deep. By 200 AD, the richest ores of the deposit were being exploited by underground mining operations. As has so often been true, when these really rich ores were exhausted, the major mine workings were abandoned. In 1951, the area was re-examined, first the river placers and then the ancient workings high in the mountains. This exploration indicated more rich ore at depth, and modern exploitation of the Zod area began (Glynn, 1979).

The Zod deposit is composed of oxidized ore near the surface, primary gold-bearing sulfides at depth, and mixed ore in between. Present information suggests that these mixed ore make up the major portion of the commercial ore in the Zod mine. The mineralization is reported to be localized in four fracture zones that dip steeply (70° to 85° to the south) and strike more or less east-west (Glynn, 1979).

The gold at Zod is present in the Zod ore bodies in four forms: (1) coarse native gold among the grains of gangue and ore minerals; (2) native gold as finely dispersed particles in sulfide minerals; (3) distinct telluride minerals and (4) gold derived from the disintegration of both sulfide and telluride minerals. The fine gold is the predominant form of that metal, but the ore contains coarse flakes that are as much as 1 cm across.

Polymetallic ore from the Zod deposit averages 20 grams of gold per ton; pyrite-arsenopyrite ore 40 grams per ton; and gold-telluride ore averages 300 grams per ton. The gold ranges between 850 and 950 fine. Both silver and tellurium are commercially recoverable from the Zod ores (Glynn, 1979).

In 1979, the Zod deposit was being mined both by underground (adit) and open-pit methods. The first treatment of the ore (1967) was carried out in a small plant about 5 km from the mine. Further discoveries of similar ores were made west of Lake Sevan; these last were quite close to an existing rail line. To prevent contamination of Lake Sevan (by cyanide), the processing plant was located in the town of Ararat, 30 km southeast of Yerevan and at the foot of Mt. Ararat. The processing plant is only about 7 km from the USSR-Turkey border. The rail transport of the Zod ore to Ararat required the construction of 122 km of track to connect the mine with the existing rail line that passes west of Lake Sevan. The first Zod ore was delivered to the Ararat plant in February 1976 (Glynn, 1979).

The plant also treats ore from the other mines discovered after it was decided to reopen the Zod mine. At the plant, a gravity concentrate is produced that contains the coarse native gold and tellurides. The tailings from this gravity process (containing sulfide and mixed ores) is floated to give a concentrate that includes gold-bearing sulfides and finer tellurides. The tailings from the floation operation are reground in ball mills and treated by the cyanide process to recover most of the remaining gold (Glynn, 1979).

The annual production of gold from Zod and those mines west of Lake Sevan (Megradzor, Kefan, Kirovakan, Radzan, and Stepanavan) is estimated by Glynn at 10 tons.

The ores of the Zod deposit can be considered as belonging to the gold-quartz association; the deposit itself is located in the extreme southern part of the USSR's metallogenic province of the Caucasus. The mineralized area that includes the Zod ore bodies is contained in the area of the junction of: (1) a near east-west striking ophiolite belt and (2) a zone of essentially north-south striking major faults. The ore bodies occur in the rocks of a gabbro-peridotite massif of Turonian-Cenomanian age (lower upper Cretaceous); this massif is contained in an upper Cretaceous volcanic-sedimentary rock sequence in the structural complex.

The gabbro-peridotite mass is cut by dikes of quartz porphyries that probably are genetically associated with the mineralization. The Zod ore bodies, in turn, are confined to complexly structured contacts between gabbros and peridotites in a part of an east-west fault. The gabbro entered the peridotite mass as a series of subparallel, en echelon apophyses; the apophyses are oriented in a generally east-west direction. This same complex contact zone is made even more so by the quartz rhyolites dikes that cut the gabbros.

Along the same strike as that of the quartz-rhyolites dikes that cut through the gabbros, zones of major hydrothermal alteration were developed in these rocks and along the gabbro-peridotite contacts as well. Locally, these alteration zones are found in both the peridotites and even in the volcano-sedimentary rocks. This hydrothermal alteration produced quartz-carbonate-talc bodies with a wide range of thicknesses. It is among these hydrothermally altered bodies that the ores were emplaced. The actual ore bodies are either quartz veins or segregations of quartz-carbonate veinlets heavily impregnated by sulfides.

The number of ore and gangue minerals in these quartz veins and quartz-carbonate veinlets segregations is huge. These include: pyrite, arsenopyrite, marcasite, sphalerite, pyrrhotite, chalcopyrite, tennantite-tetrahedrite, freibergite (Ag-bearing tetrahedrite-tennantite), stibnite, and gold. Appreciably more rare galena, sulfantimonites of lead, jamesonite, boulangerite, zinkenite [$Pb_6Sb_{14}S_{27}$], meneghinite [$Pb_{13}CuSb_7S_{24}$], geocronite [Pb_5SbAsS_8], berthierite [$FeSb_2S_4$], melonite [$NiTe_2$], altaite [PbTe], sylvanite [$AgAuTe_4$], nagyagite [$Pb_3Au(Te,Sb)_4S_{5-8}$], hessite, calaverite, petzite, tellurobismuthite [Bi_2Te_3], pentlandite, gersdorffite, ullmanite, corinite [formula unknown (to me)], polydymite [$NiNi_2S_4$], millerite, linnaeite [$CoCo_2S_4$], bismuthinite, stannite, bournonite [$PbCuSbS_3$], chalcostibite

[$Cu_6Tl_2SbS_4$(?)], weissite [Cu_4CuTe_3], rickardite [$Cu_4Cu_3S_5$], tellurium, antimony, coloradoite, emplectite, argentite, bornite, violarite [$FeNi_2S_4$], niccolite, siegenite, valleriite [$4(Fe,Cu)S \cdot 3(Mg,Al)(OH)_2$], rammelsbergite, glaucodot, löllingite, smaltite, bravoite, and chloanthite.

The principal gangue minerals are: quartz, chalcedony, carbonates, rhodchrosite, talc, sericite, zeolites, and serpentine.

The work of Konstantinov indicates that five stages of primary mineralization can be recognized in the Zod ores; these are: (1) pre-ore quartz-sericite-pyrite; (2) quartz-ferroan magnesite-pyrite-arsenopyrite, the first ore-bearing stage; (3) quartz-marcasite-arsenopyrite-sulfantimonites-tellurides, the second ore-bearing stage; (4) quartz-stibnite, post-ore; (5) late quartz-carbonate, post-ore.

This author reports that native gold is disseminated throughout the ore veins and veinlet segregations, and is normally found in connection with bismuth, nickel, and silver tellurides and with sulfosalts of antimony and lead and with sulfarsenides. Much of the gold has xenomorphic outlines as a result of filling-in spaces among earlier grains, the first of the types of ore mentioned earlier in this discussion. The particles of gold range from 0.005 mm to 1 (or even as much as 2) cm in diameter.

Although the wide variety of sulfosalts are such as to suggest strongly that the ores were formed under epithermal conditions, Konstantinov believes that the hydrothermal alteration associated with the ore was formed under conditions of temperature and pressure too high to belong to that category in Lindgren's classification. Lindgren (1933) points out, however, that the most common economically valuable minerals are gold and silver. Arsenic, antimony, bismuth, tellurium and selenium are common but only rarely of economic importance, Te, however, is recovered at Zod. The gold in epithermal deposits almost always is alloyed with silver (and on occasion with mercury). If native silver is present, it normally is a product of oxidation. In most instances, silver is found in argentite, but it ordinarily occurs in sulfosalts of antimony and arsenic, such as proustite and pyrargyrite and tetrahedrite and tennantite. Arsenopyrite is rare and pyrrhotite and magnetite are absent, but stibnite is quite common.

Quartz is the most abundant gangue mineral with chalcedony, dolomite, barite, and fluorite siderite is rare, but rhodchrosite is locally quite common.

Wall-rock alteration generally is propylitic, the typical suite being chlorite, carbonate, and pyrite with some sericite. Adularia may be in epithermal deposits in some abundance.

This short summary of Lindgren's ideas on epithermal deposits would seem to agree with Konstantinov's statement that the Zod deposits were formed, at least in part, at temperatures above the epithermal range. The wide variety of sulfosalts that are not typical of true epithermal deposits and the very considerable amounts of gold recovered from them indicates that part of the deposition of the ore occurred in the kryptothermal range. Further, the wall-rock alteration is more characteristic of the higher category (kryptothermal) than of the lower (epithermal). Nevertheless, enough of the gold and the tellurides probably were deposited in the lower-temperature range, so that the deposits of the Zod area ought to be categorized as Kryptothermal to Epithermal.

With the exception of the mercury-antimony deposits of Kirghizia and the mercury deposit of Nikitovsk in the Ukraine, Zod is only the fourth deposit in the USSR to be placed in the epithermal category in this Volume.

The close association of the gold deposits of Zod with the late Cretaceous igneous activity of the general region makes a strong case for classifying these deposits as Late Mesozoic in age.

SIBAI, BASHKIRIAN ASSR

Middle Paleozoic	Copper, Zinc Pyrite	Hypothermal-1, 2 (minor) to Mesothermal

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Notes

The Sibai mine lies in a valley at the foot of the Urals, about 290 km north of the city of Orenburg, and was discovered in 1913. The deposit is divided into two parts, the Old Sibai sector to the north and the much larger New Sibai sector to the south; both are now being exploited as open pits.

The first mining was begun in 1925 to extract the high-grade portions of the deposit with the mining of the remainder to be commenced later. In 1956, when the Old Sibai pit was 60 m deep, the removal of the overburden over the New Sibai ore body was begun. At the surface, the two pits are now joined into one; the rocks in the old pit, however, are quite flat, so that pit was exploited only to a depth of 90 m. In the much larger, steeply dipping new pit, the ore body reaches a depth of about 400 m. The total deposits is expected to yield 85 milllion tons of copper-zinc ore that will average 0.84 per cent copper.

The Sibai deposit is now highly profitable with mining scheduled to continue through 1992 at a rate of 4 million tons per year. The gold content of the ore is about 0.8 gm per ton, so the Sibai gold production is 2.2 tons per year.

The town of Sibai, near which is the mine of that name, is located at 52°43'N, 58°39'E and is about 85 km SSE of the city of Magnitogorsk, near the eastern border of the Bashkirian ASSR, and on the western slope of the Ural Mountains, the Sibai ore bodies are contained in the western limb of the Magnitogorsk megasynclinorium, specifically in the western subzone of the subordinate Uchala structural complex, and, within that last structure, the ores are restricted to the northwestern section of the short, broad Sibai volcanogenic anticline. This anticline is oriented in a generally north-south direction, is about 12 km long, and is from 2 to 3 km wide. The oldest rocks in the core of this anticline are volcanogenic rocks of the middle Devonian (Eifelian) Karamalytash group; they are overlain by the Bugulygyr red jasper horizon that is Givetian, younger middle Devonian than the Eifelian. Outward from the eroded core of the anticline is the also Givetian Ulutau group and above that sedimentary beds of upper Devonian age.

The anticline itself is asymmetrical with rocks of the western limb dipping outward at from 10° to 30° and those of the eastern between 60° to 70° in an easterly direction. At the south end of the anticline, the beds dip uniformly outward, but at the north end, the beds are cut by a diagonal thrust fault along which the north side has been dropped down. To both east and west, the anticline is bordered by normal N-S faults, two on the west and one on the east; the displacement on the east side of the east fault is probably more than 300 m; the faults, both east and west, are bordered by breccia zones as much as 70 m thick. At the south end of the anticline the periclinally dipping beds are Givetian; the beds to the north within the graben are all formations of the Karamalytash (Eifelian) group. Outside the faults, the rocks abut against terrigenous eruptive-sedimentary and sedimentary beds of both middle and upper Devonian age. In the eastern limb of the anticline, a blocky structure resulted from less impressive N-S faulting, plus diagonal and more or less E-W faults. In many instances, the larger faults were first developed prior to the introduction of the Eifelian-Givetian volcanic and sedimentary rocks, and these faults also effect the location and thickness of the various types of volcanic rocks.

The volcanic rocks of the anticline all belong to the Karamalytash group; in the area of the Sibai deposits, the volcanic rocks of the Karamalytash group have been divided into five members; these are (from bottom to top: (1) the lower basaltic, at least 400 m thick; (2) the dacitic, lavas with minor intercalations of tuffites that range in thickness from 200 to 1000 m; (3) the middle basaltic, massive and pillow lavas with interspersed seams of radiolarian jaspers, scoriaceous sandstones, and tuff-bearing sandstones with their thickness being 0 to 200 m; (4) the rhyolitic, made up of about equal thicknesses of lavas and pyroclastics with their thickness ranging from 0 to 500 m - in the upper portion of the rhyolite, there is a aggomeratic tuff of varied composition that is as much as 30 m thick, and (5) the upper basaltic formed of coarse pyroclastics and thin seams of lava.

The totality of this almost completely volcanic material is 40 per cent rhyolite-dacite. This entire sequence has been observed only in the northern part of the anticline; the higher members of the group thin out as followed south; in fact only along the eastern limb of the anticline can members (4) and (5) be followed to near its southern end. Member (3) is confined almost exclusively to the middle of the anticline. The dacites [member (2)] probably can be traced downward (in several places for up to 1000 m) into the vents of the volcanic-rock group. No evidence has been found of erosional unconformities between any two of the members of the group; such angular differences as there are between one group and the next are probably the result of different directions of lava flow.

As has been mentioned, the representative of the Givetian in the area of the Sibai ores is made up of red jaspers with thin beds of tuffites and tuff-bearing sandstone and is known as the Bugulygyr horizon; it lies unconformably on the Karamalytash.

The sequence of mainly volcanic rocks of Eifelian age was a long time in formation, being a multistage process in an area where uplift went on throughout the entire Eifelian. Submarine eruptions were succeeded by near-shore marine and even subaerial varieties. Movement on the major faults, at least, continued through the all the time of the accumulation of volcanic material, causing much of the variation in water depth (or lack of it) that characterized the Eifelian stage.

The development of the upper basaltic member terminated with the intrusion of two dike complexes: (1) diabase and (2) pyroxene-plagioclase porphyrites; also introduced were gabbro-diabase dikes. All of these dikes cut all of the rocks of the region, including the primarily sedimentary Givetian Ulutau group, this indicating that igneous activity continued after the accumulation of the entire thickness of Eifelian volcanics. These dikes are quite thin (less than 2 m) and cut all the rocks of the district, including the ore bodies.

The ores of the Sibai district are confined to the northern part of the eastern limb of the anticline where the structure is that of a graben. The eastern boundary of the graben is the Eastern fault and the western is another N-S striking fault - the Central. All of the ores are in the rhyolite [member (4)] which reaches its greatest thickness of 500 m within the graben. No ore bodies are known outside this graben.

The Sibai ore bodies, Northern and Southern, are divided from each other by a large, NE-striking fault that displaces not only all of the volcanic formations of the Karamalytash group, but the Central, ore-controlling fault as well. Blocky structures dominate on both sides of the diagonal fault, but the forms and attitudes of the ore bodies are quite different on one side of the fault from those on the other.

In both sections of the Sibai ore bodies, the ore appears to have been introduced in two stages. The Northern (Old) Sibai ore is concentrated in a layer-like structure; the eastern (and major) part of this ore body appears to lie in the rhyolite of member (4), just below its contact with the upper basaltic rocks above. The dip of the ore, therefore, is the dip of the contact which slopes at gentle angles to the east. Farther west, and separated for the major part of the ore by a fault that dips steeply east is a similar but smaller segment of ore. The total volume of the Old Sibai ore appears to be much less than that of the New Sibai bodies to the south, although only one cross section is available from which to judge. In the layer-like structure, the thickness is quite varied, ranging from 1 to 20 m.

Uniform (disseminated) ores lie on top of the so-called segregated ore that appears to be composed mainly of agglomeratic tuff fragments (clasts), the tuffs being of several silicic varieties that are cemented by what probably is a welded matrix. This bed is, as the section shows, at the top of the rhyolite member (4). The clastic fragments principally are silicified and sericitized rock that is accompanied by blocks of sulfide ore as much as 1 m in diameter. If the brecciation of the tuffs was due to volcanic activity, the rocks are pyroclastics, but the brecciation may have been tectonic.

The smaller segment of ore to the west of the fault that separates the two segments of Old Sibai ore lies beneath a fault that dips about 45° generally east. As a result of this faulting, the smaller (more westerly wedge) is enclosed within the (4)-member rhyolite.

The much larger of the two ore bodies is the New Sibai [as far as can be judged from the section shown (Samonov and Pozharisky, 1977)]. Most of the New Sibai ore body appears to be in bulges of a more massive type than is the old Sibai ore, is composed of pyrite, sphalerite, and chalcopyrite, and is most complicated in shape. It is enclosed mainly in the (4)-rhyolite member, although the ore farthest east and uppermost in the body abuts against the upper basalt (5). Two major triangular-shaped branches extend downward toward the east at quite steep angles. The larger of branches of the New Sibai ore (the one to the west) cuts off downward against the middle basalt (3). As opposed to the massive ore of the main ore bulge, downward from which the branches extend, the branches themselves contain both massive and clastic ore. This clastic ore is similar to that of the Old Sibai layered ore. In the southwestern portion of the New Sibai ore body, the ore has replaced what Ivanov (1947) thought to have been limestones developed as a small reef; fossils in this limestone have been replaced by sulfides.

Samanov and Pozharisky (1947) distinguish two stages of ore formation in the Sibai ore bodies. The first stage, they report, is massive in structure and is made up of chalcopyrite, sphalerite and pyrite. This ore type was formed in the bulges of the New Sibai ore body and probably in the lower parts of the Old Sibai ore bodies. These massive ores are composed of three types: (1) early pyritic; (2) chalcopyrite-pyritic; and (3) sphalerite-pyritic. In addition, and of minor importance, is the ore charactreized by late pyrite and quartz-carbonate that contains sphalerite (cleiophane), chalcopyrite, and tetrahedrite-tennantite. In the early ore, trace elements are limited almost entirely to cobalt in pyrite. The three types usually are associated together, and such ores are patchy, in fine veinlets, or in breccias. In the upper portions of the New Sibai ores, colloform structures have been identified and replacement textures are present.

The second-stage ores are only roughly conformable to the bedding, often have clastic textures, and are composed mainly of chalcopyrite, sphalerite, and pyrite. They occur largely in the Old Sibai ore bodies and in those small, layered bodies that are in the wall rocks of the New Sibai ores but are directly connected with them. Contained in these layered ores are thin seams composed of quartz-hematite, quartz-chlorite, and cherty oolites; in some of the nuclei of these last are occurrences of chalcopyrite, sphalerite, and pyrite.

In the rhyolite that enclosed the Sibai ores are sulfide fragments that are said to be mainly of a pyroclastic nature. These pyroclastic fragments are present in the layered branches extending outward (and upward) from the New Sibai ores and in the clastic ores in the Old Sibai body.

Hydrothermal alteration is reported [Samanov and Pozharisky (1977)] to have affected the country (rhyolite mainly) rocks of both deposits but is quite weak. The development of sericite and quartz were the most prominent effects, but chloritization has been noted at the contacts between country rock and segregated ore. Most of this alteration, apparently, is localized around the first-stage ore bodies, these aureoles being from 10 to 30 m thick.

Where the ore bodies wedge out down dip, the alteration continues for some distance beyond those points. Although the bulk of the altered rock is made up of quartz and sericite that replaced the host rhyolite; along the outer boundaries of these alteration zones, albite and epidote become prominent. Alteration of the ores by surface process has reached down into the ore bodies by as much as 100 m. In the first 20 m of this distance, the ores have been changed to brown oxides of iron; secondary copper ores have been produced down to 40 to 50 m below the surface and, along fault zones, as far down as 100 m. The principal secondary minerals in the oxidized zone are goethite, kaolinite, and hydrogoethite; in the zone of secondary sulfides, covellite, and chalcocite are important. Minor minerals in the oxidized zone mainly are hematite, cuprite, chalcanthite and, locally, malachite, azurite, jarosite, opal, and native copper.

The principal economically recoverable metals are copper and zinc; sulfur is produced principally from the pyrite. The ratio of copper to zinc covers wide ranges - from 1:1 to 1:15; the average is 1:1.6. A large number of trace elements are known - gold, silver, cobalt, cadmium, selenium, tellurium, plus arsenic, indium, thallium, gallium, and germanium as traces. It is not stated if any of these trace elements are worth recovering.

In addition to the main ore minerals - chalcopyrite, sphalerite, and pyrite, the number of minor minerals is considerable - melnikovite, pyrrhotite, and magnetite are most common with only traces of tennantite, galena, arsenopyrite, gold, hematite, bornite, enargite, freibergite, cinnabar, greenockite, and germanite [$Cu_3(Ge,Fe)(S,As)_4$]. The most common gangue minerals are quartz and calcite with siderite, chlorite, sericite, gypsum, and barite being less usually seen.

Ninety per cent of the total volume of commercial ore is of the chalcopyrite-sphalerite-pyrite type; of much less abundance and value are the pyrite and zinc-pyrite ores; in this last type, the ratio of copper to zinc is 1:15. Minor amounts of ores that also include pyrrhotite, magnetite, and siderite are found at depth along the footwalls of the New Sibai disseminated ores.

The genesis of the Sibai ores has not been fully explained. It is definitely stated [Samanov and Pozharisky (1977)] that there are two stages of ore formation. The bulk of the ore is contained in the older of these two stages with that in the younger being quite minor in total amount. In the Old Sibai deposit, the ores are contained in a layer of agglomerate tuffs the matrix of which has been welded; the presence of fragments of disseminated-sulfide ore (up to 1 m in diameter) in association with the welded tuffs indicates that the ores had been emplaced before the fracturing of the sulfide ores had taken place. Also fractured, apparently at the same time as the sulfides were broken, were the welded tuffs that now are mainly silicified and sericitized. Whether the brecciation was due to volcanic or tectonic activity is uncertain. Whatever the cause of the brecciation, it almost certainly took place after the first-stage ores had been introduced.

The first-stage ore in the New Sibai ore body appears to be more massive than that of the first-stage in the Old Sibai mineralization. The presence of both massive ore and clastic ore in the downward-trending branches of the Sibai ore bulges, indicates that the massive ore was fractured after it had been emplaced.

The second-stage ores in both Old and New Sibai ore bodies are only roughly conformable to the bedding, suggesting that they were introduced by replacement of fragments of the tuff, which fragments were largely developed in the apophyses of ore that extend upward from the bulges in the New Sibai ore body. As these second-stage ores also include seams of quartz-hematite, quartz-chlorite, and cherty oolites, it seems probable that they were formed hydrothermally and not volcanogenically, although the ore-fluids probably came from the same source (or sources) that produced the ore fluids of the first stage.

The weak, but rather wide, zones of alteration around the ore bodies seems to indicate that the first stage mineralization also was brought in after the rocks had been lithified. The tectonic or volcanic brecciation that followed the first-stage mineralization and preceded the second, however, might argue that the first stage was the result of volcanogenic exhalations; it appears more probable to me that both were the result of replacement affected by hydrothermal ore fluids.

Because the brecciation between the first- and second-stage ore formations took place within the Devonian Karamalytash group, the introduction of the ore probably should be dated as middle Paleozoic.

The presence of pyrrhotite and magnetite, particularly at depth in the New Sibai ore body, indicate that some of the mineralization at depth took place in the hypothermal range, but the major part of the ores almost certainlywas deposited under mesothermal conditions. The Sibai ores, therefore, are here classified as hypothermal-1, -2 (minor), Mesothermal. The designation "Hypothermal -2" apparently is justified by the replacement of a small body of limestone, probably a reef, in the southwestern part of the New Sibai ore body by ore minerals.

KUSINSK, CHELYABINSK OBLAST

Late Precambrian	Iron as Magnetite, Titanium, Vanadium	Magmatic-3b Metamorphic-C

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Notes

The Kusinsk titanomagnetite deposit is at 55°21'N,59°28'E, which is (Malyshev, 1937) 18 km north-northwest of the city of Zlatoust; Sokolov and Grigor'ev (1977) say that the distance is 23 km north of Zlatoust. The deposit can hardly have moved 5 km in 40 years, and the distance measures 23 km from Zlatoust to the town of Kusa on the Times Atlas of the World, 4th Edition, plate 47. The deposits are on the left bank of the small Kusa river, a tributary of the Ay, and are on the western slope of the Southern Urals in the highly industrialized area that is located between some 55 km south of Chelyabinsk and about 80 km north of Nishniytagil'.

The actual deposits are confined to a highly altered gabbroic-amphibolitic intrusion that strikes about N40°-50°NE and dips generally steeply east (70°-80°E) but may stand essentially vertically. This intrusion extends to the northeast for about 15 km; in the vicinity of the deposits, the width of the intrusion is between 0.3 and 0.8 km (Malyshev says 1 km maximum).

On the northwest, the gabbro-amphibolite is bordered by Proterozic dolomites. It was into these rocks that the gabbro-amphibolite was intruded. The rocks that lay to the southeast of the gabbro-amphibolite when it entered the area apparently also were Proterozoic in age and consisted of mica-quartz schists and quartzites. At present the southeast border area is occupied by gneissose granites that have a width, in the vicinity of the ore-bearing gabbro-amphibolite,of about 2 km. The effect of this intrusion of granite on the gabbro-amphibolite will be discussed below.

Between the gneissic granite and the gabbro-amphibolite is a contact zone of what Malyshev terms biotite-actinolite-chlorite schists that is about 10 to 20 m thick.

Originally the gabbro-amphibolite appears to have been composed mainly of gabbro or gabbro-like rocks that included lesser quantities of anorthosites, pyroxenites, and gabbro pegmatites, plus the ore bodies that are entirely in the gabbro-amphibolite. In addition to the contact zone between gneissic granite and gabbro-amphibolite, small granite dikes cut through the host rock of the ores as do veins of quartz, aplite, and pegmatite; these veins normally are near the contact zone with the granite.

Malyshev reports that the silica content of the gabbro-amphibolite decreases from the outer margins of the massif toward its center. After, or coincident with the intrusion of the granite body into the host Proterozic sediments to the southeast of the gabbro-amphibolite body, the area was subjected to considerable tectonic activity. These tectonic movements sheared the gabbro-amphibolites and caused considerable faulting that seriously disturbed the gabbro and its contained ores, already present at this time. The major fractures strike about west-northwest, displacing the ore veins by as much as 75 m. These fault-shifts apparently account for the uncertainty as to the number of ore veins in the gabbro. Malyshev says there are five of them and Sokolov and Grigor'ev say there are four; this difference appears to arise from how the faulted veins are connected together by the ore geologists studying the deposit. Since Maylshev' interpretation was based in considerable part on diamond drilling, that of Sokolov and Grigor'ev produced after much more was known of the geology of the deposit probably is more nearly correct.

The contacts between the ore and the metamorphosed gabbro-amphibolite are normally sharp, although the ore bodies are rimmed by zones of garnet, disseminated titaniferous magnetite, chlorite, and minor biotite.

The tectonically induced metamorphism of the host intrusive and its ores raised these rocks to the amphibolite grade and produced such rock as epidosites (developed by a process in which the plagioclase was converted to albite and the anorthite molecule so set free was available for the formation of epidote and zoisite and, in many places, chlorite). The result was amphibole-chlorite, and amphibole-garnet rocks, plus mica and chlorite schists. This tectonic activity also changed the granites into gneissic-granites.

The ores themselves are four (or five) layer- or vein-like bodies in which the content of titanomagnetite is quite uniform. In addition to these veins, which are elongated in the direction of the strike of the intrusion (essentially NE-SW), are local segregations of titanomagnetite. The veins of more or less massive titanomagnetite are confined to a member composed of banded metagabbroids in which thin pseudo-layers of non-feldspar and garnet-bearing feldspar amphibolites alternate. On the contrary, the segregated ore masses are principally in the strongly metamorphosed melanocratic pyroxenitic amphibolites.

Myasnikov and Karpova (1969) say that the Kusinsko-(Kopanskaya) intrusion is concordant with the Proterozic beds that underlie it; the intrusion is locallized in the limb of a large anticlinal fold and is cleanly interbedded in that structure. The K-K intrusion is characterized by alternating banding of gabbro, amphibolite, and massive and disseminated ore, giving an impression of stratification to the intrusion with the ore, either massive or disseminated, being in definite stratigraphic positions in the massif. Metamorphism has caused recrystallization of the ore, considerably changing both textures and structures in that material and producing minerals not originally in the deposit. High-temperature solutions appear to have been forced through the ore masses during the metamorphic activity.

The bodies of massive ores are from 180 to 2500 m long and have widths that normally are 2 to 3 m but may be as much as 8. The post-ore faulting is of the thrust variety with vertical displacements of 70 to 75 m. A lesser number of overthrust faults displace the ores by as much as 25 m vertically.

In the upper levels of the deposit, the ore veins dip steeply to the southeast; locally the dips may be vertical. At depth, the dip gradually changed to from southeast to southwest and gradually to nearly horizontal and finally to gently northeast. These changes in dip, and other variations in morphology of the veins seem to be related to changes in the altitude of the gneissic structures of the adjoining amphiboles. Both structural changes, therefore, are the result of plastic deformation. In places, the course of the veins are sinuously curved rather than being straight, again directly connected with the plastic flow on the host amphibolites. It appears more probable that the titanomagnetite ores were in place before the metamorphism-induced plastic flow occurred. The fractures, which later became the ore-containing structures, probably were developed prior to the intrusion of the granite body and metamorphism that accompanied it. The ores, therefore, are unrelated to the granite intrusion except as it induced changes in the structures of the ore-containing gabbros.

Sokolov and Grigor'ev report that they recognize two main types of ores in the Kusinsk deposits. The first of these is the uniform type - ores that in 90 to 95 per cent of their occurrences are made up of titanomagnetite (a strictly homogeneous solid solution of ulvöspinel $[(Fe(Fe^{2}Ti^{4})O_{4}]$ and/or ilmenite, magnetite-ilmenite [ilmenite with intergrowths (normally regularly distributed) of magnetite], and magnetite; the last two minerals are considered by these authors to be developed in the metamorphic cycle appreciably after the end of the intrusion of the gabbro. In addition to the minerals just mentioned, the ores of this type contain minor, irregularly distributed amounts of högbomite [$(Mg,Fe^2)_2(Al,Ti)_5O_{10}$], spinels, corundum, and rutile. Trace quantities of chlorite, biotite, garnet, hematite, pyrite, pyrrhotite, carbonates, and quartz have been found in these ores; most of these last probably came from the residual fluids of the Ti-Fe-rich melt. Ores of this type are foliated and jointed as a result of the tectonism associated with the granite intrusion; the joints are known to contain chlorite, phologopite, biotite, garnet, tourmaline, quartz, calcite, and sulfides. The appreciable content of what probably were minerals deposited from water-rich solutions in the various late fractures indicates that not only did the granites cause plastic deformation in the gabbros and their included ore seams but also added these water-carried minerals at a considerable time after the ore had been emplaced.

The second type of ore, the segregated variety, is composed of the same ore minerals as the uniform ores, plus minor amounts of various amphiboles (most common) and garnet, chlorite, plagioclases, sulfides, titanite, and traces of apatite.

A further complication of the ore mineralization appears (Sokolov and Grigor'ev) to have occurred when the gabbros (and more mafic rocks) were converted to amphibolites (and melanocratic pyroxenitic amphibolites). Seemingly, recrystallization of the primary titanomagnetite broke down the solid solutions of which these minerals were composed. This produced a coarsening of the ilmenite inclusions that transformed the lamellae into isolated grains. Further, the grains of titanomagnetite were converted into smaller crystals, and, around the borders of these, newly formed pure ilmenite and magnetite were depoisted. Another effect of the metamorphism was the corrosion of grains of titanomagnetite by the metamorphically formed amphibole and garnet. In the course of this process, the magnetite was redeposited but the ilmenite intergrowths were retained, causing a segregation of ilmenite from magnetite. This metamorphic conversion of the titanomagnetite produced a new type of magnetite-ilmenite ore that was much more susceptible to separation into magnetite and ilmenite concentrates than was the primary ore. The magnetite and ilmenite grains of this metamorphic ore type are of essentially the same size (0.2 to 0.3 mm).

Malyshev says that the order in which the magnetite and ilmenite crystallized from the primary melt depended on the relative quantities of the two in the melt in question. Thus, if the ore body contains under 20 per cent ilmenite, then magnetite was the first of the two to crystallize. If the melt had more than 35 per cent ilmenite, the formation of ilmenite came first. In ores in which the ilmenite content ranged between 20 and 35 per cent (and that of magnetite between 80 and 65 per cent) the two minerals were deposited at the same time, and the one is mutally included in the other.

Malyshev also states that the country rocks adjoining the ores may contain disseminated titanomagnetite; the nearer the disseminations to the massive ore, the greater the quantity of the two minerals in the disseminated form. Where the ore-mineral grains are few, they are xenopmorphic and they are contained in the interstices among silicate grains. If one of these segregations, however, is quite dense (40 to 60 per cent of ore minerals), the metallic minerals are idiomorphic.

One explanation for the concentrations of the ore minerals in the various ore seams and segregations is that of Malyshev. He considers that the ore minerals were filter-pressed out of the original gabbro melt, the Fe-Ti-rich melt remaining molten until the gabbro was at least largely crystalline; then it was forced in large part into fractures developed in the gabbro to form the ore seams. The molten ore material that did not reach the fractures crystallized as segregations or isolated disseminations.

Some objections can be raised to this explanation. In the first place, it appears certain that the ore seams were in place before the gabbros were metamorphosed. The fractures that contain the ores probably were in existence before the granite body was intruded, the gneissosity or schistosity of the amphibolites was produced by the intrusion of the granite, and the morphology of the ore seams conforms to that of the amphibolites. Thus, both metamorphosed gabbros and ore seams were deformed at the same time and after the ore melts had been introduced into fractures in the gabbro. Thus, the fractures that contain ore were not developed during metamorphism but in previous tectonic activity.

As a second theory for the gensis of the Kusinsk ores, it can be suggested that the gabbros were introduced during middle Precambrian tectonism. The Fe-Ti-rich melt separated from the gabbro melt before that magma reached its final resting place. After the magma had solidified, it was fractured, reproducing essentially straight NE-SW fractures. After, probably shortly after, the fracturing of the gabbro, the ore melt was introduced into the fractures in the gabbro to produce the primary ores seams. The ores later were metamorphosed as has been described above at much the same time that their structures were being changed from straight to arc-shaped lines. The ore segregations in the general area of the ore seams were residual masses of Fe-Ti-rich melt that had been unable to work their way downward to join the pool of molten material that later was intruded (in the molten state) to occupy fractures that thus became the primary ore seams. The disseminations of minor amounts of titanomagnetite in the amphibolitized gabbro were residual droplets of Fe-Ti-rich melt that neither joined the major pool of such molten material nor accumulated in pods large enough to form the segregations.

In either event (Malyshev's theory or the one proposed here), the ores were produced and introduced in Proterozoic time, probably late enough that the ores can be considered late Precambrian in their primary character (as are so many similar deposits throughout the world). The later metamorphism, induced by the introduction of the granite magma, well may have occurred also in late Precambrian time.

Also, in either event, the ores should be classified as Magmatic-3b, that is, late separation-late solidification of metal-oxygen rich melts. Of course, the category Metamorphic-C must be added since much of the actual composition of the ore minerals, if not the total metal composition of the ores was appreciably changed by this metamorphism.

CHIATURA, GEORGIAN SSR

Early Tertiary	Manganese	Sedimentary-A1,2,4, plus B2 or Sedimentary-A3,4, plus B2

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Notes

The deposits of the Chiatura district are centered around 42°19'N, 43°18'E on the south slope of the Greater Caucasus about 150 km N60°W from Tbilisi, the capital of the Georgian SSR. The reserves of the Chiatura district, although appreciably smaller than those of Nikopol', are second largest in the USSR and probably also so for the rest of the continental world.

The rocks of the Chiatura district cover a much wider range in ages than is true at Nikopol'; these rocks include: (1) Precambrian and lower Paleozoic granitoids, gabbroids, and crystalline schists; (2) lower Jurassic sandstones and limestones and Bajocian (middle Jurassic) porphyritic rocks; (3) Cretaceous terrigenous carbonate deposits; and (4) Tertiary Paleocene to Sarmatian (middle Miocene) clastic sediments. As is true at Nikopol', the manganiferous beds are all in the lower Oligocene and were transgressively deposited on the older rocks. In the Chiatura area, the manganese deposits lie on upper Cretaceous limestones that dip to the east at low angles; at the northern margins of the area, the beds containing manganese dip at much higher angles. In certain locations, faults displace the Mn-bearing beds by as much as several meters. The manganese beds that have been preserved from erosion normally rise above the valley levels in hills of moderate elevations. The various sub-areas of the deposit are separated from each other by stream valleys, the major valley running from northeast to southwest with (from southwest to northeast) the individual erosional-hill ore areas on the northwest valley side being: Rgani, Bunikauri-Seda-Rgani, Tabagrebi, Mgvimevi, Darkveti, and Sareki. The Darkveti-Mgvimevi complex is by far the largest. On the southeast valley side, the ore bodies form a continuous hill mass and are (from southwest to northeast) Perevisi-Merevi, Shukruti, Itkhvisi, and Pasieti. The ore horizon was gently deformed, probably in the Pliocene, into low amplitude undulations and low-dipping brachyfolds.

The rocks of Oligocene age below the ore-bearing beds lie unconformably on the eroded surface of upper Cretaceous limestones; these under-ore deposits are (from bottom to top): 0.15 to 0.50 m of basal conglomerate; then as much (to the east and northeast) as 30 m of quartz-arkose micaceous sandstones. The ore horizon itself is made up of ore beds that alternate with intercalations of sand and clay (which the Soviets consider to be opoka-like). These ore-bearing horizons may contain from 3 to 18 ore beds and may be as much as 25 m thick in such areas as Itkhvisi and Darkveti. The ore beds (or lenses) range in thickness from 1 to 50 cm and may extend along strike for as much or more than 200 m, and the total thickness of the ore horizon ranges from 0 to 14 m with the average being 4.2 m. The sandy-clayey intercalations differ irregularly in number and thickness with the number of beds of this type generally increasing to the northeast. On the contrary, the thickness of the ore beds proper becomes greater toward the west (Varentsov and Rakhmanov).

Normally, the ore horizon is made up of two parts: (1) the lower which contains most of the mineable ore and (2) the upper that includes some ore minerals but is of poor grade. Between these two parts is a ferruginous layer that is made up of iron hydroxides near the surface and glauconite- and chlorite-type minerals at depth. In addition to quartz and feldspar fragments, the sandstones that contain the ores include considerable amounts of rock fragments. The formation also is reported to contain glauconitic tuffs and tuffites, opokas, spongiolites, and siliceous-zeolitic tuffites - evidence of volcanic activity that is lacking at Nikopol'.

In the rock volumes nearest the surface, the oxide ores are mainly polypermanganites and pyrolusite; where the seas were somewhat deeper, the ores are largely manganite; this type of ore being abundant in the Seda-Rgani, North Perevisi, Shukruti, and Mgvimevi basins. In the deepest parts of the manganiferous basin of Chiatura, carbonate ore is present as nodules and oolites and cement around sandstone particles. The most valuable portions of the ore deposit are in the central and, to a lesser extent, in the western parts of the district with the oxide-ore beds growing thinner and finally pinching out in the eastern and northern directions.

Primary oxide ore is present only in small part of the Chiatura deposit, mainly Perevisi and Rgani, with such ore being closely associated nodular carbonate and oxide ores. Another peculiarity is that in some places, such as Mgvimevi, Darkveti, and Itkhvisi, the ore beds in the lower part of the formation are composed completely of carbonate ore, but the overlying beds are composed of primary oxide ore. This arrangement suggests that the control over which type of ore is now present was not due to supergene reactions (under which the upper layers would have been oxidized before the lower) but to the conditions of primary deposition. Where primary carbonate ores outcrop at the surface, however, they are oxidized by ground water passing through them.

Parenthetically, Varentsov and Rakhmanov say that, in the western portion of the deposit, initially an area of shallow water, *situated not far from the metalliferous parent rocks*, the ore-bearing beds are overlain by argillaceous and spongiolitic sandstones that are as much as 30 m thick. To the east and northeast, the manganese-bearing rocks and the sandstones both grade into clay. Further, these two beds are, in this area, separated by 10 m of clay that is similar to that of the Maikop formation (middle Oligocene). The thickness of the two clays is about 20 m, about 10 m for each type. The principal clay mineral is montmorillonite with which hydromica is intergrown. The beds above the ore and clay beds are siliceous-zeolitic tuffites and spongiolites (spongolites?) that, to the east grade into the Maikop clays. The Oligocene deposits have an average thickness of about 110 m.

Over 80 per cent of the Chiatura ore reserves are primary sedimentary manganese ores (oxides and carbonates). About 18 per cent of the reserves are oxide ores developed by supergene processes. In the vicinity of the basalt-like dikes in the Perevisi sub-area, local examples of metamorphosed ores are present. A fourth variety of ore (known as infiltration ore) is made up of manganese hydroxide; in this ore type, manganese hydroxides impregnate sandstone and chert beds that lie between zone of carbonate and manganite ores that are in the process of being oxidized. In a few widely separated locations, deposits of pyrolusite and psilomelane occur as cement or definite, but isolated, intercalations in the sandstones under the ore beds.

The primary oxide ores dominate in the western and central parts of the Chiatura district. Varentsov and Rakhmanov place 47 per cent of these ores in the oolitic category, of which there are two varieties: (1) hard pyrolusite and psilomelane oolites and (2) rather soft oolites made up of brown manganite (brown belta) and of black pyrolusite (black belta). These oolites range in size from fractions of a millimeter to as much as 20 mm with the average being between 2 and 5 mm. These oolites are cemented by siliceous material that contains fragments of quartz and feldspar, of sponge spicules, and of plant remains. In the untreated ore, the Mn-content ranges from a few per cent to 35 per cent; silica amounts to 25 to 55 per cent of this ore.

The two varieties of belta ore contain manganitic and pyrolusitic materials, and the two differ somwhat in chemical composition and contain 45 to 52 per cent Mn; 0.7 to 1.0 per cent Fe; 0.1 to 0.22 per cent P, and 7 to 12 per cent SiO_2.

A third, and appreciably less common oolite ore is a massive, black material (plasti ore) that normally appears in well-defined interbeds with the principal minerals being pyrolusite and psilomelane and the composition being 45 to 58 per cent Mn; 0.5 to 1.0 per cent Fe; 0.10 to 0.18 per cent P; and 3 to 18 per cent SiO_2.

A fourth and minor type is satskhrili ore, and it is an oolitic type but contains fragments of the plasti type, suggesting that it is appreciably younger than the plasti. It is found in interbeds that alternate with beds of the other ore types. If the cement in the common ollitic ore is high in silica, this harder material is called zhgali. Locally, the cement of the oolitic ores may be carbonate instead of silica; such ore is known as mtsvari. As this ore type is located near hydrothermal carbonate veins (Varentsov and Rakhmanov), the carbonate (apparently largely calcite) is considered to have had the same manner of genesis. These hydrothermal veins are most abundant in the Rgani and Perevisi sub-areas.

The carbonate manganese ore makes up 39 per cent of the total reserves of the Chiatura district (Varentsov and Rakhmanov) and is located mainly in the northeastern part and in the upper portion of the ore horizon. The ore layers alternate with siliceous and carbonate rocks; the thickness of the portion of the ore horizon that contains these carbonate ore layers is from 0.5 to 2 m. These authors distinguish two types of ore: (1) oolitic which contains manganese carbonate in that form and (2) a mixed ore that includes both manganese-carbonate oolites and massive fine-grained material. These oolites are essentially of the same form as those of the oxide ores. The principal minerals that have been identified are mangano-calcite, calcian-rhodochrosite, calcite, opal, and minor amounts of barite. The chemical composition of this ore type is: Mn, 20 to 30 per cent; Fe, 2 to 5 per cent; SiO_2, 5 to 40 per cent; CaO, 10 to 35 per cent; CO_2, 20 to 32 per cent.

Somewhat less than 20 per cent of the Chiatura ores is made up of oxidized ores produced by secondary (supergene) processes and was developed where carbonate ores were exposed at the surface. These ores are composed of hydrated manganese oxides and may contain minor amounts of iron hydroxides, opal, and relics of the primary ores. The chemical composition is: Mn, 30 to 35 per cent; Fe, 2 to 5 per cent; SiO_2, 8 to 35 per cent. The brown belta ores are considered by Varentsov and Rakhmanov definitely to be primary, but the black belta ores, which are derived from the brown [i.e., pyrolusite produced from the oxidation of manganite or MnO_2 produced from $MnO(OH)$], are, strictly speaking, oxidized ores, although Varentsov and Rakhmanov put them in both the primary and oxidized categories.

The simplicity of the Chiatura structure appears to make the understanding of the genesis of the formation of the ore relatively simple. This is, however, more apparent than real as it still remains uncertain if the ores were formed by the sedimentation of material eroded from rocks overlying the Georgian block or from deposition for volcanic exhalations that reached the sea floor in the Chiatura area during early Oligocene time. In either event, it seems certain that the ores are part of a series of sedimentary formations accumulated in a coastal zone of a shallow water basin in the early portion of Oligocene time. This deposition occurred in the marginal bay-like part of a basin that widened outward to the east and northeast into much more open water.

Betekhtin (1964), Avalini (1964), Strakhov, and others (1968), and Ikoshvili, 1971 believe that the manganese (and the associated clastic material) came from lower Jurassic arkosic strata, Bajocian volcanic rocks, and crystalline rocks of the Dzirul. On the contrary, Dzotsenidze (1980) and Makharadze and Chkeidze (1971) think that the manganese was supplied by hydrothermal solutions that came from volcanic hearths in the Adzharo-Trialeti geosyncline that were located south and west-southwest of the basin that now contains the manganese ores.

Varentsov and Rakhmanov state that the well-defined pattern of the ore horizon mineralization is in good agreement with the probable variations in the depositional environment. They claim that the oxide ores (pyrolusite and psilomelane) were deposited in shallow water and were successively replaced by manganite and carbonate ores that formed in deeper waters. Thus, it would appear that, during ore deposition, it is their opinion that the basin gradually sank so that conditions favoring hydroxide and carbonate formation encroached on the oxide ores formed earlier, converting those ores to oxide types. Certainly, as the manganese-bearing beds are followed seaward, the ore horizon becomes thicker and more abundant in the intercalated beds of terrigenous and siliceous materials. Thus, Varentsov and Rakhmanov seem to believe that the carbonate ores were formed during diagenesis when the primary Mn-rich materials were converted to carbonate ores by CO_2 that resulted from the decomposition of organic matter in an environment that was capable of converting carbon-hydrogen and carbon-hydrogen-oxygen molecules to carbon dioxide. These authors consider that the formation of the Chiatura deposit was essentially the same as that that produced the Nikopol' ores. This seems almost certain, but, because there is some doubt as to the manner in which the Nikopol' ores were developed, doubt also still remains as to the genesis of those at Chiatura.

In fact, some question arises as to whether or not the decomposition of carbon compounds to carbon dioxide actually was necessary for ores being deposited in a sheltered bay in which carbonate ion content well may have been high.

It seems worth considering whether or not the Chiatura ores, as the Nikopol' ores, may have been produced by the extrusion on the sea floor of volcanic exhalations high in manganese and carbonate ions and low in those of iron (as has been discussed in the description of the Nikopol' deposits). It seems more reasonable to suppose that the manganese carbonate ores deposited in deeper waters in both of these manganese deposits was the primary mineral and that the oxide minerals nearer shore resulted from oxidation of the primary manganese in the carbonates either before or after they had been deposited as primary sediments. In short, if the oxidation of the manganese occurred before that element had been precipitated, the manganese oxide ores would be primary; if after the manganese had been laid down as carbonate, the manganese oxides would be diagenetic or perhaps even the result, after lithification, of ground water acting on the primary sedimentary manganese carbonate or manganite.

In places where manganese-carbonate ores overlie oxidized ores, the suggestion has been made that the environmental conditions (Eh) changed sufficiently from early to later in the depositional cycle for conditions favoring oxide ores to have obtained when the lower beds were being laid down and for the later conditions to have had a lower Eh than earlier obtained in the area.

If the ore element (manganese) was derived from the adjacent land surfaces, by weathering and surface transportation, the deposits should be categorized as Sedimentary-A1, 2, 4, that is, inorganic and organic sedimentary reactions, plus diagenesis, plus ground water circulation (supergene processes - B2). If the manganese was derived from volcanic exhalations, the Chiatura ores should be categorized as Sedimentary-A3, plus probable diagenesis (A4) and in rocks by ground water circulation (supergene processes - B2).

TYRNYAUZ, KABARDINO-BALKARSKAYA ASSR

Middle Tertiary	Molybdenum, Tungsten Copper	Hypothermal-1 (Hornfels Skarns) Hypothermal-2 (Limestone Skarns)

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Notes

The major molybdenum-tungsten deposit of Tyrnyauz is located on the northern slope of the Main (Greater) Caucasus range at 43°25'N,42°55'E and 150 km west-northwest of the city of Ordzhonikidze. Tyrnyauz is in the Kabardino-Balkarskaya A.A.S.R.

Tyrnyauz was discovered in 1934 and has been intensively studied since; it is contained in the Pshekish-Tyrnyauz tectonic zone that has a WNW strike. This zone has been considered to be the southern boundary of the epi-Hercynian Scythian platform; more detailed work, however, indicates that the entire central Caucasian horst-anticline is involved in the epi-Hercynian Scythian platform, thus placing the Tyrnyauz ore body on the marginal part of this structure.

During the Alpine tectonic stage, this area was involved in the orogenic uplift of the Main Caucasus. This uplift is largely confined to the pre-Jurassic basement and is directly related to the area in which the junction of the Pshekish-Tyrnyauz suture with the transversal structure took place.

Of great importance in the Tyrnyauz region is the upper Devonian to Carboniferous sequence of sedimentary-volcanic rocks that compose the Hercynian eugeosynclinal complex of the Fore-Range subzone. The rocks include the following series: phyllite; extrusive (splities, diabases, porphyrites, and quartz-porphyries); tuffaceous-sedimentary (tuffs, tuffaceous sandstones, containing layers of tuffites and siliceous schists; carbonate; and sandstone-shale with intercalations of conglomerates, gritstones, and limestones. These formations have been compressed between salients of crystalline schists that have undergone metamorphism and granitization to different degrees. These crystalline schists belong to adjacent tectonic zones, the Main Range and the Bechasinian. These salients are separated from the tectonic zones by the Northern and Southern regional faults. These Paleozoic formations are confined, therefore, to narrow tectonic blocks and are overlain unconformably there by middle to upper Carboniferous and Permian sandstones, conglomerates, silts, and shales and locally the overlying rocks include conglomerates, sandstones, and shales of the lower Jurassic.

The Tyrnyauz region has been intruded by a considerable variety of igneous rocks that were introduced over a long period of geologic time. The oldest of these are ultramafics of Proterozic to Paleozoic (?) age and are present along the fault zones; most of them now are serpentinized pyroxenites. In the middle Paleozoic, the region was invaded by quartz-diorite porphyrites and trondhjemites (plagio-granites); these are concordant bodies. In the Cenozoic, the oldest intrusives are the leucogranites (about 20 m.y. old) and the appreciably younger El'dzhurta porphyritic biotite granites and the even younger rhyolites. The Cenozoic igneous rocks are confined to the area of the Tyrnyauz ore bodies and take the forms of small, complex massifs, stocks, and dikes. The Samolyet and the Pauk are two of the largest and outcrop on the northern and southern slopes of the Ullu-Tyrnyauz range, respectively. The mineral compositions of these leococratic granites are such that they have been identified as granite porphyries, granite aplites, and labradorite aplites. Where dikes of these rocks penetrate the skarn marbles, they are more mafic. In most places, these leucocratic granites cut the hornfelses. The Pauk rocks have skarn fringes 1 to 2 m thick, and these rocks, in turn, are cut by garnet-pyroxene and molybdenite-quartz veins. The Samolyet leucogranitoids cut the scheelite-bearing skarns.

The appreciably younger El'dzhurta granitoids (1.8 to 2.0 m.y. old) are produced a large massif that dips south and southwest at between 20° and 25°. They are much more uniform in composition than the leucocratic granitoids, being designated as porphyritic biotite granites. Where these granites are in contact with the marbles, they become considerably more mafic and are classed as granodiorites, diorites, labradorite aplites, and plagioclasites. Between these rocks and the marbles are skarn fringes that contain vesuvianite and molybdenite veins. The ore bodies in the Tyrnyauz region are cut by these granites as are the host-rock hornfelses.

The still younger rhyolites, 1.5 m.y. old, are in three moderately large necks that form a more or less north-south chain and are accompanied by approximately east-west dikes that may be as much as 10-m thick. The chemical composition of the rhyolites is much the same as that of the El'dzhurta granites. Still younger than the rhyolites are vitrophyre and basalt dikes (Pokalov, 1977).

The non-calcareous host rocks of the Tyrnyauz igneous massifs and ore bodies are the result of contact metamorphism that converted them to hornblende hornfelses. In the outer portions of these hornfelses (on the northern slope), the hornfelses are albite-epidote varieties. The limestones in the same general rock volumes have been changed to marbles. The exact composition of the non-calcareous hornfelses depends on their original compositon and degree of metamorphism, producing a range of hornfelses of biotite-, biotite-amphibole, cordierite-, and andalusite, pyroxene-, and plagioclase- types.

Structurally, the Tyrnyauz ore area was developed over a long time period. The basement is the early Hercynian fore-range graben; this graben was separated from the adjacent, uplifted basement by the Northern and Southern regional faults. During the late Hercynian and again during the Alpine orogenies, narrow grabens were formed within the fault zone between these two regional faults; here, upper Paleozoic and Jurassic rocks were saved from erosion.

At the present time, the sedimentary-volcanic complex of early Hercynian age dips steeply and has been made highly complex by faults of several types: (1) north-south faults along the contacts of the various lithological units; (2) the Gitche-Tyrnyauz overthrust; (3) diagonal faults, NW-SE-trending; (4) reverse faults and strike slip faults; and (5) the Central fault. This last fault is particularly important as it divides the ore field into northern and southern parts with the southern side having been upthrust from 1.5 to 2 km. Thus different rocks were eroded on each side of the fault and the resulting asymmetric pattern of host rock (and therefore of ore and contact-metamorphism) were produced. Northern and Southern faults parallel the Central.

The structure of the late Paleozoic rocks is that of a 50° to 60°S-dipping monocline that is complicated by a system of reverse faults that are conjugate withthe Northern fault and the interformational shears [type (1)] above. The down-dropping of the Paleozoic beds along these faults controlled the position of the early Jurassic sediments, a situation further confused by their folding into a syncline cut by north-south faults.

The mineralization in the Tyrnyauz ore field is complex although it does have a rather asymmetrical zoning. In the central part of the field are the Mo-W ores. To the southeast and northwest of the central ore area are skarns carrying appreciable amounts of sulfides in such locations as Tyutyu, Lesser Mukulan, and the Northern tin-bearing skarns. Still farther to the northwest, on the far side of the Central fault are skarns and quartz veins in which are low-grade mineralizations in arsenopyrite; still farther to the northwest are the antimony deposits in the Gitchie-Tyrnyauz mountains. Even farther to the northwest on the outer margin of the ore field are cinnabar and orpiment-realgar showings.

In the vicinity of the Tyrnyauz deposit, the Paleozoic rocks have been deformed into the huge Chatbash syncline with almost vertical limbs that have been further deformed by folds of lesser magnitude. This syncline also has been cut by WNW-striking faults, the Central, the Northern, and the Southern and by nearly east-west faults that feather off from these WNW faults. The result has been the breaking of the synclinal rocks into tectonic blocks that have been displaced, each relative to the others.

The major fold in the deposit is a fan-like anticline, a minor structure on the major syncline that has a core of uppermost Devonian carbonate rocks. The crest of this fold plunges steeply eastward, and the limbs dip steeply to the north and south. The southern limb of this anticline is the northern limb of an adjacent syncline, in the core of which is sandstone now converted to hornfels; the southern limb of this syncline is made up of uppermost Devonian rocks.

The anticlinal core of uppermost Devonian rocks has been injected by a minor mass of leucocratic granite; between this granite and the limestone is no more than a thin selvage of skarn. The main mass of the skarn in the limestone is at the contact between limestone and the hornfels that surrounds it. The skarn follows the folds of the rocks of the deposit emphasizing the folds and forming a huge bulge at the anticlinal crest.

In the center of the ore field are the main ore bodies of the district with the Mo-W mineralization being confined to the skarns. The hornfelses and the leucocratic granites near to, and north and south of, the skarns contain molybdenum-rich veins with only minor tungsten mineralization. To the northwest and southeast are further skarns, the mineralization in which is mainly sulfides of tin, bismuth, and copper (these are the Malyi Mukulan and Northern skarns). To the north of the Main ore field are gold-quartz veins with arsenopyrite and the gold-bearing skarns of Tyrnyauz-su. To the northwest, the Gitchie-Tyrnyauz range contains lead-antimony ores; beyond these, on the extreme northwest flank of the structure indications of cinnabar ore have been found (Pokalov, 1977).

When the mineral composition of the skarns is studied, it is not difficult to divide them into skarns formed from hornfels and skarn formed from limestone. The skarns after hornfels make up 60 to 80 per cent of the total mass of the ore skarns and those after limestone the remaining 40 to 20 per cent. The main characteristic that distinguishes the skarns after hornfels is the relics of hornblende and biotitized rocks present. Several types of skarns have been formed from the hornfels; these include pyroxene-plagioclase skarns that show strong banding; pyroxene-garnet skarns that form zones among the pyroxene-plagioclase skarns. Boundaries between skarn types are not clear but are readily seen. These hornfels-derived skarns typically contain sphene and much titanium oxide (0.8 to 1.2 per cent).

Skarns derived from marble (limestone) are in direct contact with such rocks; they are distinctly banded and are quite light-colored. In these skarns, the most abundant mineral is white wollastonite in which are lenses and streaks of pink garnet and olive-green idocrase; the wollastonite may include disseminated grains of pyroxene, garnet, and idocrase; in places, the idocrase may form veins or veinlets that cut the aggregates of garnets. These skarns normally contain appreciable relics of marble.

Veinlets commonly extend outward from the skarn bodies and enter the carbonate rocks; the outer zones of these veins are made up principally of wollastonite with minor traces of various minerals disseminated in that mineral. In the wollastonite zones of these veins, some pyroxene always is present; the amount of this mineral increases as the pyroxene zones internal to these veins is approached. The amount of garnet in the veins also increases toward the internal parts of the veins in the skarn. In some skarn veins, the outer borders of wollastonite may be missing.

It follows, therefore, that the garnet skarn zones developed from the zones of pyroxene and wollastonite skarns and the iron index of the garnet increases the more complete the destruction of the wollastonite and pyroxene may have been and the greater the proportion of garnet in the veins. It is held by Pokolov (1977) that the complex banded structure of the skarn bodies is controlled in part by the zoned structure of the replacement zones in the host rocks and in part by the simultaneous production of new replacement zones where the replacing fluids entered along systems of parallel joints.

The entire ore mineralization is definitely later in time of emplacement than the skarn as is evidenced by the uneven distribution on these ore minerals in the skarns and their consistent presence only in veins certainly later than the skarns.

In the hornfels skarn and in the marble, the molybdenum mineralization is not sharply bounded but grades gradually into less than commercial-grade rocks. In the skarn derived from marble, however, the veins are clearly defined. In biotite hornfels, the ore sectors occur in the hanging wall of the Main skarn where it is limited to the northwestern and southwestern flanks; this type of mineralization also is present in the Northern and other outlying skarns.

The molybdenum-tungsten ore bodies are in the Main ore skarn and in the neighboring Northwestern, Mukulan, Central, and Southern skarns as well as in the marble skarns. The major reserves of the Tyrnyauz area are in the Main ore skarn, essentially all of the tungsten being there and most of the molybdenum. The principal skarn mass surrounds the central mass of marble and, in turn is surrounded by biotite hornfels. Most of the skarn is on the crest of this skarn-surrounded marble body, but it also is appreciably developed around the entire marble mass. Skarn also surrounds the minor body of leucocratic granite enclosed in the center of the marble body; a relatively thin zone of marble lies between the marble and its leucocratic granite core.

The Northwestern skarn is a continuation of the Main skarn in that direction and is rather thin; it is located on the anticlinal limb a considerable distance from the crest of the anticline. It is principally mineralized by molybdenum and tungsten but also contains recoverable amounts of copper.

The Central skarn is contained within marble; the ore body is quite small and was produced by the conversion to skarn of a small layer of sandstone and shale.

The Mukulan skarn appears ot be an extension of the Main ore body; the thickest part of the Mukulan ore does not outcrop, and it is cut by a fault about which the ore body is crushed and thin. Below the fault, the skarn is along the contact between marbles and hornfelses; this layer-like body has considerable and even thickness.

The Southern skarn is an altered seam of marble enclosed in biotite hornfels.

The ore minerals in these ore bodies are scheelite, molybdenite, and molybdoscheelite [apparently a mineral intermediate between scheelite and powellite]. The ratio of WO_3 to molybdenum ranges between 4:1 and 8:1.

In the Main skarn body, W and Mo are not evenly distributed; on the flanks of the skarn concentration in the crest of the anticline, particularly the northern one, molybdenum is unusually high; in the thick portion on the anticlinal crest tungsten is quite high relative to molybdenum. As the ore is followed downward, tungsten increases. Further, with depth, the quantity of Mo^{+4} decreases and that of Mo^{+6} goes up. Despite these changes, the total amount of molybdenum remains about the same. This change of molybdenum valence means that more molybdenum is in powellite and less in molybdenite.

In the leucocratic granites, molybdenum is concentrated in the exocontact zones of the granitoids adjacent to the biotite hornfelses in the main but locally near the associated marbles. This results in most irregularly shaped ore bodies. In ore bodies in this environment, the ratio of WO_3 to Mo ranges from 1:4 to 1:8, which means that molybdenum is by far the dominant element of the two.

For the entire process of metallic and gangue mineralization, the earliest mineral is magnetite; this mineral acts as a cement in the brecciated footwall of the Main skarn. All of the scheelite in the skarns (Pek, 1962) was deposited at this time, as was a small portion of the molybdenite as inclusions in the scheelite. An appreciable portion of the molybdenum (in the Mo^{+6} state) was included in the structure of the molybdenum-bearing scheelite, essentially powellite. In the leucocratic granites and the hornfelses (rocks rich in Al^{+3} and Si^{+4}) the early scheelite was accompanied by early quartz veins that were weakly mineralized with molybdenite, and contained inclusions of pyroxene and amphibole. In skarns, scheelite normally is present with pyroxene and molybdenite with grossularite garnet.

The next event after the major scheelite-minor molybdenite mineralization was a major deposition of molybdenite with the richest Mo-bearing volume being in the Main skarn where quartz veins and veinlets were added to the earlier scheelite mineralization. Such stockworks were produced in both hornfelses and skarns. This molybdenite mineralization was followed by fluorite and small amounts of quartz and lesser amounts of pyrite and pyrrhotite. Still younger was the introduction of a melange of sulfides, such as pyrite, arsenopyrite, pyyrhotite, sphalerite, chalcopyrite, marcasite, tetrahedrite-tennantite, stannite, polybasite, native silver and gold, plus quartz, carbonates, and ilvaite [$CaFe^{2}_{2}Fe^{3}(SiO_4)_2(OH)$] that is reported to contain 1 per cent of tin. The last hydrothermal minerals were quartz and opal in veinlets (Pokalov, 1977).

Much disagreement has arisen over the absolute ages of the Tyrnyauz mineralization. Early reports gave an age of 80 to 100 m.y. for the Ek'dzhurta granites and 97 m.y. for the leucocratic granites. Later work has indicated that the true ages are 20 m.y. for the leucocratic granites and 2.0 to 1.8 m.y. for the El'dzhurta granites. The structural movements in the area appear to have taken place episodically throughout the Mesozoic and Cenozoic, so structure cannot be used to select the correct ages for the igneous rocks and the mineralization that definitely is spatially associated with them and very probably was genetically was well, particularly with the leucocratic granites. On this basis, with the priviso that further work well may reverse this suggestion, the ores are here classified as probably Middle Tertiary.

As for the temperature and pressure conditions under which the mineralization took place, they almost certainly were high and great, respectively. The depth at which the ores were emplaced probably were far engougH beneath the surface as it existed at the time of mineralization for the ores to be classified as hydrothermal-1 (in the hornfelses and leucocratic granites) and hydrothermal-2 (in the skarns).

The ore bodies of the Tyrnyauz ore region are unusual, if not unique, in that a considerable fraction of the ore (perhaps 20 to 30 per cent) was emplaced in the Al-Si-rich skarn rocks; these Al-Si-rich skarn rocks constitute 60 to 80 per cent of the total skarn volume, but they are appreciably less converted to ore than the Ca-rich skarns. In a somewhat similar geologic situation, the Pine Creek deposit of California has essentially all of its tungsten and molybdenum mineralization in the tactite formed when the metamorphically produced marbles were hydrothermally converted to skarn (or tactite).

PITKARYANTA, KARELIAN ASSR

Late Precambrian	Copper, Zinc, Tin, Iron as Magnetite	Hypothermal-2, minor Mesothermal (?)

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Notes

The town of Pitkyaranta (Pitkäranta), 61°31'N, 31°30'E, lies on the north east side of Lake Ladoga in what has been, since the end of World War II, part of the Karelian ASSR; before that time it was part of Finland and the literature in western European languages ends in 1952. The belt of metamorphosed carbonate rocks in which most, though not all, of the Pitkyaranta ores are contained is wrapped around a series of mantled granite gneiss domes (Eskola, 1952), two minor segments of which have been engulfed in later rapakivi granite. The metamorphosed carbonate-rock belt, not counting the intrusion-eliminated segments, is about 140 km long from the mine farthest northwest (Schwartz) to the one farthest southeast (Lupikko); the straight-line (NW-SE) distance from Schwartz to Lupikko is about 82 km. Along this triply curved belt, the ore fields are, from northwest to southeast: (1) the Old Mining Field [that includes, from northwest to southeast, the Schwartz, Nikolai, Meyer, Omelyanoff (4 mines), Tolvo, Klee (six mines), and Orpervoj], (2) Ristaus and New Mining Field, (3) Hopunvaara, and (4) Lupikko. The metamorphosed carbonate belt consists of two originally carbonate beds that are separated by a bed of amphibolite schist; the ores are essentially confined to the carbonate beds, the lower of which is now a mixture of skarn minerals and original calcite, while the upper is a serpertinized dolomite. All of the Old Field mines are included in the lower carbonate bed, while the Ristaus is in the upper bed. The ores of the New Field are found in both beds, as appears to be true of Hopunvaara and Lupikko as well. In the Old Field, the ores in both the northwestern and southeastern parts of that field, except for the Schwartz mine (where magnetite was produced) provided mainly chalcopyrite and sphalerite; in the central portion cassiterite was the most important ore mineral. The Ristaus mine was a magnetite mine, developed in serpentinized dolomite, the upper of the two metamorphosed carbonate beds. In the New Mining Field, Eskola finds only two mines worthy of mention, Herberz and Valkialampi; both these magnetite mines are in the lower carbonate bed, here a diopside skarn. The mines are located along the southeastern margin of the Pitkäranta mantled gneiss dome, whereas the Old Mining Field was on the southwestern margin. The Hopunvaara field abuts against the rapakivi along the southwestern margin of the Vinberg half-dome and the northwest edge of the Luipikko dome, with the entire carbonate-amphibolite sequence being overturned. The carbonate beds dip northward and westward, and the lower horizon contains only one working (worked?) mine, the small Beck (magnetite) mine, but the unworked (1952) Koskelo ore field lies almost due east of it, within a short distance of the rapakivi of the Vinberg dome. This latter mine had not been opened in Eskola's time but may have been since the Soviets took over. During the emplacement of the magnetite in the Beck mine, the skarn was changed largely to serpentine or micaceous minerals. The Hopunvaara mine is in the upper altered carbonate bed and again was a magnetite mine. Between the Hopunvaara and the rapakivi granite are the Klara mines; here the magnetite ore was in direct contact with rapakivi. Finally, the Lupikko ore field is located about 40 km south of the Hopunvaara quarry; in this area, magnetite mines were located in both carbonate beds.

Although the Pitkyaranta district is famous as being the area in which Eskola's concept of mantled domes was developed (1949, 1952), the ore deposits have no genetic connection with the domes and a spatial.one only in that the ore fields are in the rocks that mantle the domes. The genesis of the ore is completely independent of activity that produced the doming, nor does the ore district include more than a considerable fraction of the domed area. Further, Eskola points out that the ore deposits in the lower of the two carbonate horizons in the mantle rock are markedly different from those in the upper horizon. In the lower of the two horizons, the carbonate rock has been largely converted to cavernous diopside skarn that contains some grossularite-andradite. The cavities in this formation are lined with crystals of black andradite and these, in turn, are covered with ore minerals and calcite and quartz; what carbonate remains is mainly calcite. In the upper horizon, the carbonate material (here dolomite) in all the ore fields has been altered to serpentine. The ore minerals either have replaced serpentine or relict dolomite.

In the four ore fields, the mineralizations are quite different. In the Old Mining field, all the mines are in the lower horizon; in these the alteration is to skarn. In the other three fields, the alteration is so varied as to require a separate description for each field: (1) the Old Mining field already has been characterized by late black andradite and ore minerals in the lower horizon cavernous skarn; (2) Ristaus is in the serpentinized dolomite of the upper horizon with the ore minerals in serpentine and relict dolomite; ore bodies in the New Mining field are in both horizons; (3) Hopunvaara contains rhythmic crystallization in dolomite of metasomatic ore pipes and veins; (4) Lupikko includes fluorite and beryllium-bearing idocrase, plus helvite, mainly in dolomite.

The metals in the ores of the ore fields include iron, zinc, copper, tin, and minor lead, tungsten, and arsenic. Although the main ore minerals, magnetite, cassiterite, chalcopyrite, and sphalerite are found in all fields and deposit types, only the mines of the Old Mining field have yielded copper, tin, and zinc in profitable amounts, with the economic amounts of cassiterite being largely confined to the central portion of the Old Field.

In the Old Mining Field, the Schwartz mine, the one farthest northeast, produced magnetite. Although all mining had been discontinued when Eskola wrote (1952), diamond drilling prior to 1939 demonstrated that some zinc, copper and iron ores remained unmined; what the Russians may have done in the field since 1945 is unknown. The largest of the drilling-determined ore concentrations appears to have been in the vicinity of the 4 Omelyanoff mine, the ore apparently plunging off to the southeast.

In addition to the ore minerals, cassiterite, chalcopyrite, and sphalerite, cubanite is fairly common in the Old Field. Eskola agrees with Trüstedt in thinking that the ores are later than the skarn minerals which the ore minerals brecciate. The skarn minerals also are somewhat altered in the vicinity of the ores. It is not possible, in all places, to demonstrate positively that the ores are always later than the skarn, but neither is there any evidence that the sulfide or oxide ores are contemporaneous with the skarn.

Eskola differentiates between the skarn and gangue minerals, that is, skarn minerals were formed before the ores (in the main at least) whereas, the gangue minerals, such as garnet, epidote, scheelite, calcite, and fluorite, with some aragonite and barite, were formed more or less contemporaneously with the ores.

Cassiterite, according to Törnebohm (1891), occurs in part in the skarn and partly in gneiss or pegmatite in the vicinity of the skarn. The minerals surrounding the cassiterite have been considerably corroded, presumably as a result of the same process that deposited the tin oxide. Where gneiss and pegmatite are tin-bearing, they are redder than normally and often are quite friable. Much or most of the cassiterite was deposited in cavities of generally small size in the host rocks, the cassiterite either filling them entirely or lining their walls; quartz and calcite also are common cavity-filling minerals, as locally is chalcopyrite.

Eskola considers that the occurence of cassiterite in some ways resembles that of this mineral in the Erzgebigge and Cornwall, particularly as to the corrosion of the feldspars and the kaolin-like alteration products of that latter mineral. Topaz, however, is seldom seen in Pitkyaranta, and tourmaline does not occur in either the pegmatite or the skarn; where it is found in large granite stocks, it is as schorl. No lithium minerals are known in the area. The usually dark-brown cassiterite is associated with dark-green, pistacite-rich epidote in well-formed crystals.

The mineral character of the skarn differs appreciably from one end of the Old Mining Field to the other, with a much-uralitized diopside skarn being predominant in the northwest, farther southeast (the Meyer shafts) the skarn is a nearly pure, light diopside. In the Omelyanoff mines the ore is associated with much pegmatite and garnet; in the Klee mines amphiboles are the main skarn minerals, with some pyroxene and fluorite. The last mineral also appears to have been quite common in the Omelyanoff and Meyer mines. The fluorite in the Klee mines occurs as a filling of cracks in the skarn and in a granitic-aplite rock that is similar to the rapakivi aplite in the more northwesterly ore fields.

The Old Mining Field is cut by numerous faults, with which the ore bodies are at least spatially connected, the richest parts of the ore deposits being as shoots in the faults. A great chloritized shear zone follows the hanging wall of the lower skarn zone.

Although the Ristaus mine is only a short distance southeast of the Old Mining Field, it is in the upper skarn zone (as are some of the mines of the New Mining Field, Hopunvaara, and Lupikko). In this skarn zone, the most important change in the carbonate rock has been its conversion to serpentine. The main variety of serpentine in the ore is an earthy-looking material, but outside the ore the most common kind is a green, yellow, or brown serpentine with a waxy luster. Also in the ore are veinlets of fibrous chrysotile. Many inclusions of magnetite are visible in the serpentine; also common are crystals or fragments of dark sphalerite, with crystals of cassiterite, arsenopyrite, pyrite, and scheelite being less abundant. The distribution of these crystals is quite irregular, and the mass lacks bedding or foliation. The Ristaus ore body is a flattened cylinder in shape and pitches 12°SE along the dolomite (upper) carbonate layer. Skarn, as known in the lower carbonate layer, is absent, even where the altered carbonate is in contact with pegmatite; these pegmatite-dolomite contacts lack the diopside-bearing material that appears elsewhere in such situations in the skarn of the lower carbonate (limestone) layer. This pegmatite, however, has been altered, presumably by the metal-depositing solutions, for several tens of centimeters into these bodies, mainly to chlorite; the feldspars apparently have been somewhat kaolinized.

In the New Mining Field, to the northeast of Ristaus, an iron mine (1. Herberz) also is located in the upper carbonate horizon. Nearly everywhere that this rock has not been replaced by serpentine and magnetite, it remains as dolomite. The magnetite is accompanied by some sphalerite but chalcopyrite is absent in nearly all places. This Herberz mine differs further from Ristaus in containing considerable fluorite. It appears that some diopside skarn was developed in the dolomite, although it largely was serpentinized before the deposition of the ore minerals. The iron ore occurs as veins and compact masses with these in almost all instances having a banded structure; the veins may crosscut the layers or may be parallel to them. The masses of pure dolomite that initially were contained in the iron ore in places have been converted to late skarn that contains chrondrdite, green mica, fluorite, tremolite, serpentine, chlorite, idocrase, and garnet; this skarn almost certainly is post ore.

In this New Ore Field, the lower carbonate horizon has been converted to skarn, but it apparently contains only two small mines, 2. Herberz and Valklialampi. What ores these mines contained, Eskola (1952) does not say, but he quotes Trüstedt as saying that the skarn contains elongated drusy cavities that are lined with crystals of black and brown garnet, epidote, green mica, chlorite, white quartz, amethyst, calcite, and fluorite in abundance. Pegmatite bodies are rare in the New Field, but dikes of rapakivi aplite and rapakivi-quartz-porphyry are known from all mines of this field. Trüstedt reported that all contacts of rapakivi with iron ore indicated that the igneous dikes preceded the ores, as the dikes were altered by the ore-forming fluids, with their quartz and feldspar being converted to chlorite and fluorite.

In the Hopunvaara Field, the deposits included occur: (1) in the flat-dipping limestone skarn zone on the southwestern flank of the Vinberg half-dome and (2) in the lower and upper carbonate zones that border the northwestern rim of the Lupikko dome near the rapakivi contact; the deposits in site (1) lie almost directly north of those of site (2). From the Lupikko dome, a rapakivi apophysis is intruded along the stratigraphically upper contact between the limestone and the stratigraphically overlying mica schist of site (2). In the site (2) area, the entire carbonate-amphibolite complex is overturned to the west. In the stratigraphically lower carbonate bed, there is only the small Beck iron mine. To the east of the Beck mine, diamond drilling before World War II located the Koskelo ore field with considerable reserves. The Beck mine is unusual in that, although in it is exposed the lower skarn zone, the ore is of the serpentinite type, with the serpentine having been developed as Eskola says: "exceptionally in the middle zone"; what he means is not clear from his text.

The Hopunvaara quarry is mineralogically famous, having attracted the attention of many geologists from the 1830's on. Its greatest fame, however, is provided with what Trüstedt called ore-pipes. These pipes appear to have been developed through a rather complex process that began with the formation of veins and isolated masses of skarn parallel and oblique to the layering in the dolomite. The skarn bodies, isolated masses and veins, were then brecciated and the margins of the fragments rounded and serpentinized. Fissure veins then developed and expanded by replacement of the dolomite and, in numerous locations, these veins were widened into pipes. These veins replaced dolomite with the skarn and serpentine fragments remaining unaffected. In places, it appears that individual pipes were joined through extensive replacement to produce massive bodies of ore that are surrounded the skarn-serpentine fragments. The narrow, pre-pipe veins are made up of alternating dark bands of mainly magnetite and light bands of idocrase and fluorite. From the margins of the outermost magnetite layers, idocrase crystals protrude outward into the surrounding dolomite. Eskola believed that most of the growth of both veins and pipes was from the center of the veins or pipes outward. Thus, the veins and pipes both are typical of formation by replacement. In most of the pipes, the outmost zone is composed of magnetite, but in one pipe it was fluorite. Locally, the fissure through which the ore fluids moved remained open until near the end of the replacement process, when the fissures were filled by crystals growing inward. The layering appears to have been a result of rhythmic replacement of the dolomite, first by magnetite and then by mainly fluorite and idocrase; the duration of the replacement process determined how many concentric layers were formed in a given vein or pipe.

Between the Hopunvaara quarry and the rapakivi of the Lupikko dome to the east are the two Klara mines; in these mines the ores come into direct contact with the rapakivi. Eskola says that the Klara mines contained no ore pipes but were composed of ore veins and incrustations of magnetite much like those of Hopunvaara with many of them including light layers of idocrase; other light layers are composed only of fluorite. Gray, earthy serpentine of the Ristaus type encloses crystals of sphalerite, magnetite, and light brown cassiterite, these last being of large size. Again, Trüstedt thought that the rapakivi was much older than the ores.

Eskola thinks that only the small Vinberg deposits are similar to the skarn deposits of the lower carbonate horizon; this skarn largely was converted to serpentine or mica in areas where the magnetite-sphalerite ore was deposited.

The Lupikko Ore Field includes all of the western margin of the Lupikko dome south of the Franziska fault, but most of the ore bodies are just north of the point where the north-trending western margin of the dome begins to turn to the southeast. Some iron ore was mined here in the 1850's, but no major production took place until after exploration surveys at the turn of the century. At this time, two large mines were developed in the upper limestone horizon; further mining was carried out from 1919 to 1921. The limestone-ore combination is overlain and, in places, cut by a red tourmaline-bearing pegmatite and by a single dike of rapakivi aplite that exhibits contact effects much like those of the New Mining and Hounvaara Fields. Only one place is known in the lower carbonate horizon where skarn, near the hanging wall of the horizon, is in part replaced by ore.

The ore at Lupikko is quite similar to that in Ristaus in that the main ore mineral is magnetite in dull gray, green, and yellow serpentine; rhythmically banded ore veins with abundant fluorite and idocrase, such as those at Hopunvaara, also are present; phlogopite-fluorite skarn also is present. Much of the fluorite is dark and probably is equal to, or greater than, the amount of idocrase. Chalcopyrite and cassiterite are present in these ores, enclosed in serpentine, but only tin ore appears to have been in recoverable amounts, although Eskola is not specific on this point. In addition to the ever-present serpentine (in places pseudomorphous after diopside), other silicate minerals are such species as talc, green mica, and chlorite with lesser amounts of pyroxene, garnet, and epidote. Scheelite, and calcite and brucite occur here as does minor helvite. As is true at Hopunvaara, chondrodite is present in serpentine as a relic from dolomitic alteration that took place prior to the serpentinization.

In addition to those deposits, the geology of which is summarized above, a few minor deposits were found, mainly by Trüstedt's surveys; most of these were magnetite occurrences, but one, discovered somewhat earlier than Trüstedt's work, at Heposelkais was quite similar to the copper bodies of the Omelyanoff mines. One major outcome of Trüstedt's surveys was the failure to find any ores, or even magnetic anomalies, except along the boundaries between granite-gneiss domes and the overlying carbonate-amphibolite mantles. This relationship can be interpreted only in one way - that the ore fluids came from a source at depth, probably from the rapakivi magma chamber in the late stages of its activity, and moved up the channelways provided by the contacts between the domes and rocks overlying them. The evidence compiled by Trüstedt and Eskola seems definitely to demonstrate that the ore fluids could not have come directly across the rapakivi-carbonate boundary since much of the ore is at some considerable distance from any surface expression of the rapakivi.

Eskola believes that the first interpretation of the genesis of the Pitkyaranta deposits was that of Törnebohm (1891). This author established (in Eskola's opinion) the epigenetic nature of the deposits in the limestones above the domes of granite gneiss, this despite Törnebohm's originally considering that the iron ores were syngegetic. Törnebohm places emphasis on the drusy character of the skarn, a feature not duplicated in the Swedish deposits in carbonate rocks. Törnebohm points out that the skarn body, as a whole, contains a huge number of such irregular and drusy cavities that mostly are elongated in the direction of the bedding. These empty spaces still may be empty or may be filled, in whole or in part, with various species of later minerals, all of which, of course, are not represented in any one cavity. Adjacent to these drusy cavities, the skarn minerals in the solid rock always are coarser and better crystallized than elsewhere, this being true particularly in the case of garnet. Diopside is rare as a cavity mineral, although it is the dominant mineral in the solid skarn; this suggests that the bulk, if not all, of the cavity minerals formed after the development of the skarn minerals proper. The mines in the lower carbonate horizon of the Old Mining Field, contain most of the copper, zinc, and tin; although the mines in the upper limestone are chiefly magnetite mines, even though a spare representation of the metallic minerals is present dispersed in the magnetite.

Eskola states that Brögger compared the Pitkyaranta ore bodies with the contact-metamorphic deposits of the Olso region, although he does not give a reference for this statement nor does he say to what degree Brögger thought the two categories of deposits to be similar. J.H.L. Vogt (again without a reference) is said to have believed that the Pitkyaranta deposits combine features of both the Cornwall and Oslo types.

It was Trüstedt's opinion, to which Eskola generally subscribes, that the Pitkyaranta skarn was formed from limestones similar to those still preserved in the two limestone horizons of the district with the upper one showing greater similarity to its proported parent than the lower. Both of these men considered that the Pitkyaranta deposits (in contrast to the bulk of Finnish ores that range in age from 2250 to 1770 m.y.) are appreciably later with the three rapakivis in the Wiborg area having ages of 1700 to 1660 m.y., 1660 to 1640 m.y., and 1440 ± 15 m.y. and other areas (Vehmaa and Laitila) 1590 and 1570 m.y., respectively. This means that the ores must be some younger than 1450 m.y. (or late Precambrian). Further, Trüstedt gives evidence to support his contention that the ore in Ristaus, 1. Herberz, Hopunvaara, and Lupikko mines show that oldest metallic mineral (magnetite) was formed appreciably after the skarn minerals. The skarn was developed after the quite abundant post-Ladogen pegmatite in the region but before the intrusion of the rapakivi granites. After the pegmatite was introduced the skarn was formed, the rocks of the district were subjected to earth movements that broke up much of the skarn into blocks that were, in this process, rounded and their margins serpentinized. The resulting, rounded blocks are known as skarn augen. Only after all this had been accomplished was the ore deposited. As has been stated above, the close location of much of the ore to actual outcrops of rapakivi and the probable positioning of most of the remainder of the ore not far above rapakivi bodies indicates that the formation of rapakivi magma and of the ore-forming fluids were closely related processes but not as close as Trustedt's concept of the ores being the result of contact metamorphism induced by the intrusion of the rapakivi magmas. The near-surface presence of the rapakivi throughout the district is shown by windows of that rock near Nietjärvi and Vinberg and of veins of rapakivi aplite in the New Mining and Lupikko mining fields.

Since the ores normally are deposited along either the hanging or footwall of the horizon in which they are present, Eskola believes that the ascending ore-forming fluids found the least resistence to upward or lateral movement along the contacts of skarn or serpentinized dolomite with the country rocks - amphibolite, granite gneiss, pegmatite, or rapakivi. This locus of movement also explains the common, though not economically very important, presence of ore minerals in these country rocks. Of the country rocks, the pegmatite appears to have been the least permeable, indicating why the ores are concentrated on the carbonate side of the contacts between pegmatite and one or other of the carbonate horizons.

Eskola believes, as is now considered most reasonable, that the ores were not transported as gases or vapors but by high-temperature, water-rich solutions from which the ores were formed by reactions internal to the ore fluids and by reactions of these ore fluids with the skarn rocks.

The minerals of the Pitkyaranta are all those that, when associated together, are considered to have been formed at quite high temperature. The magnetite can reasonably be classified as hypothermal as also is true of the cassiterite. Not enough is known of the relationships of chalcopyrite and sphalerite to cassiterite nor of the exsolved contents of the one sulfide in the other to say that they were deposited under hypothermal conditions only. Instead, it is not unreasonable to assume that the conditions of deposition range down into the mesothermal category, although the bulk of the two sulfides probably was emplaced in the hypothermal range.

Sakela (1951) indicates that his microscopic work (1951) requires that the deposition of the ore took place in several generations. After the formation of the skarn augen and the serpentinization of the dolomite of the upper horizon, the sequence of depositional events was;(1) apatite, hematite, magnetite, vesuvianite, phlogopite, garnet, helvite, fluorite, Mn-bearing carbonate, calcite, and some sphalerite; (2) brecciation; (3) dark and light cassiterite, scheelite, and arsenopyrite, some topaz, fluorite, and quartz and calcite; (4) minor scheelite, arsenopyrite, löllingite, native silver, molybdenite, sphalerite, pyrite, pyrrhotite, cubanite, and chalcopyrite and a little gold; (5) brecciation; (6) bornite, covellite, digenite, chalcocite, sphalerite, tetrahedrite, galena, bismuthinite, and native bismuth with some calcite, quartz, fluorite, and chlorite. The amount of minerals deposited in stage (6) appears to have been so small as not to require mention of it in the classification assigned to the ores of the deposit. Saksela also presents a second and third series of ore formations, both of which appear to have been so designated because he did not realize that all the ore was younger than the rapakivi granite.

The skarn (appreciably older than the ore) was formed also under hypothermal conditions but, technically, because of its much earlier development, is not included in the classification of the ores.

The Pitkaryanta ores, therefore, are here classified as hypothermal-2, minor mesothermal (?).

The Salmi rapakivi massif (Haapala, 1977), in which the Pitkaryanta deposits are located, appears to have been emplaced in the period from 1650 to 1450 m.y. (Khazov, 1973). The first of the three phases of igneous intrusion was a hornblende granite that contains ovoid megacrysts of K-feldspar. The second phase consists of an equi-granular medium-grained granite with biotite as the main mafic mineral. The third phase is composed of two stocks of fine-grained, porphyritic (locally equigranular) granite, the dark mica of which is lithium-bearing biotite or protolithionite. In this third-phase granite, the accessory minerals include ilmenite, monazite, bastnaesite, cassiterite, and columbite-tantalite (Rub, and others, 1974).

The youngest rocks of the area belong to a Jotian graben formation located to the south of Pitkaryanta; the upper formations in this graben are basalt, porphyrite, and tuff, and the lower parts are sandstone and conglomerate.

Hapaala points out that the most important minerals of the Pitkaryanta ores are magnetite, cassiterite, sphalerite, chalcopyrite, galena, scheelite, and three beryllium minerals - helvite, bertrandite, and phenakite. The tin-beryllium-sulfide mineralization is later than the emplacement of the Salmi rapakivi massif; it was controlled by: (1) association with the rapakivi granite, especially the youngest phases of that intrusion; (2) the stratigraphy of the sedimentary rocks, and (3) faults and tectonic zones of weakness.

The ore deposits of Pitkaryanta occur in the exocontact zone of the rapakivi-granite massif itself, in the neighborhood of the third-stage granites. From what has been seen of the behavior of the contacts and the character of rapakivi windows in the schist, it appears that the rapakivi dips, in the Pitkaryanta area, to the west at low angles. The granite, therefore, probably lies at no great distance below the ore bodies even though many of the ore bodies are 3 to 4 km from the nearest visible contact.

The ore bodies nearly everywhere are located in limestone-dolomite-skarn horizons, especially where these skarn-carbonate beds are along contact with the country rocks - granite gneiss, amphibolite, pegmatite, and rapakivi granite. Pegmatite dikes, in particular, seem to have provided barriers to ascending ore fluids, so that much mineralization is located along pegmatite-skarn contacts.

Further control over the location of ore bodies was due to faults and zones of tectonic weakness. In the Old Mining Field for example, all but one of the old mines was in a chloritized shear zone that parallels the contact of hanging wall with the lower minestone-skarn horizon. The tin ore of Kitelä lies along zones of such tectonic weakness that follow the carbonate-skarn horizons (Khazov, 1973). Again, in the Old Mining Field area, the ore bodies in many places are clearly located where the limestone-skarn horizons are close to faults. The granite gneiss also has been greisenized along the so-called Franziska fault and locally the greisen carries ore-type mineralization. In fact, one small ore body was emplaced at the junction of this fault with another one.

As has been pointed out, most ore geologists studying the Pitkaryanta area have considered the tin-polymetallic mineralization as being connected genetically with the rapakivi granites. Recent studies, however, such as common-lead isochron determinations (Popov, 1975), suggest that the generation of the Pitkaryanta ore field was a complex one. Studies in the late 1960's on analogous, greisen-type Sn-Be-W-Cu-Pb in the Laitaila and Wiborg rapakivi massifs leave little doubt as to the genetic connection between the Pitkaryanta-type ores and the major tin and related ore mineralizations and the late-stage granites of the Salmi massif (Haapala, 1977).

KEMPIRSAI, KAZAKHSTAN SSR

Middle Paleozoic	Chromium	Magmatic-1b, -4 Metamorphic-C

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Notes

The chromite deposits of the Kempirsai ore region are in the major ultramafic massif of that name that is located in the extreme northwestern part of the Kazakhstan SSSR about 110 km ENE of the city of Aktyubinsk and some 80 km slightly east of south of the city of Orsk, just over the border in the RRSFSR. The geographic location of the major chromite in the massif is at 50°36'N,58°15'E with the major chromite bodies being located in the southeastern tip of the massif. Only small bodies of ore shows are present north of the southeastern marginal area of the massif. The total area of the massif (on the surface) is 920 km^2; the maximum width is in the southern part of the ultramafic body and is 31.6 km; it gradually, but irregularly, narrows toward the north to a point about 0.6 km wide; its total length is 82 km. In the latitude of Khromtau (50°15'N,58°22'E), the thickness of the earth's crust down to the Moho is 51 km, and the thickness of the 'basalt' layer is 25 km.

The arrangement of the rocks of the massif are quite different in the northern and southern parts of the massif; in the northern, the ultramafic body lies, largely conformably with the direction of the schistosity on upper Proterozoic sediments that make up the hanging wall and with the lowest Ordovician upper Tremadocian beds in the footwall. The massif in this area is homoclinal, is up to 2.5 km thick, and dips to the west at 40° to 60°. In general, the eastern contact of the massif is tectonic.

In the southern, and commercially important, part of the massif the country rocks are Cambrian gabbro-amphibolites; in other places, the ultramafics are present in the Proterozoic sediments. Between the massif and upper Devonian and lower Carboniferous rocks tectonic contacts exist. From the geologic and geophysical data which have been accumulated over the years, it seems probable that the massif is laccolithic and was intruded between upper Proterozoic and lower Paleozoic beds. In the extreme south, the contact surface of the massif dips south and southeast between 30° and 50° and passes below the amphibolites. Along the eastern margin, the contact is tectonic and dips vertically or steeply east. The thickness of the massif in the southern part may be as much as 16 km but in the western part it is down to 6 to 8 km.

A feeder channel for the southeastern portion of the massif has been recognized by geophysical work; this channel has a steep easterly dip, and it extends below the structure of the Magnitogorsk synclinorium. A channel at the northern end dips steeply (apparently) southward toward the same synclinorium. Absolute-age determinations indicate that the age of the massif is 404 to 380 m.y. or middle Paleozoic.

The surface of the massif in the southern portion has been folded into three arches; Southwestern, Tagashasai, and Southeastern, the last being the area of the Main ore field. Other arches exist, but they are out of the ore zone and can be neglected here. The depressions between the arches have been filled with amphibolites and schists. Although most of the external contacts of the massif are tectonic, places are known, particularly on the southern margin, where the country rocks have been contact metamorphosed with such minerals as garnet, idocrase, pyroxenes, and prehnite having replaced the amphibolites (Pavlov and Grigor'eva, 1977).

The main rocks of the ultramafic are lherzolites, amphibolitic peridotites, prophytitic and banded harzburgites, wehrlites, dunites, enstatite dunites, sulfide-bearing dunites, olivinites, saussuritized and amphibolitized troctolites, pyroxene-amphibole gabbros, anorthosites, chromite dunites, and various types of chromite ores; in short, a typical ultramafic melange of low-silica, high Fe-Mg rocks. Serpentinization has affected all of the rocks named, particularly those rich in olivine. The primary textures have been obliterated in the serpentinization.

In addition to the massive rocks, dikes (or veins) have been developed, both ultramafic and mafic types. The ultramafics are: bronzitites, actinolitites, hornblendites, sulfide-bearing diopsidites, and websterites, and olivinites. Also present in the dikes are: gabbro-diabases, a hypabyssal dike rock that probably has been highly altered (beerbachite), olivine norites, and anorthosites.

The predominant peridotite is massive harzburgite, some of which is porphyritic and is in the less eroded peripheral portions of the massif. The banded harzburgites probably are the most widely developed peridotite.

Massive harzburgites usually are found where erosion of the roof has been greatest, being particularly prominent in the three uplifts previously mentioned; these rocks have large and small grains of orthopyroxene (largely altered to bastite), quite evenly distributed in the serpentinized olivine matrix. The banding in the harzburgites is achieved by alternations of thin bands high and low in pyroxene grains or chains. The principal harzburgite minerals are olivine and orthopyroxene with chrome-spinel as the main accessory. The alteration serpentine is normally lizardite, but antigorite occurs in contact zones with the gabbro-diabases, plus bastite, tremolite, clinochlore, brucite, magnetite in fine powder, talc, and rare phologopite. Percentages by volume are: olivine, 14; serpentine 67.8; orthopyroxene, 2.5; bastite, 7.3; tremolite, 3.6; talc, 1.9; clinochlore, 0.3; spinel, 1.5; magnetite, 1.1. On the assumption that bastite, tremolite, talc, and clinochlore came from the pyroxenes and the serpentinite and magnetite from olivine, the volume percentages of the primary minerals in the banded harzburgite were: olivine, 82.9; orthopyroxene, 15.6; chrome spinel, 1.5. In the massive variety, the percentages similarly derived were: olivine 79.5; orthopyroxene, 19.0; chrome-spinel, 1.5. Locally, a little clinopyroxene (diopside) is present in the harzburgites, although it probably was developed in post-magmatic time in connection with the intrusion of dikes and veins of pyroxenites. The rock textures are hypidiomorphic-granular, and the present textures are presumed to be due to appreciable recrystallization.

Chemically, the harzburgites were high-magnesia ultramafics, all of the primary minerals having been low in iron. The serpentinized rocks contain not more than 25 per cent relic olivine and normally much less; serpentine is 73-98 per cent; chrome-spinel no more than 1.5 per cent; dust (probably magnetite) 1.0 per cent. The lizardite forms a network stringers, randomly oriented.

Enstatite dunites differ from other varieties only in the presence of enstatite (or bastite) in amounts from 1.0 to 10.0 per cent, otherwise they are quite similar to the harzburgites.

The sulfide-bearing dunites are present in any quantity only near the ore-bearing complexes of the Main ore field and are near the bodies of chromite ore. Rocks in the Main ore field are much lower in iron than those in the chromite-ore-poor northern part of the massif; what connection there is between this lack of iron in the south and the presence of chromite ore bodies is not clear.

The uncommon wehrlites appear to have been produced by metasomatism of primary olivine in dunites by diopside. The lherzolites (clinopyroxene-poor) are present in both the ore fields and scattered among the harzburgites - they are not common. Troctolite (olivine-bearing gabbro, forellenstein) occurs on the western margin of the massif (away from ore).

The segregations of chromite ore occur in many parts of the Kempirsai massif and are widely different in their depths beneath the surface. Further, the ore bodies are widely varied in size, and the various ore types in the lenses differ from one ore body to another, the distribution of the ore bodies depends not only on their textures but on mineral composition, internal structure of the massif, and the depth of erosion. The location of arched uplifts is important in exploration as the individual ore fields are confined to these anticlinal structures. The massif contains four ore fields, in which are located 160 ore deposits and mineralized rock volumes (not of ore grade). The northern Batamshinsk uplift contains the Batamshinsk ore field. In the Southwestern and Tagashasai and Stepninsk ore fields are groups of small deposits and prospects. The Main ore field is located in the Southeastern uplift.

The Stepninsk ore bodies are in the zone where the troctolites and harzburgite are in contact; they are of small size, a few 10's of meters long and a few meters thick. They are low quality because the chromite contains no more than 40 per cent Cr_2O_3 and has 30 per cent Al_2O_3. The ores and prospects in the Southwestern and Tagashasai ore fields (in the south) and in the Batamshinsk ore field (in the north) are located in peridotites almost exclusively; only a few minor segregations are in dunites. These bodies are slightly larger than those in the Stepninsk, and their ores are low in chromium and high in aluminum; some chromite concentrate, however, may have 43 to 45 wt per cent of Cr_2O_3.

The Main ore field contains all the high-quality chrome ores; none is aluminous. These ores are arranged in two more or less north-south groups, the Eastern and the Western. These ore zones are crowded with dunite segregations in the peridotites, and the ore segregations are concentrated in these dunites. The dunites normally are sulfide-bearing. This concentration of ores in dunites differs from those of the other fields just mentioned where the ores are in peridotites.

The Western ore zone has a length of 22 km and trends NNE; near its southern end, it joins the Eastern ore zone. The Western ore zone deposits (from south to north) are: Millionnoe, Almaz-Zhemchuzhina, Pervomaisk, Nos. 16, 29, 29a, 31, 39, Khrome-Tausk, and Geofizicheskoe VII, still farther north are the Komsomol'sk No. 4, Geofizicheskoe V and III, 20 years of the KAZAKH SSR, Geofizicheskoe XII, Aleksandrovsk, Iyun'skoe, and a major number of prospects. The ore bodies all dip to the west at from 15° to 75°.

The Eastern ore zone has about the same length as the Western and is more or less parallel to it. The Eastern ore zone deposits (from south to north) are: No. 21, Spornoe, Sputnik, Gigant, Geofizicheskoe II and VI, Solov'evskoe, 40 years of Kazakh SSR, Molodezhnoe, Vraplennoe, plus several prospects. These ore bodies have a dip of 0° to 50°E. In the southern part of the zone, at least the deposits No. 21, Spornoe, Voskhod, and Molodezhnoe pitch to the south; farther to the north, the pitch of the Vkraplennoe deposit (and perhaps of others) is to the north. It also appears that, in the Western ore zone, the same pitch characteristics obtain for the ore bodies there.

Of the ore bodies of the Western ore zone, the following are classified as large: Millionnoe, Permonmaiskoe (also spelled Pervomaisk, see above), Almaz-Zhemchuzhina, 20 years of Kazakh SSR. In the Eastern ore zone, the large bodies are: No. 21, Molodezhnoe and 40 years of Kazakh SSR (these last two are almost immediately adjacent, the one to the other).

The chromite ore bodies consist mainly of lens-like bodies of major size that differ widely in shape, from long and narrow to broader and more compressed. Some vein-like ore bodies are known (as in the "20-years of Kazakstan ore body"). Schlieren-like bodies of chrome ore are quite uncommon but are known in the Geofizicheskoe XII deposit. Individual ore bodies may be as much as 1.5 km along strike down to a few 10's of meters; widths range from 150 m down to a few meters. The number of ore bodies in a given ore deposit differs greatly, from one at the Moldezhnoe, through five at Almaz-Zhemchuzhina, to 40 at the "40-years of Kazakhstan ore body", to nearly 100 at Millionnoe. These ore bodies are separated by dunites, pyroxene dunites, and, less commonly, by periodotites.

The segregations of ore generally have their long dimension aligned in a nearly north-south direction; one deposit (Molodezhnoe) strikes about N45°E and one (Geofizicheskoe) is about equidimensional. In places, the shapes of the ore bodies have been complicated by gentle folding. Essentially all of the ore bodies have been fractured by post-ore movements along about E-W lines; displacements of fractured blocks may be from a few 10's to 300 m.

Almost exclusively, the ore bodies occur in dunites, and their contacts with country rocks are sharp except in a few places where thin zones of alteration (1 to 2 cm) are present; the zones show a depletion in chrome-spinel grains and a decrease in their size.

The ore bodies are quite different both in external shape and in internal structures. Some characteristics, however, seem to be true of all (or most) of these bodies. Both in cross- and longitudinal-sections, the margins of the bodies have a lower density (less segregated chromite) than the more central portions; the margins also show greater differences in structures and grain sizes of the chrome-spinels, a dominance of fine-grained and varied-grain-size ore minerals, and commonly segregations of nodular ores.

Internally, the concentration of chromite segregations is higher than in the margins, ores are more massive, grains size are more uniform and grains commonly are coarse. Structures remain constant over appreciable distances both length- and cross-wise. In the centers of coarse and thick ore bodies, a rough interlayering is provided by variations in grains between medium and normal sizes; banding also results from differences in the density of chromite segregations and in the sizes of the grains that make them up. The banding generally conforms to the attitude of the ore body concerned.

Although the structures (textures) of the ores have considerable variety, they have been generally separated into three main types: (1) segregated, (2) uniform (massive), and (3) nodular. In turn, the segregated ores are divided into three subtypes; (a) densely-segregated, (b) medium-segregated, and (c) rarely segregated. Further, these subtypes can be divided on the basis of grain size into: (i) fine-grained (up to 1 mm), (ii) medium -grained (1 to 3 mm), and (iii) coarse-grained (more than 3 mm). The commercially high-grade ores of the southeastern part of the massif are densely segregated with 70 to 90 per cent of chromite and 30 to 10 per cent of gangue minerals. A still further distinction of the ores on the basis of grain distribution puts them into classes - banded, schlieren-banded, even segregations, and patchy with these, in turn, categorized as eutaxitic (banded) and ataxitic (unbanded). The most common ore types are those that are either varied in grain size or coarse-grained in which the grain sizes range 1 and 2 and 3 and 4 mm, respectively (Pavlov and Grigor'eva, 1977).

Large ore segregations normally contain large- and coarse-grained, densely segragated ores, accompanied by both massive and nodular types. The massive ores, as would be expected, include little gangue material, being made up of medium- and coarse-grained chromite. The nodular ores occur in most of the deposits but they compose far less of the ore volume than do massive and segregated ores. Nodules usually are compressed and are therefore elongated in one direction.

The chemical compositions of the raw chromite ores have an appreciable range; in one deposit; for example, in the Almaz-Zhemchuzhina deposit, the Cr_2O_3 content for massive (uniform) ore is 58 to 59 per cent; for densely segregated ore is 50 to 57 per cent; for medium-segregated ore is 37 to 49 per cent, and for only poorly segregated ore is 28 to 36 per cent. The average Cr_2O_3 content for the entire ore production for the district is 49.05 per cent. Massive ore contains only 2 to 2.5 per cent SiO_2, but poorly segregated ores have 8.1 per cent. The FeO content ranges within narrower limits, from 10 to 15 per cent with an average of 12.5 per cent. The FeO content is, of course, lowest in massive and densely segregated ores.

The primary mineral composition of the ore, before any changes by post-magmatic and hypogene alterations, was essentially nothing but chromite and olivine. Such ore as has been much altered since its primary deposition, however, contains many secondary (about 30) minerals; these include secondary olivine, chrome-diopside, chrome actinolite, uvarovite, serpentine, chrome-chlorites, fuchsite (?), brucite, iron oxides, various sulfides, and several autometamorphic minerals (e.g. magnesite, opal). Sulfides occur only rarely in the ores although they may be in the dunites. The chrome-spinels are quite similar to the accessory chrome-spinels in peri-ore dunites but are quite unlike those in the harzburgites associated with the ores.

For further information on the particular ore bodies in the main ore zone, see Pavlov and Grigor'eva (1977), sepcifically 20 Years of Kazakh SSR, 40 Years of Kazakh SSR, and Molodezhnoe deposits.

Pavlov and Grigor'eva believe that the distribution of the ores, the scale of mineralization, and the quality of the ores was influenced by; (1) the shape and dimensions of the intruded ultramafic massif; (2) the composition and degree of differentiation of the ultramafic rocks; (3) the positions of the feeder channels along which the silicate and ore-silicate magmas entered the Kempirsai system; (4) the presence of the arched folds in the massif favorable to the spread of mineralization; (5) the degree of erosion after the rocks and their contained ores had been solidified.

The lens-shapes of the bulk of the chromite bodies in the dunites and the presence of veins of chromite ore in several of the ultramafics suggest that the chromite was introduced at the same time as the silica-bearing ultramafic magma and that both materials were intruded in the molten state. The molten chromite material initially was almost entirely (in the Main-ore-zone magma, at least) dissolved in the dunite magma. During the crystallization of the dunites, some molten chrome-spinel was trapped in the mesh of crystallizing olivine. The proportion of trapped chromite determined if the rock formed was sufficiently rich to form ore and whether the ore formed was one of the segregated varieties or was sufficiently rich in chromite to form uniform-(massive) or nodular-type ores.

If this explanation is correct, then the bulk of the chromite became immiscible early in the crystallization cycle and crystallized later, in general, but only slightly later than did the accompanying silicate melt. The chromite was not an immiscible melt early enough to form layers such as are present in the Bushveld or the Stillwater complexes, but was separated in time to form the pods of lenses so characteristic of podiform chromite deposits. The distribution of the chromite pods over considerable vertical distances in the massif indicates that the entire (or nearly the entire) portion of the magma classed as dunitic was rich enough in dissolved chromite to allow for the production of such pods in the resulting rock. Thus, the basic classification of the ore is Magmatic-1b, that is early separation, early crystallization. To that category must be added Metaorphic-C to cover the effects of later tectonic metamorphism and Magmatic-4 to account for the deuteric alteration that produced serpentine and other such minerals.

BEREZOVO, KURGAN OBLAST

Late Paleozoic	Gold	Mesothermal to Leptothermal

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Notes

The Berezovo goldfield is at approximately 54°40'N,64°26'E in the Kurgan Oblast in the Eastern Urals and about 100 km south-southwest of the city of Kurgan.

The rocks in the goldfield are greenstones that range in age from lower Cambrian to lower Carboniferous. The four recognized groups are (from bottom to top): (1) Agafurovo group, lower Cambrian to Ordovician and consisting of metamorphosed diabases and porphyrites (used in the sense of porphyrys with plagioclase phenocrysts and not alkali feldspar ones); (2) Iset group, Ordovician to basal Silurian and composed of quartz-chlorite, quartz-sericite, and carbonaceous-cherty slates with intercalated tuffites, sandstones, and limestones; (3) Adu group, uppermost Devonian, mainly diabases, porphyryites, and equivalent tuffs and lavas; (4) Berezovo group, lower Carboniferous and made up of interbedded chart-clay and chert-carbonaceous slates, tuffites, phyllites, tuffaceous sandstones, and tuff breccias. In the goldfield proper, the most common rocks are those of the Berezovo (4) group and the Adu (3) group. The lower two groups are so deeply buried as to have been encountered only in deep drill holes. Within this sequence of volcanic and sedimentary rocks are serpentinized peridotites and pyroxenites in layer-like or laccolithic bodies.

Structurally, the Berezovo ore field is contained in the Eastern Urals anticlinorium and is composed of the Paleozoic greenstones just described. The anticlinorium is bordered by deeply penetrating fractures that are nearly N-S in strike. These fractures controlled the intrusion of older gabbro-peridotite bodies and of Hercynian granitoids. Since the ore field also is bounded by igneous rocks on the north and south as well, it is completely surrounded by the various igneous massifs - Hercynian granitoids on the west and east, gabbro-peridotite on the north and serpentinized ultramafics on the south (Borodaevskaya and Rozkhov, 1977).

The ore field proper is contained in a subordinate syncline that is centrally located in a block that is cut off between more or less N-S striking faults; the axis of this syncline also strikes about N-S and plunges at a low angle to the north. The structure of this Berezovo syncline is made more complex by second-order isoclinal folds that range in strike between NE-SW and N-S. The rocks also are cut by strike-slip faults, the strike of which is more or less E-W and the dip of which is quite steep, by strike faults that strike in a generally north-south direction and also have steep dips, by thrust faults and accompanying (?) shear zones, both of which also have near N-S strikes and dips of as much as 30°W. All this variety of fault structures was developed before the granite-like dikes were intruded; these granite dikes are essentially parallel to the various N-S-striking fractures.

Larger masses of granite and granite-type rocks cut irregularly through the older rocks, igneous and sedimentary, these granite-like masses probably being the sources from which the various dike-filling rocks came.

The (1947) work of Borodaevskaya and Borodaevsky was able to identify five types of dike rocks (listed in order of their injection): (1) plagio-syenite porphyries, (2) lamprophyres, (3) granodiorite porphyries, (4) granite porphyries, and (5) plagio-granite porphyries. The dikes appear to form two main but separate groups with the dikes that compose them being more densely spaced in the southern part of the area and diverging northward. Each of these groups is made up of several clusters of such closely spaced dikes with the distance between individual dikes being 5 to 10 m and that between clusters being from 80 to 100 m or even more. In the two groups of dikes, the dips are to the west in the east group and to the east in the west group. The dikes become less numerous toward the more or less east-west fault that is occupied by gabbro and peridotite; on the contrary, the number of dikes is larger the farther south the dike-group are followed. The more mafic dikes extend for only short distances and are quite thin. The more silic dikes are far longer (up to 9 to 10 km and, unusually as much as 20 km) and appreciably thicker.

The separate character and time of intrusion of the dike-rock types (1), (2), and (3) from those of types (4) and (5) is shown by the presence of the first three types in overthrusts and shear zones and the last two types in strike-slip joints. Thus, not only the rock types differ in the igneous rocks they contain but the ages of formation of the various fault-types must have been appreciably different.

The gold is contained in two types of veins: (1) veins confined to the dikes and (2) veins outside the dikes in the volcano-sedimentary and igneous rocks that surround them - krassyk veins. Thus, all the gold-bearing veins have a direct connection with the dikes which must have acted as channelways for the gold-carrying ore solutions.

The most important of the dike veins are steeply dipping veins that are essentially perpendicular to the walls of the dikes that contain them. Since the dikes are more or less north-south in strike, the veins are appoximately east-west. These veins were designated as "ladder" veins because they are, in section, much like the rungs of a ladder. In plan, they also are parallel to each other and resemble a ladder lying on the ground. In addition to these true ladder veins, the dikes contain flat-lying, diagonal, and veins that parallel the course of the dikes. If the dikes in which ladder veins are present are followed along strike, the veins are found to be quite unevenly arranged; parts of the dike are full of them and other parts are almost free of them. This pattern of vein location is directly dependent on the manner in which the dikes responded to tectonic stress; where the dikes broke readily, the fractures were abundant; where the dikes were competent, fractures did not occur and no space was provided for the ore fluids to fill.

Of all the types of granite-like rocks, veins are most rich in dikes of types (4) and (5) [see above] where they dip steeply and where they are in platy schistose rocks. The dip of these rich veins is 70° to 80°, usually to the south but locally to the north. The strikes of the veins range between E-W and N70°W. The veins generally are plate-like in shape, but they may be lensoid, in which case, they are arranged en echelon. The length of the ladder veins is governed by the thickness of the dike in which they are; they do not extend beyond the margins of the dikes.

The main gangue mineral of the ladder veins is drusy quartz with the open spaces among the quartz crystals being filled, in most instances, by sulfides. If the quartz crystals completely fill sections of the vein, the sulfides are in isolated lenses. Sulfides also may be contained in cracks in the quartz, these cracks appearing to have rather random orientations.

In contrast to the ladder veins, the diagonal veins are more regular in shape, are much longer, and extend to greater depths; veins of this type normally are made up of alternating thin bands of quartz and sulfides. The north-south veins are irregularly spaced within the dikes but are much the same in structure to the diagonal veins.

The ladder, diagonal, and north-south veins are bordered by aureoles of beresite. [Beresite, a term proposed by Rose in 1837 when he studied these deposits, is a rock resembling greisen and consisting of about 35 per cent quartz, 63 per cent mica, and 2 per cent pyrite; it is a typical example of the type of alteration to be expected to be produced by moderate- to high-temperature hydrothermal solutions acting of granite. Helmhacker (1892) considered the "fresh" rock at greater depths in the Beresovo mine to be a quartz porphyry. Johannsen (Petrography, v. 2, p. 23) says that beresite properly "should be applied only to dike rocks that have been altered to greisen-like material".] This beresite, therefore, is definitely an alteration product of the granitoid rocks of the Beresovo dikes produced by the same ore fluids that deposited the gold and its associated sulfides.

The other major type of veins in the Beresovo deposit are those known as krassyk veins, and they occur in the wall rocks of the gold-bearing dikes. They have the same strike and dip as those ladder veins in the dikes, and they (the krassyk veins) are continuations, in many instances, of the ladder veins in the granitoids of the surrounding rocks (volcano-sedimentary and intrusive). These krassyk veins usually form groups of 4 or 5 and, in places, as many as 9 or 10 that make up a combined parallel and en echelon pattern. These krassyk veins may continue for distances of 200 to 300 m and in some places up to 1 km and may merge and separate again.

The greater number of krassyk veins are confined in two zones. The northern one was developed in serpentinites and gabbros and, to a much lesser extent, in diabases and tuffs. This zone includes all of the best gold-bearing krassyk veins of the orefield. The zone strikes N50° to N60°W and is about 2 km wide. The southern (second) zone is contained in tuffaceous shales. tuffs, and diabases; this zone strikes NW to N80°W and is up to 1 km wide. In the normal case, a krassyk zone consists of a main vein with a family of thinner and shorter veins on either side. These veins are composed mainly of quartz with lenses of sulfides in which pyrite is the dominant mineral but in which gold can be a commercially important mineral. These sulfide lenses may be joined to each other with thin stringers of sulfides.

The krassyk veins are surrounded by alteration aureoles of listwänite, a schistose rock composed various proportions of quartz, dolomite, magnesite, talc, and limonite. The listwänite appears to be the equivalent of beresite, being produced by the same solutions that deposited gold and sulfides in the ladder veins and beresite in the bordering dike rock. When there solutions encountered the volcanic and sedimentary wall rocks of the krassyk veins, the reaction produced the listwänite (Borodaevskaya and Rozkhov, 1977).

The ore mineralization developed in the dike (ladder) and krassyk veins is essentially the same. Quartz is the main mineral and appears to have been introduced in several generations; less important gangue minerals are such as ankerite, dolomite, and calcite. The metallic minerals make up from 2 to 10 per cent of the total vein content and include pyrite, tetrahedrite, scheelite, aikenite [$PbCuBiS_6$], galena, and chalcopyrite. It has been suggested that the ore mineralization was introduced in four stages: (1) quartz, (2) quartz-pyrite, (3) polymetallic (mainly tetrahedrite, galena, and aikenite) with gold, and (4) carbonates.

Most of the gold was deposited with stage (3) but some, in very fine powder, is associated with the pyrite of stage (2). Where there are lenses of stage (3) sulfides in the veins, gold is present in the greatest amounts; the fineness of the gold ranges between 800 and 930; apparently the major alloy element with the gold is silver, but this is not certain.

The distribution of the gold depends on the distribution of the sulfides; the more sulfides, the more gold. The highest concentrations of gold have been found in the central parts of the veins and in the hanging walls of the dikes. Where the veins are most abundant, it normally follows that the sulfides are also most abundant, so, in such situations, gold is, of course, likely also to be most common. With minor modifications this statment also applies to the krassyk veins, although these veins do not have such definite outer limits as those provided by the hanging walls of the granitoid dikes.

In the wall rocks of both types of veins, some sulfides have been introduced into the beresites and listwänites, respectively. These alteration aureoles range in width from 0.1 to 0.4 m to 1.0 to 1.5 m in thickness; the beresite zones around the dike veins normally are gold-bearing and much of that material can be mined commercially.

The age of the wall rocks of the Beresovo veins strongly suggest that the age of the ore mineralization was late Paleozoic; I have seen no data that controdict this assumption.

The sulfides associated with the gold indicate that the gold was deposited in the mesothermal to leptothermal ranges, although the scheelite may suggest a somewhat higher temperature-pressure range. Probably the quantity of scheelite present is insufficient to justify adding hypothermal to the designation in the modified Lindgren classification.

PECHENGA (PETSAMO), MURMANSK OBLAST

Middle Precambrian	Nickel, Copper, Cobalt, Platinum Metals, Silver	Magmatic-2a, Metamorphic-C

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Notes

The ore deposits of the Pechenga (Petsamo) region are centrered around the town of that name, at 69°28'N,31°04'E, that is within 12 km of the Norwegian border in an area that was, before World War II, the extreme northeastern part of Finland. Pechenga is about 45 km northwest of the major port of Murmansk and some 25 km southwest of the nearest point on the Barents Sea.

The oldest rocks in the district are Archean in age and are the granite gneisses of the basement. The oldest of the supracrustal rocks lie unconformably on this basement and are designated as the Tundra series by the Soviets (Pt_1). Pt_1 is a Soviet shorthand for the lowermost Proterozoic, just as D_3fm refers to the upper Devonian Famennian stage or C_1v to the lower Carboniferous Visean. The eruptive and tuffaceous and sedimentary formations that make up this Tundra series consist of an early and major introduction over the entire Pechenga area of submarine spilitic lavas into an area that was steadily being geosynclinally downwarped. The axis of this synclinorium strikes northwest; in the Pechenga area, these beds are on the northwest flank of the synclinal structure and normally dip 30° to 45°SW but may be as steep in places as 80°SW. These spilitic lavas were slowly covered by a thick sequence of shallow marine sediments (sand and clay) derived from the adjacent land surface; this sequence included intercalations of volcanic and carbonate rocks, the entire sequence being denoted as the Tundra series.

This Pt_1 phase was ended by the early Karelian orogeny in which the sedimentary-volcanic beds of the Tundra series formed into steep isoclinal folds with a northwest strike; appreciably faulting also accompanied the orogeny. Closely combined with these tectonic developments was the development of synorogenic masses of granodiorite and plagioclase-microcline granites, probably by rheomorphism (partial anatexis) of the granite gneiss of the Archean basement.

The next rock group to be developed was the Pechenga series that had much in common with the Tundra. While this series was being formed, the geosynclinal sinking continued (although with the axis offset somewhat to the northeast). The rocks that were formed were masses of diabase flows and tuffaceous-sedimentary rocks that eventually reached a thickness of 10 to 12 km (Pt_2). Originally, the Finns named this series the Tunturi formation (Wegmann, 1929) or the Petsamotunturi formation (Sederholm, 1930), and these names can be considered the essential equivalents of the Pechenga. The rocks of this series cover an irregularly shaped area, the greatest length of which (northwest-southeast) is about 70 km and its greatest width (northeast-southwest) is some 35 km. The rocks of this series lie on the complexly metamorphosed and folded rock of the Archean gneiss and the Tundra series with which their slightly metamorphosed character contrasts strongly. Haapala (1969) says that polymict conglomerates mark the contact between the highly metamorphosed rocks beneath and the Pechenga beds above.

In contrast to Väyrynen (1938), who divides the Pechenga series into nine separate zones, Glazkovsky, and others (1977) divide it into four major eruptive sheets, separated by beds of tuffaceous and sedimentary rocks. From the bottom upward, these eruptive rocks are: (1) fine-grained metadiabases, metaporphyryites, and amygdaloidal rocks 1200 to 300 m thick; these are overlain by 100 to 200 m of quartzites, sandstones, and dolomites lenses; (2) metadiabases, metaporphyrites, albitophyres, keratophyres, and tuffs 1200 to 1600 m thick - the upper portion of this sequence contains (or is overlain by) lenses of conglomerates, quartzites, arkoses, phyllites, and dolomites as much as 50 m in thickness; (3) aphanitic diabases, amygdaloidal rocks, and tuffs - at higher elevations and along strike, is a well-established group of sandstones that average about 1000 m thick - this mainly diabasic rock series is the one that contains the mafic and ultramafic rocks in which are included the Pechenga ores; (4) along a thrust plane above rock series (3) was introduced a sheet of augite diabases in which there are seams of picritic diabases, tuffs, and tuffites that total over 4000 m in thickness; into this sequence were intruded sills of gabbro-diabases that are enclosed in turn in the tuffaceous-sedimentary sequence included in the augite diabases,. Of the total Proterozoic rock thickness, Glazkovsky and others (1977) report that about 80 per cent is mafic igneous rock.

Following the end of the accumulation of the Pechenga series, late Karelian orogeny produced new folds, accompanied by thrust and normal faults and the injection of the ultramafic and mafic magmas that contain the nickel-bearing ore bodies of which (including those now, at least, uneconomic to mine) there are at least 20. This orogeny also marked the end of the folding of the Pechenga geosynclinal zone. The blocks of Archean basement on both sides of the geosynclinal fold belt probably also were affected by these same orogenic movements and these produced, in the Archean rocks, structures into which ultramafic and mafic magmas of Proterozoic age were introduced, and in these nickel-bearing rocks some copper-nickel sulfide ores were produced, including those of the Allarechen region south-southwest of Pechenga and apparently quite close to the Finnish border.

Glazkovsky, and others (1977) report that the Pechenga district contains 110 nickel-bearing, ultramafic and mafic intrusions and 60 gabbro-diabase bodies in addition. These all occur, as stated above, in rock series (3). The igneous bodies are in layers or lenses that are from 2 to 700 m thick and from 200 to 7000 m long. All have been affected by the same folding that deformed the rocks that enclose them. Their 30° to 60° dip to the south and southwest conforms to that of their host rocks and, of course, brings them to greater and greater depths as they are followed in a general southerly direction.

In the central part of the ore field, where the host rocks have undergone the greatest downwarping, about three-quarters of the total volume of igneous is located, and this area contains all of the known occurrences of economically mineable nickel ore. The small intrusive bodies to either side of the ore-bearing belt include no mineable ore.

The igneous rocks are primarily related to the folds in the sedimentary rocks and are thickest in the axial volumes of these folds (Glazkovsky, and others, 1977).

The surface outcrop of the folded, ore-bearing igneous rocks and their enclosing volcanic and sedimentary beds is arc-shaped, convex to the northeast and about 27 km in length. The actual structure extends beyond the ore-bearing portion for several kilometers at both ends. The designations of the ore bodies that were assigned by the Finns are as follows (from west to east): (1) Kaulatunturi; (2) Kotseljoki; (3) Kammikivi; (4) Ortoaivi; (5) Soukerjoki; (6) Mirona; (7) Pilgujärvi; (8) Onkitunturi. The Soviet ore geologists recognize: (1) Kaula; (2) Promezhutochnoe (Intermediate); (3) Flangove; (4) Kammikivi; and (5) Western Ortoaivi. These deposits (of the western ore node) are located on the western limb of the Central Downwarp and are contained in a zone of tectonic crumpling confined to one interlayer in the upper part of the tuffaceous-sedimentary sequence. On the eastern limb of the downwarp are the deposits of the eastern ore node: (1) Kierdzhipor and (2) North Onki, and these also are contained in a zone of crumpling that is confined to a single inter-layer; this interlayer is immediately adjacent to the underlying diabases of the third sheet. Between these two nodes, in the central part of the downwarp, are igneous rocks and ore deposits that are divided between two sedimentary belts. One of these is located in the basal portion of the tuffaceous-sedimentary sequence and contains the Eastern Ortoaivi, the Northern Soukeriioki (Russian spelling as contrasted with Finnish above), and the Mirona deposits. The other belt is contained in the upper portion of the tuffaceous-sedimentary sequence and includes the Raisoaivi deposit and the minor mineralization of the Soukeriioki Creek area. Obviously, an exact correlation cannot be made between the Finnish and Soviet deposit names, but there is enough correspondence among them to make the identification of their relative positions quite simple.

The igneous rocks of the ore-bearing zone are contained between overlying flows, pillow lavas, and minor pyroclastics and underlying agglomerates, tuffs, and massive diabase bodies. The boundaries of this ore zone with the non-ore-bearing zones above and below normally are well delineated, but local intrusions of the rocks of the ore zone into the neighboring rocks make the zone boundaries in such areas uncertain (Haapala, 1969).

The actual host rocks of the layered, ultramafic to mafic ore-bearing igneous rocks are phyllites and concordant intrusions of diabase; locally, mafic dikes are present. In this zone, the phyllites are the most common rocks, but some coarser-grained sediments, up to and including conglomerates, occur. In some places (mainly in the western area), the layered rocks are persistently in one horizon and may be several kilometers long; their width is from a few 10's of meters to several hundred. The layering is broadly conformable to the attitude of the intrusions; in the usual instance, the sequence in the ultramafic-mafic layer, from foot-to hanging wall, is; (1) a fine-grained contact phase; (2) serpentinite, produced by the alteration of olivine-rich rocks; (3) pyroxenite; and (4) gabbro - more silicic phases are known, but they are not common. The serpentinite normally is the most abundant rock type, and gabbro the next, but local differences in the amounts of the rocks in the ore zone may be appreciable. The various rock types lens out horizontally, and masses of both serpentinite and gabbro are known that have no connection (at least that is exposed) to the layered units. The contact between serpentinite and pyroxenite ranges from sharp to nearly so, but the change from the contact phase into serpentinite is gradual over short distances, but the gradation from pyroxenite to gabbro is almost imperceptably gradual. If a body of ultramafic rock has been converted entirely to serpentinite, the hanging-wall contact is fine-grained.

Although the hanging-wall block has been little affected by the intrusion of the layered rocks, the ore zone shows strongly the results of tangential movements. The intrusions are considerably, though irregularly folded, and the numerous shear zones are unevenly distributed across and along the layers, but these are controlled by the intrusive contacts (Haapala, 1969).

As the major synclinal folds are centrally located, and these folds control in positions of the mafic and ultramafic rocks, and the ores are concentrated in the axial portions of the folds. On the basis of internal structure, Glazkovsky and his colleagues (1977) have separated the nickel-bearing intrusions into two varieties: (1) differentiated intrusives; from bottom to top, these are composed of altered peridotites (wehrlites, now serpentinites); pyroxenites, gabbros; and monzonitic gabbros; (2) undifferentiated, usually quite thin and made up completely of serpentinites or gabbros. These two varieties of intrusions appear, however, to be of the same age; the differences between them being due, apparently, to the thin ones having crystallized quickly and the thick ones much more slowly. In both types, the serpentinization was due to autometamorphism.

Bodies of igneous rock may bear the same name, gabbro for example, but may be quite different in composition, testifying, as Glazkovsky and his colleagues point out (1977) to the heterogeneity of the parent magmas of the nickel-bearing intrusives. They are convinced that some of the magmas separated into peridotite and gabbro fractions, respectively, even before they were injected into their final resting places.

The average composition of the parent magma of the Pechenga ultramafic-mafic rocks must have been about that of a melanocratic olivine gabbro and probably was produced by partial melting of a deep-seated peridotite. The only characteristics that suggest it may have been somewhat contaminated by assimilation of, or mixing with, basalt are a relatively large content of titanium and a very small amount of contained chromium.

Haapala (1969) points out that the ore deposits are consistently located in synclines that range from gently downward warps to sharp troughs or even in the minor curvature at the end of an ultramafic lens. In the usual instance, the ore deposits are an abnormal addition to the succession of layers in an ultramafic body. The ore deposits in most cases are located low in the layering of any given ultramafic and are, therefore, closely associated with serpentinites. The ores may consist of footwall serpentinite that contains far more than the usual sulfide content associated with non-ore serpentinite. The enrichment is locally so great as to constitute a distinct layer-like body that includes enough disseminated sulfide to make it high-grade ore. In some places, such high-grade disseminations stop at the upper edge of the fine-grained serpentinite of the contact with the foot-wall [originally a fine-grained layer of pyroxene and amphibole produced during the relatively rapid cooling of the margin (the lower one in this case) of the ultramafic intrusion]. In other places, there may be a massive separate layer of Ni-Cu sulfide between the intrusive and the phyllite of the footwall.

Haapala (1969) indicates out that nickel-bearing sulfides occupy a large portion of the layered intrusions and are closely connected with the serpentinites. Most of the disseminated sulfides are in the basal section of these altered ultramafics, and these enriched areas also may include a layer-appearing concentration of high-grade sulfides. In some places, this layer reaches no farther downward than somewhere within the fine-grained lower margin of the serpentinite. In others, however, this mineralization not only extends to the contact between serpentinite and the footwall phyllite but actually spreads out along this contact as a layer of massive sulfides.

For each deposit, the dividing line between ore and waste is economic, but these contacts (both between high-grade disseminations and the low-grade variety and between massive layers and the mineralized phyllite) normally are easily recognized and conform in general to the intrusive structures. In general, the core of the mineralized body is high-grade ore, and this core dies out in all directions to give this ore a lens-like shape; the lateral dimensions of these high-grade ore bodies are much more extensive than the vertical ones.

The principal ore type in the Pechenga district, however, is disseminated ore in serpentinite; all other ore types are present only in close proximity to the disseminated ore.

Another ore type, of relatively minor economic importance, is composed of mineralized shears or breccias along the footwall; this type, reasonably, occurs most commonly where the host rocks have been most broken. In contrast, locally, within disseminated ore bodies are volumes of serpentinite that contain only minor quantities of sulfides and have sharp contacts with the enclosing and mineralized serpentinite.

Glazkovsky and his colleagues (1977) agree that the ore bodies are confined to the ultramafic and mafic igneous masses, principally being in the serpentinitized peridotites in the lower portions of such rock assemblages. Such few ore bodies as are entirely in the phyllites are readily followed to their parent intrusives; ore bodies of this type are breccia- or vein-fillings. Such breccia- or vein-filling ores are not only in the phyllites, but similar ores can be found in the serpentinites as layered injections or as small veinlets and injections. Glazkovsky and his colleagues consider these ore types to be epigenetic in the serpentinites. These authors also emphasize that the deposits of the Western ore node differ considerably from those of the Eastern; those of the Western node have been appreciably eroded and, therefore, exhibit the lowest portions of the ore bodies, whereas, those of the Eastern node have barely been affected by that process. A result of this lack of erosion is that the Eastern node ore bodies have such great vertical length (down dip) that they had not yet (1977) been completely explored.

Haapala recognizes four types of ore: (1) disseminated ore; (2) sulfide layers; (3) mineralized phyllite; and (4) breccia ore. Glazkovsky and his colleagues have designated the ores as: (1) uniform (= disseminated); (2) breccia-like (= breccia); (3) segregated in serpentinites (= sulfide layers); and (4) veinlet-segregated in schists, that is phyllites (= mineralized phyllites). The latter group of authors add information as to the nickel, copper, and cobalt ratios in the four ore types: (1) 48:19:1; (2) 56:22:1; (3) 55:24:1; and (4) 47:48:1. Haapala does not give ratios but says that "averages of nickel:copper ratios for the disseminated mass and main breccia have almost the same value, with nickel running about twice that of copper. In the mineralized phyllite, copper generally equals or supercedes the amount of nickel. This says almost what was given in figures by the Soviet group.

The major constitutents of the ore are pyrrhotite, pentlandite, and chalcopyrite, although Eliseev, and others,(1961) report that bornite is present instead of pyrrhotite in certain disseminated ore; some bornite is known in the Kaula breccia as small grains and as lining in shears. Minor sphalerite occurs in all ore types. Such mineralization is found in all major sulfide deposits throughout the world with the exception of the locally abundant bornite at Pechenga.

In the average ore occurrence, the boundary between the ultramafic rocks and the phyllite is designated as the fine-grained contact phase. This phase is no more than a few meters thick and, in many places less than that. It appears to be composed, now, of pyroxene and amphibole, but originally, it probably represented a chilled contact and probably represented the average composition of the intrusive mass of which it was the first portion to be solidified. Above the contact phase is serpentinized peridotite. The serpentine minerals in almost all instances are pseudorphous after olivine but with a chlorite and biotite matrix. Only a few remnants of olivine are known, and a few phenocrysts of pyroxene still are present. In some places the pyroxenes are hydrated and talc may be quite common in both the matrix and the phenocrysts. Magnetite is quite common, but ilmenite is confined to the contact phase. The large number of minor minerals includes: violarite, bornite, cubanite, mackinawite, and vallerite. Rare minerals are; chloanthite, niccolite, cobaltite, millerite, galena, and the platinoids. In the autometamorphism, serpentine and chlorite were produced from the olivine and, locally, from the pyroxene, and amphiboles and chlorite from the pyroxene. Chrysotile asbestos is present in joints and cavities. During this same process, magnetite was produced during the autometamorphism of olivine, carbonates from olivine and pyroxene, and biotite probably directly from the deuteric solutions. The Soviets consider that hydrothermal metamorphism produced talc, carbonates and pyrite.

All this is typical of autometamorphosed peridotites. The bulk of the sulfides are positioned around the pseudomorphs. If the ore is low grade, then the sulfides are in irregularly scattered patches and specks; some sulfides are rounded and cut across the internal structures of the matrix minerals or are present in them as inclusions (Haapala, 1969).

Where the ore is of high grade, the specks coalesce into networks; these networks usually are fine-grained enough to outline the forms of the former olivines. The net may be entirely of sulfides, but various proportions of magnetite and silicate may be included; the silicate is mainly antigorite. The sulfides consist of large individual pyrrhotite crystals that are irregularly surrounded by zones of pentlandite, chalcopyrite, and magnetite.

The silicate material (dominantly antigorite) associated with the sulfides stands out from them due to the differences in texture and their being separated from each other by magnetite coatings. Islands of serpentine may be enclosed in a mixture of antigorite fibers and sulfides. If the antigorite was formed after the sulfides were emplaced, as appears probable, the conversion of olivine to antigorite does not seem to have had any appreciable effect on the composition of the sulfides interstitially present (originally) in relation to olivine, but the sulfides may have been rearranged in space. The relative abundance of serpentine (antigorite) to sulfides almost certainly shows the same relationship of olivine to sulfide.

The magnetite, apparently produced during the autometamorphism, exhibits closely packed groups or chains that may surround the sulfides or even penetrate into cracks in them. Thus, the magnetite, as is the antigorite, probably is later than the sulfides. The possibility that the sulfides have been somewhat rearranged is shown by the presence of sulfides in veinlets. How much of this rearrangement is due to autometamorphism and how much is due to the tectonically generated metamorphism that followed the intrusion of the ultramafic rocks and their included sulfides is uncertain, but both probably were important.

A considerable argument for the effect of tectonic metamorphism on the primary sulfides is shown by the breccia ores in which massive primary sulfides were broken by the same forces that disturbed both the underlying phyllite and the overlying serpentinite as fragments of both of these rock minerals are found as fragments in the sulfide matrix of the breccias. Of course, in some breccias the sulfide material has been forced into bedding or fracture planes in the phyllite to produce what Glazkovsky and his colleagues call mineralized plicated phyllite. The presence of minor carbonate and quartz in the breccia matrices probably is the result of late-stage hydrothermal solutions generated during autometamorphism.

The presence of chalcopyrite in the various ore bodies as veinlets that both cut and isolate pyrrhotite and pentlandite indicates that the well-documented greater mobility of the $FeCuS_2$ under metamorphic pressures also obtains at Pechenga.

Glazkovsky and his colleagues point out that chlorite, present as intergrowths with sulfides, is in greater abundance the greater the degree of alteration of the rocks. This again appears to be a result of autometamorphism, the greater the severity of which, the greater the development of chlorite and of its intergrowth with sulfides. How much of the production of chlorite is a result of internal generation of autometamorphic fluids and how much to hydrothermal fluids exogenous to the ultramafic magma-rock complex is uncertain. I suspect, however, that the fluids produced during autometamorphism were largely or entirely responsible for this and also for the remainder of the later alterations of the ultramafics with the tectonic metamorphism to be charged only with those effects, such as chalcopyrite veinlets, that are due to pressure alone.

From these facts and the reasoning from them, it would appear that the entire process of the development of the pre-tectonic Pechenga deposits. resulted from reactions internal to the ultramafic magma. What happened during the period of tectonic activity is another matter.

The more complex the tectonic effects on the host rocks and on the ultramafic rocks that contain the ores, the more complex the structure of the ore deposits themselves. This is particularly true of the deposits of the Western ore node, the deposits of which were emplaced at appreciably greater depths than those of the Eastern as already has been discussed. Such complexity is clearly shown in the Kaula deposit (as can be seen from Figures 3, 4, and 5 of Glazkovsky and his colleagues and Figure 3 of Haapala). Especially, it should be noted in Haapala's Figure 3, that a sulfide layer is shown on the top of the Kaula ultramafic body. As it seems almost certain that this situation only can have resulted from the folding back of the intrusive on itself during the tectonic phase, it follows that the sulfide layer was present before the folding took place. I can envision no mechanism that possibly could produce the sulfide (massive) layer on the top of the ultramafic intrusion during the pre-tectonic crystallization of the intrusive, so it can be where it is only if it was folded back on itself more or less around the initially central horizontal axis of the intrusive lens.

The content of platinum-group metals (PGM) in the Pechenga ores is low, much lower than other deposits of this type. Such PGM metals as are present are concentrated in the chalcopyrite-pyrrhotite-pentlandite ores of the layers or lenses of the peridotite-gabbro massives that belong to the intrusive phase of the Pechenga basaltoid association (Godlevsky, 1968). To a lesser extent, the PGM metals also occur in the mineralized plicated phyllite beneath the ore bodies. The disseminated ores, although poor in the metals of the PGM, are relatively enriched in platinum and palladium. The layer-like segregations of the most massive ore along the tectonic contact between the intrusions and the (normally) underlying country rocks are the richest locations for PGM. The Pd:Pt ratio of these deposits averages 1:2; the distribution of the PGM, however, is most uneven. Palladium is particularly concentrated in pentlandite and platinum and rhodium in pyrrhotite. Unusually, so far as ores of this general type is concerned, the chalcopyrite of the Pechenga ores contains them in even lesser amounts than is true of pyrrhotite. The only two Pt-minerals in the ores are sperrylite and native Pt. The largest (though still small) amounts of PGM in the Pechenga ores are found in the massive and breccia ore of the Kammikivi and Kotsel'vaar deposits (Cabri, 1981).

It seems apparent, from what has been said here, that the sulfide material in the ultramafic-mafic magmas injected into the volcanic-sedimentary beds in the Pechenga suite contained so much sulfide that, somewhere in the crystallization cycle, more sulfide was present than the dominant silicate melt could hold in solution. This sulfide material most probably separated from the ultramafic-mafic magma in the molten state and did not begin to crystallize until the silicates largely had crystallized themselves. This meant that some of the molten sulfide mass was able to find its way to the base of the silicate mass where it crystallized in layers largely composed of sulfides. By far the bulk of the sulfides, however, crystallized in the interstices among the olivine grains, these being the more strongly connected physically, the greater the concentration of sulfide to silicate in the volume in question. Thus, it can reasonably be said that the separation of the sulfide melt was relatively early in the crystallization cycle of the ultramafic-mafic magma but that its solidification took place only after the bulk of the silicate melt had solidified. Except for the concentration of appreciable, but minor, portions of the sulfide material available, in largely massive layers in the lower reaches of the ultramafic body, most of the molten sulfide crystallized not far from where it had separated from the silicate melt in which it originally was dissolved. Thus, the Pechenga sulfides, in the first stage of their history as solid materials, should be classified (under the modified Lindgren classification) as Early Separation, Late Solidification, that is Magmatic-2a.

This was not, however, the end of the story. In the severe tectonic activity of late Karelian time, the volcanic-sedimentary rocks that were hosts to the ultramafic-mafic magma and the rocks that crystallized from these magmas were strongly folded. This folding was so strong, at least locally, that the massive sulfides, basal to the formation in the crystallization cycle, were forced into brecciated masses of phyllite and serpentinite to produce the breccia ores that compose a minor, but high-grade, part of the Pechenga ores.

It should be obvious from the preceding discussion that autometamorphism (deuteric alteration) produced important effects on the ultramafic rocks. Except, however, that the conversion of olivine to serpentine probably caused some slight change in the position of sulfides within the altered ultramafic bodies, the autometamorphism had little effect on the contents of nickel and copper in the ore, and it would serve no useful purpose to include this category in the classification of the Pechenga deposits.

The tectonic metamorphism, produced by the late Karelian orogeny, however, did much to change the position of the ore minerals, even though it probably did not add to the total nickel-copper content of the area. The highest-grade ores are those of the massive sulfide matrices in the basal portions of the deposit and certainly deserve mention in the classification as Metamorphic-C.

Much has been said about the hydrothermal effects on the Pechenga ores, but it seems to me that most of these were accomplished by deuteric solutions and not by hydrothermal fluids from an exogenous source. No term connoting hydrothermal activity in the strict sense, therefore, is included in the Pechenga classification.

SADON, NORTH OSETIAN ASSR

Middle Mesozoic (or younger)	Lead, Zinc	Hypothermal-1 to Mesothermal

Arkhangel'skaya, V. V. and Vol'fson, F. I., 1977, Geotectonic positions and systematic stratiform lead-zinc deposits: Nauk Press, Moscow, 274 p.

Azhgirei, G. D. and Yefremov, G. M., 1950, The geological-structural features of the region of polymetallic deposits of northern Osetia: The Natural Resources of Northern Osetia ASSR, Nauka Press (A.N. USSR), Moscow, p. 126-152 (or A. N. Kazakh, Alma Ata?)

Azhgirei, G. D., and others, 1968, Geological-petrographical features of the middle Devonian porphyry intrusions of the southeastern Rudnyi Altai: Inst. Higher Learning, Geol. and Exploration, Reports, no. 7, p. 27-28 (Izv. vyssh. ucheb. Zaved., Geol. i Razv.)

Bashkina, E. F., 1972, On the conditions of localization of the lead-zinc mineralization of the Sadon Deposit: Problems in the Formation of Ore Pipes, Materials from a Symposium, Novosibirsk, p. 275-281

Knyazev, I. I., 1960, Distribution patterns of zinc-lead mineralization in the Karatau: The Fundamental Ideas of N. G. Kassin on the Geology of Kazakhstan, Akad. Nauk Kazakhstan SSR Press, Alma-Ata, p. 379-400

Kruglov, V. A., 1979, The influence of the structural and lithological factors in the localization of hidden ore bodies of the Sadon deposit: Akad. Nauk USSR Reports, Geol. ser. no. 8, p. 117-129

Prokopenko, N. M., 1950, Phases of mineralization in the lead-zinc deposits of the Sadon region: The Natural Resources of the North Osetian ASSR, Akad. Nauk. Kazakstan Press, Alma-Ata, p. 163-180

Rudnev, V. N., 1941, The Sadon polymetallic deposit: Sevkavtsvetmetrazvedka, Nal'chik (northern cavernous non-ferrous metal survey?) (Nal'chik, a city on the north slope of the Greater Caucasus)

Rychkov, A. I., 1977, On the forms of endogenic reaction rims of the Sadon deposit: Geology and Prospecting for Deposits of Commercial Materials, Moscow, p. 97-101

Rypinsky, E. G., and others, 1969, Conditions of localization of polymetallic mineralization in the Sadon and Zgid deposits: The Geology and Mineral Resources of Northern Osetia, p. 47-51 (see Azhgieri, G. D., above?)

Smirnov, V. I. and Gorzhevsky, D. I., 1977, The Sadon deposit, in Smirnov, V. I., Editor, 1977, Ore Deposits of the USSR: Vol. II, Pitman Pub. Co., London, p. 198-201

Trofimov, N. N. and Pobokova, O. P., 1977, Element indicators of lead-zinc mineralization and of major discontinuous dislocations of the Sadon group of deposits: Geology and Prospecting for Deposits of Commercial Minerals, Moscow. p. 102-131

Notes

The Sadon lead-zinc deposit is located in the North Osetian ASSR, a small political subdivision that descends from the crest of the Greater Caucasus mountains, from an elevation of over 4600 m to reach the south side of the Terak River that eventually enters into the Caspian Sea. Sadon is nearly 60 km SW of Ordzhonikidze, the capital of this ASSR, and probably is at an elevation of perhaps 3500 m above sea level.

The deposit has been explored and mined since 1853; at the present time, the deposit is essentially exhausted.

The Sadon ores are contained in a complex vein that has filled a major fault; this fault cuts not only Paleozoic granites but also Jurassic flows that overlie them. This structure dies out in the Jurassic rocks that form the western and eastern limbs of the Sadon-Zgid anticline. Most of the vein is contained in the anticlinal core of Paleozoic granite; locally this granite grades into granodiorite that form a facies of the granite massif; these silicic rocks are basically biotite-microcline-plagioclase granites (plus the granodiorites just mentioned). On the north limb, the granites pass under north-dipping flows of andesite-dacite porphyrite and related tuffs; a basal conglomerate lies between these rocks and granites beneath them. All these rocks are of Toarcian age, that is lower Jurassic just below the Bajocian. On the southern limb, middle Toaracian shales have transgressed over the Paleozoic granites; a major thrust fault, south of the contact, gives the granite immediately under the shales a wedge shape. At some stage in the structural histroy, NW-striking joints cut the granite; these have been filled by dikes of andesite-dacite porphyries, probably from the same magma chamber that supplied the temporally related porphyrite flows of the same rock type on the north limb.

The strike of the complex vein in which the Sadon deposit is contained in N30°E to N50°E; this fault has been followed on the surface for about 4 km and down dip for as much as 1.5 km and for another 300 m by drilling. The fault (and the two major branches into which it is divided as it is followed to the northeast) is of the strike-slip variety with the NW-side being displaced to the southwest by nearly 1500 m; the dip is nearly vertical. The northwest side also is downthrown by about 300 m. The southeast boundary of the fault is appreciably more clearly marked than is the northwestern, is designated as the Eastern strike-slip fault, and has been filled with gouge produced by the movements of the two sides of this fault relative to each other. In contrast, the northwestern bounding fault is far less well shown, it being possible to trace it through a series of various ore veins and joints that are collectively known as the Western reverse fault.

Although ore is present in the two main faults, most of the ore of the deposit is located (Kruglov, 1979) in subordinate fissures that strike out from the Western fault in the higher levels of the fault zone and from the Eastern in the lower levels at somewhat flatter angles than the main faults. These minor fissures completely cut up the rock between the two main faults in a most intricate pattern. These subordinate fractures take off from curved segments of the main faults. The thickness of the minor fissures range from a few centimeters up to 20 m; the average thickness is between 1.0 and 1.5 m.

In addition to these lead-zinc-bearing veins, single veins, distinct from the main and containing mainly pyrrhotite, are present; these veins well may be different in age and in age of mineralization from the lead-zinc ones.

According to Kruglov (1979), the faulting took place over a major span of geologic time, the area being composed of several types of faults, the strikes of which appear to range between north-south and east-west. Fault development occurred in five stages: (1) ancient (pre-intrusive time); (2) pre-volcanic extrusion time; (3) pre-ore time; (4) intra-ore time; and (5) post-ore time. Stage (1) was marked by the development of the main faults; these were intruded by dikes of aplites and pegmatites. Stage (2) was the time of reactivation of the older faults and the intrusion of pipes and dikes of various silicic rocks. Stage (3) saw the formation of the ore-bearing faults and fissures in the granitoid rocks; these were later filled by ore bodies in several stages. Stage (4) saw the filling of the veins and veinlets with ore and movement on the faults coincident with the arrival of the ore. Stage (5) resulted in weak movements in the earlier of the ore fissures. The directions of displacement during the various stages of fracturing were the same during the entire geological history of the district. The ores veins contain appreciable amounts of agglomerate; in the upper horizones, these were localized by the fissures; at middle depths by shale zones; and in the lower horizons by the contacts between dikes and eruptive rocks.

In the Sadon deposit, the main ore minerals are galena, sphalerite, and quartz with their introduction being in four phases. The phases [all during Kruglov's stage (4)], were separated by the crushing of the mineral material that already had been deposited. In phase (1) a weak mineralization of quartz and pyrite was precipitated; in phase (2) a moderate mineralization of quartz, calcite, and pyrrhotite, plus lesser amounts of pyrite, arsenopyrite, chalcopyrite, sphalerite, and galena; in phase (3), the main ore phase, much sphalerite, galena, quartz, and calcite, plus some pyrite, pyrrhotite, arsenopyrite, chalcopyrite, tetrahedrite, native bismuth, and manganosiderite; in phase (4) the main mineral was calcite with some aresenopyrite and sphalerite (cleiophane) (Smirnov and Gorzhevsky, 1977). Kruglov (1979) gives a slightly different arrangement of minerals in the four phases, his being less detailed but the principal minerals are essentially the same.

Wall-rock alteration during ore formation converted the granites around and between the veins to quartz, sericite, and chlorite-containing rocks.

Rypinsky and others (1969) point out that: "the most intense mineralization has been observed in the sector of the eastern fault" where the strike is between N38°E and N40°E and the dip is vertical or steeply NW. This sector is estimated to have contained 50 per cent of the Sadon ore tonnage. Two other ores zones (Khoda and New) are located where the Eastern fault has a strike of N46°E and contain about 30 per cent of the ore reserves. Another 10 per cent of the reserves are located where the Eastern fault has a strike of N26°E to N30°E (Rypinsky, 1969). Thus, it is readily apparent that the mineralization in the Sadon faulted zone is far from evenly distributed.

As is to be expected in a deposit of this type, in which the temperature of ore deposition was lower nearer the surface and higher lower down, galena is most abundant, relative to zinc, in the higher levels, whereas sphalerite and pyrrhotite increase in importance relative to galena with increasing depth. A definite change of lead content by a minus 0.53 per cent occurs with each 100 m of added depth in the mine workings, but it also apparently is true that the amount of sphalerite, in actual terms, also goes down as greater depths are reached. This decline in zinc content is -0.40 per cent over each 100 m of greater depth. Since the rate in decline of zinc is less than that of lead, the ratio of lead to zinc goes down from 0.54 in the upper levels of the deposit to 0.30 in the lower horizons. If there had been an actual increase in the amount of zinc with depth, instead of one relative to lead only, the lead-zinc ratio at depth would have been perhaps, as low as 1.0:1.0 or even less in comparison with a ratio of possibly 2.0:1.0 in the upper levels.

Trofimov (1977) has suggested that ore bodies not outcropping at the surface, that is "hidden" from immediate view, may be discovered by studying the contents of such elements as mercury, boron, iodine, and silver in the near-surface rocks or soils. The forms assumed by these geochemical anomalies at the surface are due to tectonic dislocations of the rocks beneath and, therefore, to the forms of ore-controlling structures. Further, there appears to be a definite relationship between the depth of a particular ore body (hidden or not) and the character of the aureoles as seen at the present surface.

The mineralization of the Sadon deposit is typical of many hydrothermal deposits in which the mineralization in the upper levels appears to have been deposited under mesothermal conditions and that at lower levels under hypothermal, in this case in non-calcareous rocks, that is hypothermal-1. The considerable quantities of arsenopyrite and pyrrhotite in the lower levels and their essential absence in the higher seems to confirm this classification.

The age of the ores cannot be older than middle Mesozoic (because the extreme northern end of the ore body is in lower Liassic rocks). The development of the last stages of faulting and of some of the mineralization may have taken place in the late Mesozoic or even, perhaps, on into the Tertiary. It appears best here, however, to classify the deposits as Middle Mesozoic.

The Zgid lead-zinc deposit is slightly north and west of that of Sadon. Although the details differ appreciably, the ore here is located dominantly in granite of similar age and character to that of Sadon, the major structure also is the Sadon-Zgid anticline. The main Zgid vein penetrates only a short distance into the Liassic shales and flows that are intimately folded into Paleozoic granites. The main ore-containing fault strikes NE-SW and is cut and displaced by NNE-SSW faults. The main vein has been traced along strike, with some breaks, for 3.5 km and along dip for 1200 m. The dip is steep and ranges between NW and SE. The horizontal displacement along the ore-bearing fault is perhaps 100 m and the vertical displacement about 70 m. The main fault has been cut and displaced by N-S, NW, and E-W pre-ore faults along which post-ore displacements of small-scale have occurred. The mineralization is apparently of a lower temperature than that at Sadon and probably should be classed as mesothermal without any reference to hypothermal.

KACHKANAR, SVERDLOVSK OBLAST

Late Paleozoic	Iron, Titanium Vanadium	Magmatic-3b

Borisenko, L. F., 1977, The Gusevogorsk deposit [vanadium content], in Smirnov, V. I., Editor, Ore Deposits of the USSR: Vol. I, Pitman Pub. Co., London, p. 260-263

Fominikh, V. G., 1977, Ore(-bearing) anorthosite of the Gusevogor deposit: Mineralogy of Geochemistry of Ultra-Basites of the Urals. Mineralogy no. 13, Pub. of the Inst. of Geol. and Geochem of the Ural Scientific Center, USSR, issue 125, p. 62-65

Fominikh, V. G., and others, 1967, The pyroxenites of Kachkanar: Ural' Branch, Akad. Nauk Press, Sverdlovsk

Fominikh, V. G., and others, 1976, Mosaic structures of titanomagnetites of the Gusevogor deposits and their influence on magnetic properties: Accessory and Ore Minerals of the Urals, Pub. of the Inst. of Geol. and Geochem. of the Ural Scientific Center, USSR, issue 118, p. 21-29

Razin, L. V., 1977, Manifestation of platinum mineralization of the Kachkanar type, in Smironov, V. I., Editor, Ore Deposits of the USSR: Vol. III, Pitman Pub. Co., London, p. 119-120

Razin, L. V. and Yurkina, K. V., 1971, Minerals of the platinum metals in the ores of the Gusevogork titanomagnetite deposit (middle Urals): Geology of Ore Deposits, v. 13, no. 2, p. 102-109

Sokolov, V. A. and Grigor'ev, V. M., 1977, Kachkanar deposit, in Smirnov, V. I., Editor, Ore Deposits of the USSR: Vol. I, Pitman Pub. Co., London, p. 35-37

Notes

The Kachkanar deposits are located in the Sverdlovsk Oblast at 58°45'N, 59°23'E in the Isovsk region of that oblast, about 30 km NW of the Nishnyaya Tura railway station, in about the center of the platinum-bearing belt of the Urals. The Kachkanar massif of gabbros and pyroxenites consists of the Kachkanar massif proper and the Gusevogorsk massif which contains eight areas of ore segregation; these are: (1) Northern, (2) Intermediate I, (3) Intermediate II, (4) Intermediate III, (5) Eastern, (6) Main, (7) Southern, and (8) Western.

The Kachkanar massif, including all eight divisions of the Gusevogorsk, is an ore-bearing gabbro-pyroxenite pluton and takes up an area of some 110 km^2. It is about equidimensional and appears to be laccolithic in shape. This pluton is enclosed in a variety of igneous and sedimentary rocks. Along the eastern boundary are plagioclase porhyrites and extrusive diabases that are Silurian in age. On the western contacts are Ordovician mica and siliceous schists. The borders of the northern and southern parts of the pluton have been converted to amphibolites. One-half of the total area of the pluton is composed of the two pyroxenite massifs with Kachkanar lying to the west of the fragments that make up the Gusevogorsk massif in its surface expression.

Actually, the Gusevogorsk massif is composed not only of pyroxenites but also of hornblendites and gabbros and has a north-south extension of 8.5 km and a maximum east-west width of about 3.5 km (Borisenko, 1977).

The Kachkanar massif has its maximum dimension of 5.5 km in a NE-SW direction, and its average width of 3.2 km in an east-west direction.

Of the eight areas of ore segregation in the Gusevogorsk massif, apparently only (6), the Main vein, has been exploited, within this "vein" the area of commercial ore is 1.1 km^2. In this "ore" area, however, there are subareas some of which contain low-grade and some are too low to be designated as ore at all. These subareas are between 1000 m^2 and 2200 m^2 and are nearly equidimensional. Mineralization had been proved (in 1977) down to 500 to 600 m below the surface, and the drill holes bottomed in ore.

It is reported that the contacts between the massifs (both Kachkanar and Gusevogorsk) and the country rocks are tectonic. In such zones of tectonic disturbance, the proportion of amphiboles was at its highest. The massif has a concentrically zoned structure in which the stratification is no more than vaguely indicated; this structure suggests that the massif is a brachysyncline. The axis of this syncline strikes NW-SE; it plunges toward the center of the massif at 30° to 35° in the northwest and from 70° to 80° in the southeast.

The Gusevogorsk pyroxenite massif, with which the vanadium-bearing titanomagnetite ores are spatially and genetically connected, is elongated in a north-south direction, its length being about 8.5 km and its width as much as 4.6 km; its area is some 22 km^2. The entire massif dips at about 75° to 80° to the east.

The pyroxenites of the Gusevogorsk massif are composed of diallage, olivine, hornblende, and plagioclase varieties. In the central portion of the massif, the main ultramafic rock type present is diallage pyroxenite. The predominate variety in the northeastern and southwestern parts is olivine pyroxenite. The hornblende and plagioclase pyroxenites normally lie in the transition zone between pyroxenites and gabbros. Mainly in the olivine pyroxenites, lenses of wehrlite occur; some lenses of olivinite also are present. These ultramafic rocks are massive rocks in which the grain size may be medium, coarse, and gigantic. Locally indistinct banding can be observed in these pyroxenites, and the wehrlites and the olivinites in places are highly serpentinized.

The Gusevogorsk pyroxenites contain numerous igneous veins (or narrow dikes) that are composed generally of anorthosite but some of them are gabbros or fine-grained pyroxenites (locally known as gusevites).

The titanomagnetite mineralization is contained normally in diallage and hornblende pyroxenites; in a few instances, it occurs with other types of ultramafic rocks. The veins mentioned in the preceding paragraph seldom contain any ore.

The segregations of ore have been discovered, mainly by drilling, in the eight areas mentioned in the first paragraph of this discussion, plus one more known as Vyisk. The shapes of these segregations are quite complex, and the transition from ore-bearing pyroxenites to non-ore-bearing olivine pyroxenites is gradual. Almost all the titanomagnetite in the ore-bearing rocks fills the interstices among ferromagnesian silicates.

The areas of both ore-bearing pyroxenite and below-ore-grade pyroxenites have been invaded by many dikes of hornblende and quartz-plagioclase rocks that are as much as 2 m in width; their strikes are quite varied and their dips as well (between 20° and 90°).

The ore bodies were produced by the segregation of masses of mainly titano-magnetite and less commonly by schlieren and veinlets of massive ore. Essentially all of these bodies are in pyroxenites , gabbros, and hornblendites and only in minor number in the olivinites and peridotites. Sokolov and Grigor'ev (1977) report that the ores have been divided into five types: (1) coarse (grains more than 3 mm in diameter); (2) medium (1-3 mm); (3) fine (0.2-1 mm); (4) very fine (0.05-0.2 mm); and (5) dispersed-segregated (less than 0.05 mm).

Titanomagnetite is the main ore mineral and contains 2 to 18 per cent of exsolved ilmenite. Vanadium is present in the titanomagnetite, probably in solid solution as coulsonite (FeV_2O_4) - isostructural with magnetite and highly soluble in that mineral. The ore-grade vanadium-bearing titanomagnetite appears to be confined to the ores in the Gusevogork massif where the V-content ranges between 0.20 and 0.48 per cent and averages 0.42 per cent. The titanomagnetite segregations also contain minor amounts of pyrite and pyrrhotite and even less chalcopyrite, pentlandite, and bornite, plus native platinum and platinoids; some pyroxenes and amphiboles and even less chlorite and biotite form the gangue minerals.

In the various concentrations of titanomagnetite, the iron content averages 15 to 18 per cent and ranges from 14 to 34 per cent. Commercial ore, apparently must contain at least 16 per cent iron. TiO_2 ranges between 0.8 and 2.0 per cent; for the entire deposit, the V_2O_5 lies between 0.05 and 0.31 per cent. The platinum metals are measured in tenths of a gram per ton.

The vanadium is separated from the iron, being concentrated in the convertor slags. Reserves in 1977 are reported to have been 3.5 million tons in Gusevogorsk and 2.6 million tons in Kachkanar proper; the average grade of the ores in iron is 16.6 per cent. To 1977, open-pit mining had been carried out at Gusevogorsk, and the Kachkanar deposits were being prepared for mining. Ore of this grade could not be mined at a profit in the capitalist countries, but different standards appear to apply in the USSR.

The major reserves of titanomagnetite ores are located in the Main, Northern, Western, and Intermediate I subareas of the Gusevogorsk pluton. The greatest vanadium content is in the Western subarea ore segregation (0.1 per cent V). Fominykh and others (1967) says that the average grade of the ores in the Gusevogorsk pluton is: Fe, 16.7 per cent; V, 0.08 per cent; and Ti, 0.73 per cent. Allowing for the grades given by Sokolov and Grigor'ev (1977) probably being expressed in V_2O_5 content and not V, these grades of Borisenko (1977) are not too different from those given by the former two authors. The content of vanadium in the ores becomes greater as the size of the titanomagnetite segregations grows larger. The titanomagnetite may contain as much as 0.48 per cent vanadium. The concentrations of vanadium in the rock-forming minerals is far less than in the titanomagnetite; in hornblende, less than 0.09 per cent; in diopside less than 0.03 per cent; and in the olivine less than 0.003 per cent. In ilmenite, the vanadium content is less than 0.1 per cent; in the sulfides it is negligible.

In the Kachkanar concentrator, the titanomagnetite concentrate contains 0.35 per cent vanadium, the agglomerate (whatever that may mean) 0.4 per cent, and the silicate tailings 0.037 per cent vanadium.

Of the five structural types of ore, the finest contains 0.071 per cent vanadium and the coarsest 0.089 per cent; the change from finest to coarsest in vanadium content is quite gradual.

The ultramafic rocks of the Gusegovorsk pluton are quite similar to those of such districts as Nizhnivtagil' and Noril'sk so it is not surprising that some PGM minerals are contained in the Gusevogorsk titanomagnetite segregations. At Gusevogorsk, the platinum-bearing minerals are most irregularly arranged. The main platinum mineral in this deposit is polyxene that contains 88 per cent platinum and 8.6 per cent iron; other platinum-group minerals, platinierous iridium, iridiferous osmium, copperite, and native osmium are present, but these are far less abundant than polyxene. It is notable that the xenomorphic intergrowths and aggregates of polyxene grains may be as large as 0.2 mm. These platinum-bearing minerals are closely associated with the highest grades of native gold, although nowhere is the gold present in anything but minor amounts. Appreciable amounts of placer platinum-bearing minerals (and presumably of native gold) are known in the Bol'shoi Gusevki rivers.

The Kachkanar ores appear to have been formed entirely during the crystallization process of the ultramafic intrusives. For all practical purposes, the titanomagnetite is a component mineral of the ultramafic rocks in which it is now contained. The titanomagnetite did not come out of solution until late in the crystallization cycle of the ultramafic magma; this is shown by the consistent locations of the titanomagnetite in the interstices among the silicate crystals. The variation in Fe-Ti-V content from one part of the ultramafic rocks is a reflection of the content of these elements in the magma at the time it began to crystallize. The scattered character of the segretations of ore grade in the host rocks is a demonstration of the accuracy of this statement. Where there was enough Fe-Ti-V-oxide material, the amount that crystallized in a given rock volume created ore; where there were lesser amounts of the oxides, the rock is below ore grade. The deposits, therefore, should be categorized as Magmatic-3b, that is, late separation (of the oxide material from the silicate melt) and its late crystallization relative to the silicate minerals. The crystallization history of the PGM must have been essentially the same as that of the Fe-Ti-V oxides.

The Kachkanar-Gusevogorsk intrusion almost certainly is of the same age as those in the Nizhniytagil' and other platiniferous areas of the Middle-Ural which makes them late Paleozoic, and they are so classified here.

NIZHNIYTAGIL', SVERDLOVSK OBLAST

Late Paleozoic	Platinum Metals	Magmatic-3b
(Primary)	(Gold, Placers Only)	(Primary)
Quaternary (Placers)		Metamorphic-C
		Placers-1B, 1C
		(Secondary)

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Notes

The dunite massif that contains the primary deposits of the Nizhnetagil' ultramafic complex centers around 57°55'N,59°58'E and is about 125 km NNW of Sverdlovsk. The complex has its greatest length of at least 13.5 km in a NNE direction. According to Razin's map (1977), modified after Vysotsky and Zavaritsky and Betekhtin, the center of the complex is composed of dunite with a generally rather narrow margin of what Razin calls apo-dunite. By apo-dunite he means modified dunites similar in composition to peridotites, pyroxenites, gabbros, and mafic alkaline rocks; to some degree these apo-dunites are serpentinized. On the east border of the dunite (and its apo-dunitic margin) lies a much larger mass of gabbro-diorite; this rock group also appears on the west side of the dunite, but this western gabbro-diorite is much narrower than that on the east side. Beyond the diorite in essentially all directions, except directly to the east, is apo-gabbro and apo-diorite amphibolite. Minor amounts of the gabbro appear to have been uralitized. The entire igneous complex is surrounded by sedimentary rocks, mica schists, limestones, and crystalline schists. In Quaternary times, platinum and gold placers were developed on the surface of the dunite; they now are worked out. The total arrangement of the ultramafic complex is far more complex than this description indicates, but in general the relative arrangements of the various rocks are correctly stated.

If anyone today knows (or anyone ever knew) how much platinum and gold were recovered from the complex, either from primary ore of placers, nothing is said about it in the Soviet literature.

The chromite segregations in dunite that contain the primary platinum group metals and minerals are confined to pipe-like bodies of this ultramafic rock, of which the Nizhnivtagil' dunite body is best known; these intrusive bodies make up the Ural platiniferous belt. Although most of the dunite massifs are more or less oval in outline, that at Kachkanar (discussed under that heading) appears to be at least larger than the others and perhaps is more lopolithic in character. The depths to which the dunite bodies descend beneath the surface do not appear to be more than 1.5 km.

In the gabbros that surround the pyroxenites (apo-dunites) at Nizhniytagil' in places are quite continuous; in others, they are fragmented. In several of the ultramafic massifs, the pyroxenites occupy much larger volumes than the dunite with which they are associated. The diagrams given by Zavaritsky and Betekhtin show that the dunites are the central and lower portions of the complexes, but locally erosion may have gone so deep, or the dunite masses were so small that they are now completely missing from the eroded remains of the complex. The dunite, in the less eroded complexes, may be covered, in turn, by pyroxenites and gabbros, and even by granites. This rather simple scheme may be complicated in places by the appearance of isolated bodies of dunites and pyroxenites at different levels in the gabbros.

The gabbros (locally they are gabbro-norites) frequently are olivine gabbros, but large parts of the gabbro contain no olivine at all. In such gabbros as contain clinopyroxenes, these minerals are somewhat to appreciably amphibolitized. The gabbros differ considerably in regard to size of grains and character of textures with banded structures, however, being widely present.

Gabbros, in which the monoclinic pyroxene was altered to green or brown hornblende, are designated as gabbro-diorites. Feldspar normally is quite fresh, but it may be, in considerable part, replaced by zoisite. Such gabbros are quite common in the Nizhniytagil' massif proper at Mount Mamynikha. The amphibolitized gabbro may be closely associated with diorites and quartz diorites, but these rocks types are separated from the dunite by a melanocratic phase of the gabbro that forms a transition zone with pyroxenites. Locally, these rocks (known as tilaites) include biotite and potash feldspar in their composition.

The pyroxenites and peridotites that enclose the dunite bodies are composed of monoclinic pyroxene and lesser amounts of olivine; some xenomorphic magnetite may be present. In rare instances, the pyroxenites may be associated with hornblendites.

The constitutents of the dunites are olivine and chromite, and the former are normally more or less serpentinized.

Dike rocks of a considerable variety cross-cut some or all of the major rocks of the complex. The dike-types that cut gabbros and diorites are designated as aplites and lamprophyres, whereas those occupying similar relations to dunites are called issites (i.e., hypabyssal rocks composed of hornblende with minor amounts of green pyroxene and even less labradorite, plus accessory magnetite and apatite) - the term is no longer recommended by the Glossary of Geology. Other dike rocks cutting the dunites are gabbro-pegmatites, plagioclasites (i.e., anorthosites), and albitites (i.e., albite phenocrysts in an albite groundmass). Although most of the dunites contain these dike rocks, the Tagil' massif (the largest) completely lacks them because it has been deeply eroded enough to have removed the issite veins that normally are confined to the upper parts of the dunite bodies.

In the southern part of the dunite of the Nizhiyntagil' massif, the igneous body is penetrated by tongues or bands of pyroxenite that extend diagonally northeast and northwest, following joint planes in the dunite. The gabbros that lie just outside the narrow pyroxenite band on the west side of the dunite body have been highly altered and now are green schists. On the east side of the dunite, the gabbro has been altered to gabbro diorite that exhibits a strong banding that suggests that the original gabbro magma flowed around the solid dunite-minor pyroxenite body when the gabbro was injected.

Near the present surface, the dunite is partially serpentinized with the alteration extending out from the same set of joint planes that localized the bands of pyroxenite in the dunite. A considerable number of these bands contain serpentine that has no remaining trace of its original mesh texture. Between the dunite and its narrow enclosing rim of pyroxenite is a band of antigorite (serpentine) that also shows no remnant structure of dunite; brucite is a characteristic mineral of this antigorite-rich rocks.

Drilling carried out in the early days of geological examination of the dunite massif shows that serpentinization ends at about 450 m beneath the surface; below this depth, the dunite is composed entirely of olivine except for 1.5 to 2.0 per cent of chromite.

At a depth of 500 m, the chemical composition of the dunite was determined by Zavaritsky and Betekhtin to be: SiO_2, 40.03%, Al_2O_3, 0.57%; Cr_2O_3, 0.50%; Fe_2O_3, not determined; FeO, 8.29%; MnO, 0.06%, MgO, 48.8%; CaO, absent; loss by ignition 1.28%. One thing is obvious from this analysis - the olivine is forsterite.

In this primary dunite, the olivine texture is euhedral as are the grains of chromite. Locally, vugs have been encountered in the dunite that contain amorphous serpentine, but, between the olivine and the serpentine, the walls are overgrown with uvarovite (Cr-garnet), Cr-chlorites, Cr-vesuvianite, Cr-diopside.

Mertie (1969) reports that at least 600 of various sizes and shapes of chromite bodies were found in the Nizhniytagil' massif; although many of these contain platinum or platinum-bearing alloys, the majority do not. The size of these bodies ranges from small pockets 1 to 2 cm in diameter (and not included in the figure of 600 given above) to large, elongated masses that may have vertical dimensions of 20 to 30 m. The Gos-shakhta (State shaft) deposit reached a depth of 150 m and had a diameter of 6 to 7 m with, of course, the Pt-bearing portion of this pipe composing only a small part of its total volume. The 600 ore bodies (or barren masses) were distributed quite irregularly in the dunite; in places they occur in groups of two or more, but single ones are quite common. The bodies that are associated in groups may all be morphologically quite similar or differ greatly in shape, structure, mineralogy, PGM content, and chemical composition of the chromite of which they are largely made up.

Apparently, the original dunite magma contained Fe^{+2}, and $Cr_2O_4^{-2}$ ions dissolved in it. The platinum-group elements probably were carried in the metallic state, probably as native metals in solution in the silicate melt. As the melt temperature dropped the silicate minerals, almost entirely olivine, in the restricted volume of that melt, crystallized. Being heavier than the melt, they tended to sink downward, but crystallization appears to have been so rapid that the network of olivine crytals shortly achieved approximate structural stability, although sizeable pockets of molten material remained trapped in various parts of the now considerably solidified dunite magma. Some of the Fe^{+2} and $Cr_2O_4^{-2}$ reached a concentration in the dominantly silicate melt that they could no longer be held in solution in the melt and separated from it, probably in the molten state. This molten chromite accumulated as accessory minerals in the interstices among the olivine crystals or in the larger pockets that existed in the dunite. The platinum-group minerals, however, appear to have remained in solution in the silicate melt, as the early separated chromite is particularly poor in PGM.

As the mainly olivine crystallization continued, the proportion of chromite and PGM in the residual melt increased in relation to the silicates. Eventually the chromite-PGM portion of the melt became almost entirely immiscible in the molten silicate fraction, and the two melts existed side-by-side with the chromite and PGM melts being combined into one melt.

Not all of these chromite-rich melts, but any means, contained enough molten PGM alloy that, when the chromite melt crystallized, to become ore; it included too little PGM ever to be considered as ore of the platinum-group metals. In other instances, however, there were appreciable amounts of PGM in the chromite, and these metals crystallized anhedrally in the chromite mass in which they were incorporated.

From the data given above, it appears that early in the crystallization cycle of the dunite magma, a chromite-rich molten fraction began to separate from the dominant silicate melt. Further, the PGM-Fe content of the silicate magma was unacceptable in either silicate or chromite structures, but the PGM-Fe component appears to have been, initially at least, more acceptable in the silicate melt than in the chromite melt because the chromite that crystallized early (and was incorporated among the olivine crystals) is essentially barren of PGM. Later in the crystallization cycle of the dunite magma, the content of PGM (small though it was) became large enough that some (or most) of it became immiscible in the silicate melt. This molten PGM seems to have been acceptable in mutual solution in the chromite melt in contact with the silicate melt, so that the later chromite contained PGM that crystallized anhedrally in the larger masses of chromite (the formation of which will be discussed below). This PGM-chromite relationship is shown clearly in illustrations in Schneiderhön (1958).

What was the fate of this later PGM-chromite melt? Much of the PGM-bearing chromite is (or was) found in PGM-chromite schlieren, normally contained in steeply dipping pipes or lenses. Apparently the fracture necessary to localize these pipes were formed after the olivine has crystallized sufficiently to break under stress. Into these fractures zones, PGM-chromite melt was injected in such a manner as to promote the arrangement of the chromite in these schlieren structure. These pipes have been followed downward for 10's to 100's of meters.

In other situations, pods or lenses of PGM-chromite melt appear to have accumulated in the solid olivine and remained there until they, in turn, had solidified, not having been moved essentially from the places in which they initially were entrapped. Not all of these pods, as has been mentioned, were rich enough in PGM to be worth mining, but they probably were composed of chromite that had settled out early from the silicate melt, before PGM was available to be mixed with it.

Thus, the chromite shows three types of bodies. Type (1) was the pods of lenses of chromite that separated early from the silicate melt and, therefore had no PGM associated with them.

Type (2) was the schlieren, nests, and branching veins of PGM-bearing chromite created when the molten material was driven into structures created by tectonic stress after the dunite was solid enough to fracture. Some of these fractured bodies were so extensive, that the PGM-bearing chromite acts as a breccia cement or extended to such depths as to be major pipes.

Type (3) ores are the pods of chromite that accumulated late in the crystallization cycle of the dunite magma (and therefore, contained PGM) and were not moved by tectonic forces after they had been collected.

Thus, only types (2) and (3) of the chromite accumulations are (or were) PGM-bearing. Type (1) of the accumulations separated too early in the chain of solidification events to contain PGM.

In contrast to most workable bodies of PGM elsewhere in the world, those in the Nizhniytagil' massif are composed almost entirely of alloys or solid solutions of PGM. PGM compounds are negligible to lacking in the Nizhniytagil' ores. These solid solutions are referred to by Soviet geologists as "polyxene", a term not recognized by Fleischer. The most abundant alloy is iridium-rich polyxene, in which the components are: (1) Pt 77 to 81 per cent; (2) Pd 0.2 per cent; (3) rhodium 0.4 to 2.1 per cent; (4) Ir 7.0 to 7.6 per cent. Minor minerals are: platiniferous iridium and osmiferous iridium; the platinum content of these two alloys is 13 per cent and 2 to 5 per cent, respectively. Of PGM sulfides, only laurite (RuS_2) has been reported; it contains essentially no platinum or palladium.

Razin (1977) believes that the Nizhniytagil' dunite body, prior to erosion, in its upper levels contained platinum sulfides and arsenides [such as are present in the Ingaly type of Pt-deposit (Amur Oblast, 54°21'N, 124°32'E)]; this concept, of course is difficult to prove. It may simply be that the Nizhniytagil' deposit contained such ratios of PGM and sulfur and arsenic that all of the Nizhniytagil' mineralization was the same as what is now known, and the same reasoning applies to the Ingaly type.

The classification of the Nizhniytagil' deposits is based on the apparent fact that the PGM did not separate from the dunite melt until late in the crystallization cycle and, therefore, did not themselves crystallize until late. Thus, these deposits should be categorized as Magmatic-3b. Since the location of the PGM, if not their formation, depends in large measure on metamorphic (tectonic) effects, the term Metamorphic-C should be added to the classification.

The final event in the solidification history of the dunite was the conversion mainly of the border areas of the Nizhniytagil' dunite to serpentine, plus lesser amounts of calcium and magnesium carbonates, brucite, magnetite, and even rare native copper. Very little PGM ever have been found in the serpentine. This process of serpentinization probably was autometamorphic (deuteric), although a hydrothermal genesis for this serpentine has been postulated. As these deuteric reactions did not affect the PGM content of the Nizhniytagil' ores, Magmatic-4 is not included in the classification of the deposit.

The placer deposits derived from the primary bodies of the Nizhniytagil' massif in all probability were developed during Quaternary time, both Pleistocene and Recent. They should be classfied as deposits mechanically concentrated, being contained in both eluvial and alluvial placers, that is Placers-1B and 1C. Although gold apparently never was recovered from the primary ore at Nizhniytagil', enough was obtained from the Ural placers associated with the dunite massif to include gold among the metals recovered from the deposit.

Igneous activity in the Urals, which reached a climax during the junction of the European and Asian plates at the end of the Paleozoic, almost certainly was responsible for the primary PGM deposits of Nizhniytagil', and they should be classified as late Paleozoic in age.

KERCH, UKRAINE SSR

Quaternary	Iron	Sedimentary-Ala Ground Water-B2

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Notes

The Kerch Peninsula forms the eastern extension of the Crimea and provides the western one of the two peninsulas that almost shut off the Sea of Azov from the Black Sea. The city of Kerch (45°22'N,36°27'E) is near the eastern tip of the peninsula that is bordered by various ore bodies (and the spaces between them). The total length of the mineralized zone is at least 130 km from the most westerly expression to the farthest east ore body on the Sea of Azov. In addition to the belt of ore bodies that occurs almost continuously along the north coast of the Kerch peninsula, ore bodies are present in a continuation of the north-coast belt along the east coast bordering the Kerch Strait (Kerchenskiy Proliv) for essentially its entire length of some 40 km (straight line). The mineralized area most centrally located is the Novoselovsk depression that is 30 km from the east coast, 25 km from the north coast, and about the same distance from the south coast; it is quite isolated from the other ore bodies on the peninsula.

For all practical purposes, the rocks of geologic interest in the ore-bearing area of the Crimea are Cretaceous, Miocene, and Pliocene, where one of the finest records of the Cretaceous and Tertiary periods is to be seen. The Crimea can be considered as a northwestern extension of the Caucasian fold belt, the southern half of which in the longitude of the Crimea has been sunk beneath the Black Sea. Physiographically, the Crimea can be divided into two parts. By far the larger part, the one comprising about 80 per cent of this land area is the Crimean steppes where the surface rocks are Pliocene in age. The really mountainous part makes up the rocks along the southern coast, extending as far east as the beginning of the Kerch peninsula proper. As these mountains are crossed from north to south, they begin as rounded hills, but the structures become steeper to the south and end in steep cliffs along the coast-resort area of which Yalta is the best known locality. The Crimean mountains end sharply on the south, sloping steeply off into deep water, perhaps the result of a great graben developed during the Cenozoic. The Crimean mountains are of the Alpine type but almost certainly are not connected with the Alpine structures of the more westerly parts of Europe in a continuous chain.

The Crimean steppes, of which the north section of the island is a part, have been thought to be a fore-deep directly to the north of the mountain range. Although the mountains are mainly composed of Mesozoic rocks, the fore-deep is filled with a rather appreciable thickness of Paleogene, Neogene, and Quaternary sediments. The building of the Crimean mountains seems to have coincided with that of the Carpathian and Balkan mountains and were developed from mid-Cretaceous to mid-Tertiary times. The mountain system is a complex anticlinorium, whereas the steppes essentially are a subsided portion of the east European platform in which the thick Tertiary sediments lie on a Paleozoic and Precambrian floor.

The geologic map of the Kerch peninsula shows an appreciable band of Pliocene sediments along the western part of the north coast; in the eastern third of the peninsula, the Pliocene beds are confined to a narrow band (a few km wide) that reached the Kerch strait at the city of that name. To the south of this band are Miocene formations; these formations contain inliers of Cretaceous rocks. These inliers do not conform, even roughly, to the locations of the iron ore deposits of the peninsula.

The iron ores of the Kerch peninsula are of two types; both are concentrations of sedimentary ollitic iron ores. [In the translation of Academician Smirnov's "Ore Deposits of the USSR", the the translator says that the ores are in rocks of Kimmeridgian age. Actually the Russian text says that are they are middle Kimmerian, Kimmerian referring to tectonic movements and not to the age of formations.] Since, the ores, in places, are underlain by Pontian limestones, they cannot be older than oldest Pleistocene. [Routhier says that they are Pliocene in age which is much closer to the truth than the translator's statement that they are Kimmeridgian, that is, uppermost Jurassic, which they cannot be.] It would seem, from the Russian text, that Sokolov and Grigor'ev (1977) refer to "Kimmerian" as a stage and not as a designation of a series, over considerable time, of tectonic movements. At any event it seems clear that the ores are either early Pleistocene or latest Pliocene with the odds favoring the former.

The first of the two ore-containing depressions are large brachysynclinal structures (troughs) and the second are pseudotectonic downwarps or, more properly, compensation downwarps in the zone of development of mud volcanism (these were designated as depressed synclines by Shnykov, 1965).

In each of the depressions (troughs and downwarps) the "Kimmerian" ore layer (in the Russian sense) is underlain by either Pontian-stage limestones or clays of the lower "Kimmerian". The ores are overlain by clays of the upper "Kimmerian". Although some of the ore actually outcrops at the surface along the margins of the depressions, it normally extends to depths of as much as 70 to 100 m or even, in the centers of the troughs, to some 140 to 180 m. The ore layers are 25 to 40 m thick in the central parts of the depressions but are as thin as 0.5 m along the ore deposit margins.

In either of the two types of depressions, there are two types of ore. Those ores that are primary, or essentially so, have not been subjected to the oxidizing reactions caused by surface waters and so retain very much their primary mineral character. These primary or "tobacco" ores have, to a considerable extent, been preserved because waters under high pressures rise from the Pontian aquifer under the "Kimmerian" beds, entering joints, cavities, and capillaries in the ore layer, thus preventing the oxidation of the tobacco ores that would occur if surface, oxygen-bearing waters could get at them. This Pontian aquifer normally is confined to the central parts of the two types of depressions which makes the outer annular zone of the deposits most subject to oxidation.

The stratigraphic section in a typical ore-bearing depression is that of the Kamysh-Burun trough that is located a short distance southwest of Kerch. At the base of this section is a layer of argillaceous coquina, the top of which contains appreciable siderite. The next layer (2) is, in the basin centers, composed of secondary brown ore. The next layer (3) at the same stratigraphic level as layer (2), but up dip from it, is made up of tobacco ores and reaches almost up to the outcrop but is very narrow there. Between layer (3) and layer (4) are isolated spots of sandy clays and argillaceous sands with abundant bivalve shells. Layer (4) is directly above the brown ore and is roe ore; it dies out up dip at some distance from the outcrop. Layer (5) is directly above the roe ore and grades up dip from tobacco ore into brown ore. Layer (6) lies above all of the sequence through layer (5) except where, near the center of the basin, it has been nearly or completely cut out and filled with the clays of layer (6); these layer (6) clays reach almost up to the outcrop. Layer (7) is sandy clays that fill valleys in layer (6) toward the center of the trough and lie over it rather smoothly from there almost to the surface. Layer (8) is composed of calcareous loams and covers all the other layers from the center of the depression to the outcrop.

The "brown" (or oxidized) ores are, for reasons just outlined, confined to the upper levels of the depression-contained ores and usually completely rim the trough or downwarps in which they are collected. Locally, however, these brown ores appear to have formed at depth where channels permitted the influx of surface waters.

In addition to these two ore types, a minor and very thinly scattered variety is the mangano-siderite-rhodochrosite oolites and concretions (roe ores); these appear to be distinguishable from the oxidized brown ores only by their appreciably higher Mn-content. Although now only brown ores are mined, some roe ore may be recovered with them, but no separate manganese product appears to be developed from them.

The tobacco ores contain, as major minerals, hydroferrichlorite [probably iron-rich clinochlore or chamosite - $(Fe^2_5Al)(Si_3Al)O_{10}(OH)_8$], although this is not certain, ferrimontmorillonite [apparently montmorillonite in which Fe^2 is substituted for Mg], and hydrogoethite [optically identical with lepidocrocite - $FeO(OH)$ with adsorped or capillary water added to the lepidocrocite formula], in what are described as "ocherous and compact varieties". This ore also contains carbonates of the manganosiderite-rhodochrosite series. The oolites commonly contain fragments of quartz and feldspar. Locally, the phosphates, vivianite and kerchenite (= kertschenite an oxidation product of vivianite), and hydroxides of manganese; psilomelane (probably used here in the sense of romanechite $[BaMn^2(Mn_8O_{16})(OH)_4]$, vernadite $[(\text{delta-}MnO_2)\text{-}(Mn^4,Fe^3,Ca,Na)(O,OH)_2 \cdot nH_2O]$, and pyrolusite. Realgar has been noted rarely.

In the brown ores, the main minerals are hydrogeothite and ferrimontmorillonite with psilomelane (romanechite), gypsum, aragonite, calcite, kerchenite, pyrite, quartz, feldspar, and glauconite. The ore fabric of these ores is oolitic and pisolitic.

Thus far (1977), no method had been developed for benefitiating the tobacco ores to a useable product, so they are not mined; only the brown ores are.

The grade of the brown ores is: Fe, 37.7 wt. per cent; MnO, 3 per cent; SiO_2, 18 per cent; Al_2O_3, 5 per cent; MgO, 1 per cent, CaO, 1.75 per cent; V_2O_5, 1.2 per cent; P, 1 per cent; S, 0.07 per cent, As, 0.13 per cent.

The reserves of both ore types in the peninsula were estimated (1977) as being 1.7 billion metric tons with .0.517 billion tons being brown ores. Open-pit workings are employed at the major Kamysh-Burun and El'tigen-Ortel deposits.

Smirnov [1977 (1962)] divides sedimentary iron ores into three types; (1) oxide, (2) carbonate, and (3) silicate. The major difference between the two ores types (brown and tobacco) in the Kerch deposits is that, in the primary tobacco ores, hydroferrichlorite (or chamosite ?) is present. In the secondary brown ores, the chamosite has been eliminated with, presumably, the content of the other two principal minerals, probably mainly hydrogoethite, having been increased. Thus, both of the ore types in the Kerch area are mixtures of Smirnov's types (1) and (3) with type (2) ores (as has been mentioned) being far less important than either of the other two.

The differences between what have been designated as "brown" and "tobacco" ores may actually have been original sedimentary differences with the brown oxide-silicate mixture of the deposits having been formed around the margins of the depression-basins in the primary oxidizing zone and the tobacco oxide more silicate ores having been developed in the more central parts of the troughs.

On the contrary, it seems more probable that the original extent of the primary ore bodies was appreciably greater than the extent they have now, the outer portions of the individual troughs having been removed by erosion, turning what once were a few or several much larger ore bodies than those at present known into the much smaller ones that now remain.

If the later explanation is the correct one, the present character of the brown ores was produced by secondary, near-surface waters attacking, particularly, the hydroferrichlorite and converting it into hydrogoethite, thus turning the tobacco ore into brown by raising the Fe-grade.

It appears most probable that the erosive processes that put the materials from which the ores were deposited into the surface-water circulation took place in the then more extensive Pleistocene (or Pliocene) Crimean Mountains and that these waters were brought northward into the extensive Pleistocene sea that covered what are now the Crimean steppes and there deposited the ores. The ancestral Crimean mountains to the south later were dropped into what is now the Black Sea graben.

The possibility of some (or all) of the iron having been provided from volcanic emanations seems unlikely in view of the lack of volcanic materials in the stratigraphic section of the Pliocene-Pleistocene in the Kerch peninsula portion of the Crimea.

On the basis of this reasoning, the Kerch orion ore deposits should be classified as "deposits chemically concentrated by interaction of solutions - inorganic reactions", that is Sedimentary-Ala.

KRIVOY ROG, UKRAINE SSR

Middle Precambrian	Iron as martite, hematite, hydrohematite, hydrogoethite, magnetite	Sedimentary-Ala Metamorphic-C Hydrothermal, (probably Hypothermal-1) Metamorphic-C

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Notes

The Krivoy Rog iron ore basin, the second largest concentration of iron ore in the Soviet Union after the Kursk Magnetic Anomaly, runs in a slightly east of north direction for some 325 km. The city of Krivoy Rog is located at 47°55'N,33°24'E, and the northern portion of the ore basin lies between that city and Kremenchug, (49°03'N,33°25'E), a distance of some 130 km. Kremenchug lies on the north of the artificially drowned valley of the Dnepr river and some 100 km southwest of the battlefield of Poltava where the forces of Peter the Great defeated those of Charles XII of Sweden in June of 1709, effectively ending the dreams of Swedish hegemony in north-eastern Europe. Occurrences of ferruginous rocks of the belt extend for another 125 km to the south of Krivoy Rog and some 70 km north of Kremenchug.

The Krivoy Rog basin lies in the western part of the Dnepropetrovsk region in the drainage areas of the Ingulets river and its tributaries, the Saksagan and Zheltaya rivers. The basin ranges from 2 to 7 km in width. The major mineable ores are enriched iron ores than range around 46 per cent iron; huge volumes of ferruginous quartzites contain magnetite in sufficient quantities to give them a grade of about 10 per cent iron.

The discovery of the ores of Krivoy Rog took place in the 1770's, and much fragmentary geologic work was done in the next 100 years, but it was not until near the end of the 19th century that detailed and accurate geologic work was begun in the ore basin. In the 1930's, a permanent geological survey was established for the Krivoy Rog ores, and this body, assisted by geologists from numerous other scientific organizations, made remarkable advances in the finding and understanding of the ores. The principal events in the period since 1930 were: (1) the deciphering of the multi-bed stratigraphic scheme that made possible the understanding of the structure of the basin; (2) the theoretical basis for searching for deposits not blessed with surface outcrops; (3) the development of the metamorphic theory for the formation of high- and low-grade ores; and (4) the study of the petrography, petrology, metamorphism, and mineralogy of the rocks and the ores contained in them.

The large-scale mining of the high-grade ores of the Krivoy Rog basin began in 1875 and was confined to open-pit operations. Currently, 12 commercial enterprises are engaged in the underground mining of iron ore; further, the five largest mining organizations in the USSR - Inguletskii, Yuzhnyi, Novo-Krivorozhskii, Tsental'nyi, and Severnyi - apparently confine themselves to the open-pit mining and processing of ferruginous quartzites.

By the 100th anniversary of the beginning of mining in the basin, 1975, 2 billion tons of ore had been extracted from the various deposits.

The rocks of the Krivoy Rog basin are Archean and Proterozoic crystalline rocks of the Ukrainian shield; the only other rocks in the area are of Cenozoic age. Within the Krivoy Rog basin, the central portion that contains the iron-bearing Proterozic beds is bordered on the east and west by highly folded and metamorphosed Archean rocks that consist of gneisses, granites, migmatites, amphibolites, and schists. The sedimentary rocks of Proterozic age have been folded into a complex geosyncline; these rocks are divided into (from bottom to top): (1) lower arkose-quartzite and phyllite sequence; (2) middle ore-bearing ferruginous quartzite and slate sequence; and (3) upper quartzite-sandstone-slate sequence that contains lenses of marble (originally limestone). To indicate how complex the stratigraphy is, Belevtsev (1962) divided sequence (2) into seven to nine horizons of ferruginous quartzites with which are interbedded a variety of slates, such as quartz-sericite, chlorite-sericite, and others, and microquartzites. This group of rocks is known as the Krivorozhakaya series, and its major members (from top to bottom) are: (1) lower iron ore, (2) middle schistose, (3) upper iron ore, (4) quartzite-sandstone, and (5) schistose. The upper four ferruginous beds (as recognized by Belevtsev, 1982) are contained in the upper iron ore with the third ferruginous being contained in the middle schistose and the first and second ferruginous in the lower iron ore.

In detail, the seven ferruginous members differ appreciably, one from the others. All of them, however, appear to have been derived by hydrothermal alteration of ferruginous quartzites and their later (or even concomitant) metamorphism, of which more later.

Most of the richer ores are concentrated within the fifth and sixth ferruginous quartzites and schists in what is known as the Saksaganskaya formation, that is the beds containing the seven ferruginous quartzite horizons.

The Krivorozhskaya series has been divided by Belevtsev (1982) into five formations (from the bottom upward): (1) Novokrivorozhskaya, (2) Skelevatskaya, (3) Saksaganskaya (the major ore-bearer mentioned above), (4) Gdantsevskaya, and (5) Gleevatskaya. Formation (1) is composed principally of metabasites; amphibolites, quartz-biotite schists, sericite metasandstones, quartzites, and tuff schists are of lesser importance. This formation is as much as 2000 m thick.

The lower part of formation (2) is made up of quartz metasandstones and arkoses that locally become metaconglomerates. In their higher portions, the beds mainly are schists. Laterally, the formation grades into a carbonate-talc-sandstone horizon. Normally, the thickness of formation (2) is 50 to 60 m; in a few places it may be as much as 250 to 300 m thick.

Formation (3), the one that contains the seven ferruginous quartzites, has these alternating regularly with seven horizons of schists. The iron-bearing horizons are variously composed of magnetite streaks and layers of hematite-magnetite, silicate magnetite, martite, hematite-martite quartzites, and jaspillites. For all practical purposes all the iron ores of the basin are confined to this formation. The greatest thickness of the formation is 1500 m in the central part of the Saksagan' district (Komintern and Frunze mines).

Formation (4) is largely made up of schists, metasandstones, feldspar quartzites that lack ore, and dolomites. Belevtsev has been able to divide this formation into two members. At the contact between formations (3) and (4) are rich beds of ore of the Ingulets type.

Formation (5) is composed of metasandstones, schists, and metaconglomerates; this formation is divided into four members.

Structurally, the rocks of the Krivoy Rog basin (Krivbas) an intricately folded synclinorium, the axis of which has a more or less north-south strike and is known as the Krivorozhskii first-order structure. The structures are essentially isoclinal and are complicated by further folding of higher orders of which the major examples are the Osnovnaya and Saksagansk synclines and the Tarapako-Likhmanovsk and Saksagansk anticlines, the western limbs of which are cut by steep thrust and normal faults. Farther south, the basin consists of the Ingulets (Likhmanovsk) syncline, the western limb of which has been cut off by a steep thrust fault. The syncline structure is best seen in the neighborhood of the Iljich and Dzerzhinskii iron mines. The axis of the synclines plunge to the north at as much as 40°. Dips on the limbs of the synclinal and anticlinal folds dip between 45° and 80°. The largest of the overthrusts cutting the folds of the Krivbas is the Saksagan' overthrust.

The ore bodies of the basin are (from south to north): (1) the Ingulets (Southern) ore field, (2) the Saksagan'(Main) ore field, (3) the Pervomaisk, Annovsk, and Zheltorechensk (collectively the Northern) ore field, and (4) the Popel'nostovsk ore field. In the Northern ore field, the Proterozic rocks are cut by granite intrusions, also of Proterozic age.

The ferruginous quartzites consist of magnetite, magnetite-hematite, and hematite varieties, but they are of commercial importance only (at present) where they are made up of the unoxidized magnetite and magnetite-hematite types. From the huge volumes of these two types of ferruginous quartzites put through the concentrating plant, it is possible to make a high-quality magnetite concentrate.

The rich ores, on the other hand, are made up largely of iron oxides and hydroxides that are irregularly distributed among the ferruginous quartzites and have such forms as layers, columns, stocks, and lenses.

In the Main (Saksagan') ore field, the ores are confined entirely to the ferruginous quartzite layers, principally on the limbs and troughs of the synclines. The ore bodies on the limbs of the folds appear to be located most commonly where these parts of the synclines have been crumpled by transverse zones of gentle folds. Such tectonic structures generally are accompanied by the development of cleavage and the movement of beds relative to each other and by faults along the strike of the bedding and in zones (or nodes) of crushing and jointing. Since these tectonic structures are not equally developed throughout the various ferruginous-quartzite horizons, the ore bodies are separated from each other by less altered (metamorphosed) volumes of ferruginous quartzites. These individual ore bodies, however, appear to coalesce as the troughs of the synclines are approached and agree in strike and dip with those of the synclinal troughs.

The iron ores of the Krivbas usually are subdivided into: (1) rich ores in which the iron content ranges between 46 and 70 per cent with the average being about 57.5 per cent and (2) ferruginous-quartzite ore in which the grade ranges between 10 and 45 per cent with the average being about 34.3 per cent. The rich ores are not subjected to mineral-dressing processes, but the ferruginous quartzite ores must be so beneficiated.

The rich iron ores differ in the ore minerals contained in them; the following are examples of these differences: (1) 'blue' ore - martite and hematite-martite; (2) 'red-blue' ore - martite-hematite-hydrohematite-hydrogoethite; (3) 'red' ore - hematite-hydrohematite-hydrogoethite, and (4) (in the Northern ore field only) magnetite and magnetite-speculatite. Further mineralogical differences exist among these four types. Type (1) also contains chlorite, sericite, pyrite, carbonates, clay minerals, quartz, and apatite in minor quantities. Type (2) ore includes also sericite, quartz, clay minerals, alunite, and sphene as trace and accessory minerals. Type (3) additionally has kaolinite, clay minerals, chlorite, and carbonates. Type (4) also contains amphiboles, aegirite, biotite, albite, quartz, carbonates, and chlorite, isolated grains of pyrite, pyrrhotite, and chalcopyrite have been seen. The average content of iron in these ore types is; (1) 63.7% (2) 62.3%, (3) 57.5% and (4) 54.0%. In phosphorus, the content is: (1) 0.26%, (2) 0.08%, (3) 0.088%, (4) 0.04%. On sulfur, the content is: (1) 0.043%, (2) 0.03%, (3) negligible, (4) 0.15%.

In the Krivbas, there is a definite relationship in mineral composition between those of the enclosing ferruginous quartzite and schist and of the rich iron ore so enclosed. The fifth and sixth ferruginous horizons (counting from the bottom of Saksaganskaya formation) contain most of the rich iron ores of the basin. In the fourth horizon, rich ores are present in the southern mines of the district (the Dzerzhinskii, Kirov, and K. Liebknecht mines). Still other rich deposits lie along the contact zone between the Saksaganskaya and the Gdantsevskaya formation above it; some rich ores are present in basal metasandstones, barren quartzites, and schists of the Gdantsevskaya formation.

The major rich ore mines of the Saksagansk type in the Krivbas are (from south to north): (1) V. I. Lenin; (2) R. Luxemburg; (3) 20th Party Congress (pit); (4) Frunze; (5) Komintern; (6) K. Liebknecht; (7) Kirov; and (8) Dzerzhinskii. They contain 85 per cent of the rich iron ore reserves and amount (down to a depth of 1500 m) to just over 1100 million tons. The 14 major rich iron ores of the Ingulets types are located mainly in the Krivorozhskii basin, and the most important of these are confined to the Likhmanosk syncline and to the nose of the Osnovnaya syncline. Their reserves (down to a depth of 400 to 800 m) amount to 81 million tons.

The rich iron ores of the Perovmaisk type occur in the northern part of the Krivbas (1st of May and Zheltaya Reckka mines). The reserves of this type are about 200 million tons, and they will be mined (together with the associated ferruginous quartzites) by underground methods.

The ferruginous quartzite reserves belong mainly to the Skelevatsk type that eventually will be of commercial value. They consist of two types: (1) non-oxidized ores - magnetite, iron micaceous magnetite, silicate-magnetite, and siderite-magnetite and (2) oxidized ores - martite, ferromicaceous martite, goethite-hydrogoethite-martite, disseminated hematite-martite. The potentially workable ferruginous quartzites can be, on a geological basis, divided into: (1) deposits in the closed portions of folded structures - their strike lengths are 2.0 to 2.5 km and they are from 50 to 400 m in horizontal thickness in fold limbs and up to 100 m in the troughs of the folds; (2) deposits confined to the monoclinal limbs of folded structures;

they may be tens of kilometers long and 30 to 400 m thick; (3) deposits in rock volumes where the folded structures have been transversely deformed into blocks that exhibit a homoblastic texture; and (4) deposits of outliers of various structures engulfed in granitoidal rocks.

These quartzites do not differ greatly in mineral composition, although local lateral and vertical variations may appear in a single deposit. These deposits may be oxidized down to as much as 1500 m, and locally may be highly leached.

The metamorphism of the rocks of the Krivbas is in the range of the green-schist facies in the Saksagansk district but is in the epidote-amphibole facies in the Southern and Northern districts. The reserves of oxidized ferruginous quartzites are over 1 billion tons, and mining has begun on such bodies. The reserves of non-oxidized quartzites in the Krivorozhskii basin are more than 17 billion tons.

The formation of the ores of the Krivbas was a multistage process (Belevtsev and others, 1982). The first step was the sedimentary development of the parent rocks of the ferruginous quartzites. The second step was the further introduction of iron and the development of major mineral transformations under regional and dynamothermal metamorphism. During the green-schist facies development, SiO_2 was leached from certain tectonically favorable zones in the ferruginous rocks; this resulted in a rise in the grade of iron and a reduction in rock volume with the iron being converted mainly to magnetite. Where rocks in this metamorphic stage did not lose silica, the end product was low-grade ferruginous quartzites or jaspillites. The third step was the changes of the iron-bearing rocks in the epidote-amphibole facies; these resulted in an intense migration of iron and its concentration by replacement of large portions of the rock volumes of ferruginous quartzite with sufficient iron to turn the quartzites into high-grade ore.

Deep-seated oxidation of the secondary rich iron ores resulted in silica removal, the oxidation of magnetite to hematite, martite, and various hydro-iron minerals, and the decomposition ferruginous carbonates and silicates. Such abyssally oxidized ores (martite-type) have been followed downward in the Saksagansk district for 2500 m.

It would seem that hydrothermal fluids (of whatever manner of genesis) must have taken a far larger part in the enrichment of the Krivbas ferruginous quartzites than Belevtsev seems to believe. This hydrothermal transformation probably took place at quite high temperatures.

Thus, the classification of the ores has several parts, the first of which is Sedimentary-A1a which assumes that the materials of which the ferruginous quartzites initially were composed were brought in in solutions that reacted with each other to produce the initial minerals of what were later to be (after appreciable metamorphism) ferruginous quartzites. The next step was essentially regional metamorphism in which the major event was the conversion of the initial sedimentary materials to ferruginous quartzites; this step did not appearently include much help from thermal waters and is classified as Metamorphic-C. The third step was the removal of silica and the concentration of iron by high-temperature solutions of unknown genesis and is loosely classified as "hydrothermal". This hydrothermal (probably "Hypothermal-1) phase appears to have affected the entire basin and was not confined to the Northern ore field. Hydrothermal activity in the Northern field probably was longer continued and was more intense as is indicated by the epidote-amphibole facies exhibited in that portion of the Krivasbas. Finally, the abyssal oxidation must have been carried out by solutions capable of appreciable oxidation, though not necessarily of surface origin, although they may have been so produced. From the diagram given by Sokolov and Grigor'iev (1977), it would appear that the total volume of porous, semi-friable ores "of deepweathering crust" in the district must come near equalling that of all of the volumes of all other ore types combined.

One problem that does not appear to be fully discussed in the Soviet literature on Krivoy Rog is: how much iron, if any, has been added from exogenous sources to the original iron formation of the Krivbas to produce the total amount of iron now known in the basin deposits? Is it possible to have developed the huge amounts of various enriched material without having had available additional iron above that initially present? To reach anything near an accurate conclusion, much detailed analysis of the present rock types and of completely unmetamorphosed iron formation would be necessary as well as estimates of the actual quantities of the various ore types that were present when the deposits were first discovered.

It is unreasonable to attempt to reach a conclusion on this matter with the data currently at hand, but a sound determination of the amount of iron added from outside sources (if any) is essential if a full understanding of the genesis of the Krivoy Rog ores is to be attained.

NIKOPOL', UKRAINE SSSR

Early Tertiary	Manganese	Sedimentary-A3, A4 or Sedimentary-A1, A2, A4 plus, for both, Ground Water-B2

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Notes

The city of Nikopol' (at 47°34'N,35°25'E) is about in the center of the Nikopol' manganese mineral belt that extends from Ingulets (47°43'N, 33°16'E) to Bol'she Tokmak (47°13'N,35°43'E), a distance of about 170 km (straight line or 250 km along the arc of which the 170 km is the cord). In this 250 km arc, the distance actually occupied by Mn-bearing ore beds is about 25 km. The district is divided into three parts: (1) West Nikopol', (2) East Nikopol' and (3) Bol'she Tokmak. The western and eastern Nikopol' areas are separated by an uplift volume of the crystalline basement, as a result of which the beds in the western area do not continue directly into the eastern. The two areas are operated by different mining and dressing complexes. The Bol'she Tokmak district is the largest of the three and apparently contains the major reserves not only of the Nikopol' district in the broad sense (including the Bol'she Tokmak) but of the entire continental world. The manganese ores were accumulated in fold depressions in the crystalline basement.

The oldest rocks in the area are granites, gneisses, and metabasites of Precambrian age and making up the southernmost part of the Ukrainian shield. This shield dips at low angles to the southeast, and its upper surface is composed of an alternation of low ridges and shallow hollows (the ore being in much younger rocks accumulated in the fold depressions mentioned above).

An appreciable portion of basement still is covered by weathered material derived directly from the crystalline rocks of this basement. No rocks between Precambrian and Eocene are known in the district. The first of the Eocene beds, deposited transgressively from southeast to northwest, are the sands, clays, and coals of the Buchaksk stage. The Buchaksk sediments were followed by the clays, marls and siltstones of the Kiev stage (also Eocene) that mainly overlie the Buchaksk but also notably transgress beyond them to lie directly on the crystalline basement.

The next rocks to be formed in the area were those of the Oligocene Kharkov stage, of which the Nikopol' manganese formation is the basal member. The Nikopol' formation is clearly unconformable on the underlying rocks; to the south it rests on members of the Buchaksk and Kiev stages and, to the north, on the crystalline shield itself. The deposition of the Nikopol' formation, like the Eocene beds that preceded it, took place transgressively from southeast to northwest. The basal portion of the Nikopol' is composed of light-colored orthoquartzites and quartz sands, and these are overlain by green-gray-yellow clays (made up mainly of hydrated gaulconites, hydromica-montmorillonite, zeolites, quartz, and glauconite); there are some whitish clays as well. The Nikopol' beds do not cover the entire area of the district, partly because of later erosion but principally because these beds are confined largely to depressions in the Original basement. This follows because the Eocene sediments only partly filled these ancient depressions or, farther north, had never been deposited at all. The Nikopol' beds are, of course, markedly discordant on the rocks on which they rest, be those older rocks basement material or Eocene beds. This lower part of the Khar'kov footwall series ranges in thickness from several centimeters to 25 to 30 m the thickness of these beds increases to the south with the dip of the basement rocks, being especially thick the deeper depressions in these ancient rocks (Varentsov and Rakhmanov, 1977, 1980).

The beds of manganese ore are distributed transgressively at different stratigraphic levels in the lower Khar'kov series just described. Regularly any one of these beds of manganese ore thins out to the northwest and is replaced along strike by deposits of sand and clay; these last are, in many places, coal-bearing and contain appreciable amounts of glauconite. Such transitions are not always easily seen because the Oligocene sequence has been eroded considerably in the ore areas.

To the south, the thickness of the manganese-bearing layer increased but the amount of manganese decreases due to a considerable increase in the proportion of clay to manganese but also to an acute reduction in manganese content as well.

The two main areas in which the manganese ores were deposited and were not later removed by erosion are: (1) that of the Nikopol' deposits proper (West and East), which are located northwest and northeast of that city and (2) the Bol'she-Tokmak deposits some 100 km farther to the ESE; to these are added some smaller ore-bearing areas near the town of Ingulets and some others in the Ingulets-Dneper interfluv. The thickness of the manganese layer is 0.5 to 4.0 m. Except for some rare places where the manganese bed outcrops (or outcropped before mining), the depth of the bed below the surface ranges from 15 to 130 m with the latter depth being reached only in the deposits of Nikopol' proper. These depths explain the universal use of open-pit mining in the entire manganese district.

The manganese ore layer provides a definite marker bed that maintains a definite stratigraphic position within the lower Khar'kov, the base of which is defined normally by a band of glauconitic sand from 20 cm up to 1.5 m thick. The ore layer constitutes a rather uneven interlayer of manganese concretions, nodules, and lenses cemented by clay to silt material. The accumulations of these concretions and lenses within the lower Khar'kov may be, more or less confidently, traced for several 10's to as much as a few

100's of meters in the direction of the strike of the beds. In the manganese-bearing layers, the total content of actual Mn-ore may be as much as 50 per cent by weight but normally averages between 15 and 25 per cent.

The sediments above the ore are separated from that material by a band of iron hydroxides or by considerable enrichment in glauconite. The iron hydroxides lie over oxidized manganese ores and the glauconite over carbonate ones. Above these hydroxides or glauconite bands, the rocks are composed of green-gray massive or laminated clays that are made up largely of montmorillonite with a little glauconite and even less quartz grains. These beds range from 25 cm downward to a few centimeters where fully represented; they may have been much eroded or entirely removed just before the late Oligocene.

Within the manganese layers, three zones have been recognized: (1) the nearest shore and least buried - the oxide zone; (2) the next deeper - the mixed oxide-carbonate zone; and (3) the manganese carbonate zone; it appears that minor amounts of oxide minerals of manganese were deposited before the Mn-carbonates. This view must now be regarded as rather simplistic as the work of Varentsov and Rakhmanov (1980) indicates (see below).

The Nikopol' formations that contain the manganese ores were first developed as a result of intense weathering of the ultrabasites to the north of the present Nikopol' ore bodies. The manganese appears to have been brought into solution as $Mn(HCO_3)_2$, which is much more soluble than $Fe(HCO_3)_2$. Such solution is likely (Borchert, 1980) to have taken place at a pH of about 7 and over a rather wide range of Eh, thereby producing a definite and major separation of the two elements. The removal of Fe from these source rocks would have required an appreciably more acid condition and a more reducing value of Eh. Such a lower Eh could have been produced, Borchert thinks, only by decomposition of organic matter. Such organic matter was apparently lacking in the Ukraine of Oligocene Khar'kov stage time. A contrary suggestion to this method was proposed by Hewitt which he envisioned as being the local introduction of hydrothermal fluids to the sea floor, these fluids being confined to the areas that now contain the Nikopol' managanese deposits.

Under Borchert's hypothesis, the separation of the manganese from the bulk of the silicon and aluminum contained in the original source rocks was accomplished by "humic acid" solutions produced in swamp-forest and/or moor facies. Such iron as does accompany the manganese probably would have been separated largely from the Mn by the conversion of the HCO_3^{-1} complex with iron to colloidally sized particles that were rapidly destabilized and precipitated on contact with saline sea water. This removal of iron permitted the manganese, when conditions became correct for its deposition, to be deposited in a largely iron-free state (the amount of $FeCO_3$ in the Mn-carbonate ore appears to average about 2.5 per cent). The primary, essentially oxide-hydroxide-carbonate ores probably were formed under extreme reducing conditions. In ores formed in an environment in which an appreciably higher Eh obtained, manganite [MnO(OH)] appears to have been the first mineral to have formed with the manganese carbonate later replacing the manganite under higher Eh conditions. In the northern (shoreward) part of the Nikopol' deposits, hypergene (supergene) reactions (taking place in a higher Eh environment than the primary deposition) converted the primary ores to: (1) manganite ores; (2) pyrolusite (MnO_2)-manganite ores; (3) pyrolusite ores; and (4) "polypermanganite" ores (by which Varentsov appears to mean a variety of more highly oxidized manganese ions). Each of these ore types has been divided into various structural varieties.

The only ores that retain their primary character and texture are the carbonate and manganite-carbonate ores. The corrosion of primary manganite pisolites and their replacement by Mn-carbonate were an earlier stage in ore formation than the hypergene transforamtions. Hypergene reactions appear to convert the carbonate and carbonate-manganite ores to manganite-pyrolustite ores, the amount of pyrolusite depending on the degree to which manganite was converted to that mineral. Under hypergene conditions, the terminal and stable products of oxidation of the primary ores are the pyrolusite ores in the central portions of masses of which are relict structures of manganite and carbonate composition.

The polypermanganite ores include manganates, the anhydride of which (according to Varentsov) is manganese dioxide. The principal polypermanganite is, according to Varentsov, the barium-bearing variety of permanganite, that is: $BaMn^{2}Mn^{4}_{8}O_{16}(OH)_{4}$, which Fleischer calls romanechite. Another of the polypermanganite is cryptomelane $[K(Mn^{4},Mn^{2})_{8}O_{16}]$. The ores of this type are, of course, the most highly oxidized of the Nikopol' deposits and normally are rather widely distributed, but essentially no ores consist completely of polypermanganite minerals, but the ores that contain varied amounts of the polypermanganites are much more widely distributed. The polypermanganites were most readily produced in the more uplifted portions of the Nikopol' deposits. In places, however, pyrolusite was formed after the polypermanganite minerals as a result of the dehydration of the latter. In areas of much ground water circulation, however, collomorphic incrustations were formed after pyrolusite and consist of hydrated types of MnO_2 with absorbed cations. The suggestion has been made that the polypermanganite minerals were produced by the absorption of hydrated varieties of MnO_2 from solutions of alkaline, alkaline-earth, and other metals.

Varentsov and Rakhmanov (1980) consider, therefore, that the primary of the South Ukrainian basin took place in two phases: (1) accumulation of hydroxide compounds in a situation of high Eh values and a quite intense addition of ore-forming components from near-bottom waters and (2) transformation of carbonate forms, during diagenesis, to manganese hydroxides, in this event, a medium with low pH was necessary. Later after post-Oligocene tectonic movements, ground water in the more uplifted portions reacted with the primary and diagnetic minerals to produce pyrolusite, polypermanganite minerals, and manganite.

The source rocks from which the manganese in the various deposits of the Nikopol' and other South Ukrainian deposits was obtained were the lower Proterozoic mafic metavolcanics that closely coincide with the northern boundaries of the ore basins. The metavolcanics were topographically higher up and farther north than the present loci of the manganese ores, so drainage into the then-existing basins was readily achieved.

Contrary to what has been presented above, it is possible that the manganese needed in the formation of the Nikopol' (and associated Mn-deposits) may have been derived from volcanic-exhalations extruded on the sea floor. Those who support this concept, despite Borchert's opinion (summarized above), argue that it would not be possible to separate, by Borchert's mechanism, the huge amount of iron set free by weathering so completely from the much smaller quantities of manganese so liberated. Certainly, the amounts of secondary iron minerals present in the general vicinity of the various Nikopol' manganese does not seem of anything like the necessary order of magnitude to account for the iron that must have been weathered-out from the ultramafic rocks. Hewitt (1966) has suggested that the manganese needed was introduced by hydrothermal solutions that poured out onto the sea floor in the Nikopol' basin with the manganese ore beds being thickest in the depressions on the basin floor. Hewitt points out that the bed of manganese was deposited from almost at the strand line seaward for distances measured in hundreds of meters. Near the strand line the ore is composed entirely of such manganese oxides as (in 1966 terminology) pyrolusite, psilomelane, and wad. Further down the slope of the basin, the content of oxides declines and that of rhodochrosite increases until, in the deepest parts, the ore is entirely manganese carbonate. [At the time of Hewett's paper, the presence of oxides of manganese in minor amounts in the deep-seated manganese carbonate was recognized but he says nothing about the relative age of oxides to carbonates.] Hewitt states that, where the Mn-bearing bed is wholly constituted by manganese oxides, the Mn content ranges from 45 to 48 per cent with iron running from 0.5 to 1.0 per cent and that of calcium from 1 to 2 per cent. In the carbonate ore, the Mn content ranges from 10 to 20 per cent, the Fe content from 1 to 8 per cent, and the CaO content from 13 to 34 per cent. Hewitt does not, however, say anything about how the up-slope oxide ores were formed from the carbonate variety. Almost certainly, however, he thinks that the higher content of nascent oxygen in the near-surface waters was responsible for converting the manganese in the volcanic exhalations reaching the sea floor to oxides instead of the carbonates deposited at greater depths. But did this conversion take place during primary deposition of the manganese or was it a supergene (diagenetic) process? Borchert does not answer this question, nor does Hewitt.

Varentsov (1980, p. 33) points out that the manganese oxides resulted from the oxidation (or alteration) of more reduced manganese compounds; thus, he conceives of the manganese carbonate being the major primary mineral and that the oxides were apparently seldom, if ever, deposited as primary minerals but were produced by diagenetic (supergene) processes. Of course, he considers that the Mn-carbonates were deposited as sediments from waters rich in manganese brought in from the surrounding land surface, but his belief that the oxides were produced from the carbonates would apply even if the carbonates were produced from volcanic exhalations.

But why was the initial precipitation of manganese in the Nikopol' basin in the form of manganese oxides? In a general way, of course, this can be explained as saying that the Eh of the exhalative solutions or of the drainage from the adjacent land surface was higher than it later became. But why was this so? This situation is more readily explained if the depositing solutions were of hydrothermal character and had, when they first reached the sea floor, been of a higher temperature and higher Eh and, for a short time, deposited oxides instead of carbonates as is the normal condition in hydrotheral deposits in general. What would change the Eh of surface waters from more oxidizing to less oxidizing and then permit the sea water to become more oxidizing again is less readily explained unless the sea floor rose after the initial deposition of the manganese carbonate, permitting the waters in contact with the near-shore manganese carbonates to be of sufficiently high Eh to convert the carbonates to oxides through supergene (diagenetic) reactions.

Hewitt does not think that the manganese oxides were diagenetically produced but resulted from the entry of the hydrothermal solutions into the near-shore sea water with its high nascent-oxygen content and its concomitant high Eh so that the manganese introduced into the nearer-shore water would have been deposited directly in higher-valence states than farther out into the sea basin. The lack of lower valence-state manganese minerals, such as braunite and hausmannite ($Mn^2Mn^3_2O_4$), farther from shore would have been caused by a greater abundance of carbonate ion in that water volume and an appreciable lower oxygen potential, coupled with the prevailing lower valence-states of the manganese. The small amounts of iron in the deposits appear to be confined to montmorillonite; the lack of siderite strongly suggests that the solutions that brought in the manganese were low in iron as is true of such hydrothermal fluids as deposited the late-stage rhodchrosite at Butte.

In the Nikopol' ores, the ratio of manganese to iron ranges from 8 to more than 20 to 1 with the usual range being between 15 and 20 to 1. Even allowing for the separation of iron from manganese derived from ultramafic rocks as described by Borchert, it is difficult to believe that the separation could have been so complete in weathering and transport processes. The ratio of phosphorus to manganese is vanishingly small, where as the amount of clastic impurities, mainly quartz and claye-glauconitic material, is rather high. The SiO_2 content of the oxide ores rarely drops below 5 to 7 per cent and that of carbonate ores ranges between 30 and 70 per cent.

Although iron in the Nikopol' ores is low, the related minor elements - Ni, Co, Cr, V, Cu, and Mo - appear to have been concentrated above what they would have been in ultramafic rocks; again this suggests that the manganese came from a hydrothermal fluid rather than from a weathered terrane.

Thus, it does not appear to be certain whether the Nikopol' ores were formed by sedimentary process acting on material released from ultramafic rocks by weathering and surface transport or by sedimentation from volcanic exhalations introduced into the sea where the Oligocene rocks sequence was being deposited.

The major argument against the Nikopol' ores having obtained their manganese from volcanic exhalations is that the Oligocene section contains no volcanic formations, which means that, unless Oligocene igneous magmas (for which there is no evidence in the Oligocene strata) can be discovered, a source for volcanic exhalations does not exist. On the contrary, iron-manganese amphibolites (in which iron averages 26 per cent and the manganese less than 0.1 per cent) are present in the basement rocks to the north of the Oligocene sedimentary basement. It is, however, difficult to see (even allowing for the soundness of Borchert's hypothesis for the separation of iron from manganese) how the nearly clean separation of the two ferrous metals could have been so thoroughly accomplished.

Granted that the volcanic-exhalative hypothesis should be found to be sound, the Nikopol' ores should be categorized as Sedimentary-A3, that is, deposits chemically concentrated in quiet waters through the introduction of fluid igneous emanations, plus probable action of circulating ground water (supergene processes - B2).

If the weathering-surface transport-sea water deposition-diagenetic modification hypothesis is correct, the deposits should be categorized as Sedimentary-A1, 2, 4, that is inorganic and organic sedimentary reactions, plus diagenesis, plus probably the action of circulating ground water - B2.

At this stage, it seems impossible to accept without reservation, either of these two hypothesis, and further study of the problem remains to be done.

NIKITOVSK, UKRAINE SSR

Early Mesozoic (?) Mercury Epithermal

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Notes

The mercury deposit of Nikitovsk is located in the Donets basin, on the east side of the Dneper river at about 48°20'N,38°03'E; the workings lie about 160 km ENE of Khar'kov and an equal distance SSW of Voronezh.

The deposit is localized by structures and favorable beds in the Gorlovo (Gorlovskaya) anticline; this anticline lies on the flank of the main Donets basin anticline. The deposits is confined to two sandstone beds, separated by a bed of middle Carboniferous shales. The ore in both beds is related to the crest (or crests) of the Gorlovo anticline and may extend down the dip of either limb which usually dips at somewhat more than 45°; the lower ore body is in much the same position in a sandstone bed immediately beneath the shales that separate it from the upper sandstone. Both ore bodies are cut by the Nikitovsk (Sekushchaya) thrust fault, and both ore bodies are displaced by this structure with the ore on the crest having been moved upward in relation to that on the limb as would be expected of a thrust fault.

The stratigraphic sequence consists, in addition to the interstratified sandstones and shale, of intercalated layers of limestone and anthracite. The folds in this sequence, folds that involve the ore beds, have a NW strike. The axis of the Gorlovo anticline has, in turn, been folded into sinuous curves that form brachyanticlinal ridges or domes. There are four of these and they are named the Chernokurgansk, Katushka, Sofia, and Chegarnick; the Nikitovsk bodies are confined to the last two of these domes. Other mercury deposits in the area include the Zheleznyansk, Chernokurgansk, and Novozavodsk; the last of these three deposits is connected with the same Sofia and Chegarnick domes as is the Nikitovsk.

The district has been cut by numerous faults of varied structural magnitude: major ones are steeply dipping overthrusts of more or less NW strike and are known as the Nikitovsk, Artemovo, Chegarnik, and Southern. In addition, the district is cut by second-order diagonal faults.

The most important of these overthrusts is the Nikitovsk, which has produced the Sekushchaya discordant zone, that has guided upward the ore fluids in all of the ore fields of the district. This zone is 5 to 20 m thick and is associated with diagonal faults of both NW and NE strike. In addition, interbed slip joints have resulted in shear zones and had major effect of the localization of the mineralization. Not only has the Sekushchaya discordant (or shear) zone served as a channelway for ore fluids but also acted to enclose the ore within that structure. Further, to aid in ore localization, the sandstone layers, even initially quite porous, were made more so by pre-ore crushing during the folding, crushing that particularly affected the sandstones on the crests of the folds and on the limbs immediately below the crests. The final control in keeping the ores in the sandstone beds was the presence of quite impervious shales above the sandstones in which the ore was deposited; these prevented the easy rise of such solutions to higher levels. Some shale interbeds in the sandstone also acted to localize the deposition of ore. Most importantly, the ores are where they are because the associated folds and fractures directed the ore fluids into the anticlinal crests where the open-space created by these structural features greatly aided in causing ore deposition.

Kuznetsov (1977) reports that two types of ore have been recognized on the basis of morphology of the bodies: (1) layered ore segregations that are confined to junctions between the Sekushchaya crushed zone and the sandstones that were so readily invaded by the ore fluids once they reached them because of the drastic jointing of the sandstones in the anticlinal arch, (2) vein-like bodies that are enclosed in transverse and diagonal joints of steep dip. In shape, the ore bodies have been separated into (a) simple and complex veins, (b) lenses, (c) stockworks, (d) layer-like segregations and (e) nests. The ore fabrics are categorized as (1) segregated, (2) fine-veinlet, and less commonly (3) massive, (4) brecciated, and (5) breccia-like.

The Nikitovsk deposits are of the quartz-dickite-cinnabar type and form in such rocks as siltstones, sandstones, shales, and conglomerates. Typically these deposits occur in terrigenous geosynclines, marginal foredeeps, and aulacogens. The mineralization of deposits of this type occur in the late stage of orogenesis; they have no direct, visible association with magmatic phenomena but certainly were deposited from hot-water solutions. These deposits are telethermal or epithermal, depending on the rapidity with which the responsible ore fluids lost pressure and temperature. The major ore mineral is, of course, cinnabar; minor quantities of realgar and orpiment are present and some arsenopyrite may be an early mineral. Some sulfosalts, in minor quantities, may be present, but they are of no commercial importance. The principal gangue mineral is quartz, but dickite is quite common and is characteristic of this type. Some alteration of the host rocks has produced quartz and argillite minerals. Kuznetsov includes the famous deposit of Almaden in this type as well as numerous deposits in the Soviet Union; these are others in the Donets basin, in the northwestern and northern Caucasus, in the Kuznets Alatau, in the Gornyi Altai, in east Sayan, in Chukotka, and in the Koryak plateau.

In the Nikitovsk deposits, the principal (and only ore) mineral is cinnabar. Stibnite, arsenopyrite, pyrite, and marcasite are rather common; bismuthinite, chalcopyrite, galena, and sphalerite are rare and are in isolated microscopic segregations. In addition to quartz and dickite, the gangue minerals include sericite and iron carbonates; unusual minerals are found in the oxidized zone, such as scorodite ($Fe^3AsO_4 \cdot 2H_2O$), antimony ochres, melanterite, and epsomite.

Five principal phases (perhaps several more) have been recognized in the Nikitovsk ores; these are: (1) arsenopyrite, (2) quartz-arsenopyrite, (3) stibnite-carbonate, (4) stibnite-cinnabar, and (5) dickite-stibnite-cinnabar.

Bol'shakov (1975) says that there are four to six stages of mineralization in the Nikitovsk mercury deposit without any clear indications that the sequence was interrupted by any major changes in ore textures or compositions of the minerals being deposited. He suggests that the mineralization was one long-continued process that can be divided arbitrarily into stages of mineralization with changes in the minerals being deposited but with no evidence of any break in the reactions taking place. The beginning of mineralization was the deposition of quartz in large quantities including quartzification of the immediate wall rock, then arsenopyrite, followed by dickite and pyrite and more arsenopyrite.

In the course of the various replacement reactions that occurred during the mineralization, Bol'shakov believes that Al and alkalies were added as were Fe, Mg, and Hg. The sandstones were altered far more than either the limestones or shales.

The cinnabar was deposited by replacement of wall rock or of fragments in the crush zone of the main fault. Massive bodies of Cinnabar contain grains of zircon that were originally in the wall rocks. Under the impermeable overlying shales, metacrysts of cinnabar developed by replacement of the quartz grains of the sandstone. These Hg_2S crystals are 3 to 6 mm in diameter and replaced as much as 20 per cent of the sandstone.

In many places, cinnabar rims large pyrite grains; these rims are of the same thickness no matter what the size or form of the pyrite.

Where shale is brecciated near the border of shale with sandstone, Hg_2S crystals fill open spaces in the breccia; microscopically, the Hg_2S crystals are seen to be full of shale inclusions. On the contrary, stibnite shows no signs of having effected replacement.

Information from the homogenization of gas-fluid inclusions gives formation temperatures as follows: quartz-240° to 290°C and cinnabar and stibnite 120° to 150°.

Soviet geologists recognize that the ore fluids may have been of magmatic origin but that they also may have come from extra-magmatic activity in which they include waters of metamorphogenic origin. The best-founded hypothesis (according to Soviet ore geologists) is that the ore fluids were released from deep-seated subcrustal foci where basaltoid or alkaline-basaltoid magmas gave off the requisite ore fluids. Their opinion related these ore fluids genetically to an early Triassic andesite-trachyandesite dike complex, the age of which is 200 to 230 m.y. The rocks of this complex have significantly higher Hg values than clarke (clarke being about 0.9 ppm). These ages agree with that of 200 m.y. (by the K/Ar method) obtained from pre-ore Nikitovsk sericite.

Thus, the Nikitovsk deposits are here assigned an early Mesozoic age. I believe that the deposits were formed near enough to the then obtaining surface to be better classified as epithermal and not telethermal, but they must belong to one or the other category.

SOLNECHNOE-FESTIVAL'NOE, AMUR OBLAST

Late Mesozoic	Tin, Tungsten Copper	Xenothermal to Kryptothermal

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Notes

The Solnechnoe and Festival'noe tin-tungsten deposits are located in the Komsomol'sk (Myao-Chaya) ore region. Center of the area is the city of Komsomol'sk-na-Amure (ca50°32'N,136°59'E), and the two mines appear to be in its immediate vicinity. The city is about 265 km NNE of Khabarovsk.

The ore area is concentrated at the intersection of several regional faults that are contained in a superimposed Cretaceous downwarp that later was uplifted. The lower portion of the downwarp is composed of Jurassic terrigenous sediments with a minor development of diabases with which are cherty and chert-clay rocks that apparently have no genetic connection with the diabases. At the time that the area was deformed into geosyncline, local sill-like intrusions of diabase were introduced. This deformation produced NE-striking linear folds; a deep-seated thrust fault in the ore-bearing area converted these folds to a more or less east-west strike.

The upper portion of the downwarp is markedly unconformable on the eroded Jurassic folds and consists of Cretaceous volcanogenic-sedimentary formations that have been slightly deformed. The lower part of this Cretaceous sequence is made up of discontinuous conglomerates, silicic extrusives and related tuffs that are as much as 700 m thick (the Kholdam group). Above these are andesites, dacites and related pyroclastics, locally over 1200 m thick, (the Amut group). These Cretaceous rocks are cut by granitoid intrusions of quite varied composition, the age of which ranges from 73 to 113 m.y. these ages average 85 to 93 m.y., making them late Cretaceous. The outcrops of these igneous masses (and of the ore bodies associated with them) have been covered by Miocene and lower Quaternary mafic extrusives. Prior to these extrusions, tin placers were formed on the weathered surface and were preserved where covered by the extrusives.

The intrusive-extrusive complex of the Komsomol'sk area became increasingly mafic as time passed, and the igneous rocks exhibit appreciable evidence of assimilation of one rock type by another. The relations of the various igneous rocks to each other are so complex that relative ages are difficult, if not impossible, to determine. One formation can definitely be dated, a diorite porphyrite that cuts all the other intrusives. The presence of a large body (90 x 40 km) of sedimentary rocks, now converted to hornfels suggests that it lies over a large intrusive massif that may be a single or a multiple intrusion. Nor is there evidence that the igneous bodies that outcrop at the surface are the result of the differentiation of a single igneous body (Materikov, 1977).

The largest intrusive massif exposed at the surface is mainly granite and is about 200 km^2 in area; it is situated at the west of the region and is elongated in general conformity with the strike of the folds. In association with these granites (certainly spatial and probably genetic) are greisens that contain weak tin mineralization and quartz-feldspar and quartz-tourmaline veins that are located in NW-striking joints.

All of the commercial tin deposits in the area are associated with small stock-like bodies of granitoids in the central part of the area that become more mafic as the age of the intrusions decreases. These igneous bodies were localized by more or less east-west-striking structures that are transverse to the folds but with apophyses that follow north-south of NW-SE joints. The dikes that fill these structures contain the mineralized tin zones.

These tin zones are in the portions of the apophyses that have been most crushed and jointed and have steep dips; they can be followed along strike for many kilometers, have widths that may be 100 m or more, and have been followed downward for, in some instances, over 1000 m. Radkevich and her colleagues have determined that the ores were deposited as near to the then-existing surface as 200 to 300 m and with their roots reaching 1.5 to 2 km below that surface.

In the mineralized zones, quartz-tourmaline replacement masses are centrally located; these are surrounded by aureoles of quartz and sericite (in quartz-rich rocks) and by propylitization (in intermediate and mafic rocks.

In a few places, the quartz-tourmaline cores grade upward into quartz-chlorite varieties that contain sulfides.

Irregularly along the mineralized zones, the quartz-tourmaline bodies have been cross-cut by more or less N-S quartz veins and veinlets that contain cassiterite and also by quartz-pyrite veins containing pyrrhotite and chalcopyrite. These veins were followed by quartz-carbonate and carbonate veins that include galena, sphalerite, and lead-antimony sulfosalts. These veins are confined to the quartz-tourmaline bodies and only where these are located in zones of propylitization in the upper Cretaceous andesites.

Tin mineralization poor in sulfides predominates where the host rocks of the ores are Jurassic sediments; as the mineralization passes upward into the volcanic rocks, copper-tin or tin-lead-zinc types take over but change to an entirely tin-free mineralization in the highest rocks of the Cretaceous sequence. Locally, however, where the granitoids occupy east-west zones, the polymetallic mineralization may go down to the lowermost depths reached by the ores. In the far north of the ore-bearing area, where granitoid outcrops are lacking, the tin mineralization is changed to a mercury-antimony type.

Two important (probably the most important) deposits in the Komsomol'sk area are those of Solnechnoe and Festival'noe. Although exact information is not available (to me) as to the locations of these two relative to each other, they cannot be far apart, particularly as both are near the unconformity surface between the lower Jurassic and lower Cretaceous rocks of the Komsomol'sk ore region (Materikov, 1977).

The Solchenoe tin deposit is confined to a major zone of more or less north-south faults and fractures, close to the point where these fractures intersect with an approximately east-west line of granitoid intrusions. This deposit is farther beneath the lower Jurassic - lower Cretaceous contact than is that of Festival'noe. The deposit was exposed by the erosional development of a deep river valley, and considerable amounts of the original deposit appear to have been lost by this process. The deposit proper extends along strike for about 8 km, and its thickness ranges from a few to 115 m; the depth of mineralization reaches several hundreds of meters. Not all of the deposit is commercially mineralized, blocks of weakly altered or mineralized rocks separating those of ore grade. Thus, the deposit is composed of many discrete ore bodies, some of which have been completely or partially covered by post-ore basalts. Where the ore bodies have been exposed at the surface, either because the covering basalts were eroded or never covered the body in question, oxidation from the surface has penetrated as far down as 100 to 110 m.

Where the mineralization is contained in lower Jurassic sandstones and siltstones, the ore bodies were emplaced mainly by metasomatism and are quite thick. The mineralized rock volumes generally split into two or more distinct zones but retain a uniform lithological composition; in the intrusive granodiorites and quartz monzonites, however, and in the surrounding hornfelses, the mineralized zone divides into many thin branches, complicating their form and structure and appreciably increasing the total width of the mineralized belt.

In the mineralized zones, the cassiterite is contained in quartz veins and veinlets; these quartz bodies are made up of fine-grained masses, frequently zoned, of elongate-prismatic crystals that form drusy and comb-like aggregates. Fluid-inclusion study indicates that the quartz crystallized between 382° and 320°C. In addition to the quartz and cassiterite, the veins contain local segregations of wolframite, arsenopyrite, and scheelite in amounts minor relative to the cassiterite. Near the borders of the veins, a nearly colorless, acicular variety of tourmaline is present. Materikov (1977) thinks that this troumaline has been remobilized from primary examples of the mineral.

The cassiterite grains that are from 0.01 to 2 to 3 mm normally are accumulated in aggregates. In drusy quartz, nests and cavities contain cassiterite that is up to 5 mm in length. The tin mineral normally is brownish-red and locally is zoned. As the ore is followed downward and from north to south, the crystal type of cassiterite changes from acicular to essentially equant. In the same directions, indium and scandium in SnO_2

decrease and tungsten goes up. Cassiterite in this deposit is low in trace elements in contrast with this mineral in other deposits in the Komsomol'sk region except for tantalum that may be as much as 0.006 per cent.

In most of the deposits of this region, the ore, where followed upward, grades into quartz-carbonate-sulfide veins and veinlets that contain mainly pyrrhotite. This is not true, in general, for the Solnechnoe deposit. Such of the veins as occur at Solnechnoe are on the southern border of the Main ore zone where the sedimentary and granitoid rocks are in contact. This means that Pb, Zn, Cu, and Ag are, in contrast to the other Komsomol'sk ore bodies, are insignificant in amount in the Solnechnoe ores.

The Festival'noe deposit, although in the same lower Jurassic rocks as that of Solnechnoe, is near to the surface of the structural unconformity between the Jurassic and Cretaceous rocks of the district. In fact, the principal ore zones at Festival'noe extend upward from the Jurassic sandstones and siltstones into a lower Cretaceous sedimentary-volcano sequence of interbedded conglomerates, tuff-conglomerates, and quartz-porphyry tuffs. On the southern margin of the deposit, some of the mineralization was buried under upper Cretaceous extrusives. Again, in contrast to Solnechnoe, where the ores are followed upward, the strictly tin (or tin-tungsten) ores grade into copper-zinc bodies.

The forms of the Festival'noe ore bodies is quite complex. Their overall shape is that of steeply dipping, linear deposits, but they are complicated, in both plan and section, by bends, swells, pinches, and numerous apophyses. The thickest portions of the ore bodies are located along the contact between the lower Cretaceous and lower Jurassic beds. In the Jurassic rocks, the thicknesses of the ore bodies becomes less, and the ore zones divide into subparallel branches and thin appreciably. In the swells and pinches in the Cretaceous beds, the former normally are confined to the horizons of tuff-conglomerates; in the more compact and uniform quartz-porphyry tuffs, the ore zones are definitely thinner; this thinning may be quite abrupt. Where the mineralized zones are near the east-west fractures, there morphology is quite complex.

The lower Cretaceous rocks contain both steeply and gently dipping veins with the latter generally linking up with the former, particularly along the dips. These veins differ considerably in thickness, the variation being between a few centimeters and a few meters, but the extent of these veins along strike is not more than 50 to 70 m. Steeply and gently dipping veins do not differ appreciably in composition, being mainly quartz with cassiterite and arsenopyrite; locally quartz and tourmaline bands alternate.

In the lower portions of the Festival'noe veins (those in Jurassic rocks), the mineral composition is much the same as such veins in the Solnechnoe deposit. The replacement of the host rocks in Festival'noe by quartz and sericite is much less intense than this process is at Solnechnoe. In the quartz-cassiterite veins at Festival'noe, arsenopyrite is common, but wolframite and scheelite are rare. The development of later-stage sulfides, such as chalcopyrite, pyrrhotite, stannite, (rare) galena, and sphalerite is weak.

The cassiterite in Jurassic rocks usually is located in spaces between the quartz crystals and may be present as overgrowths on the quartz. This cassiterite is relatively late, but there also is an earlier type that is concentrated around xenoliths in the quartz veins and in veinlets in the quartz-tourmaline replacement bodies.

Above the structural unconformity, the character of mineralized zone changes; the thickness of the quartz-tourmaline replacement masses lessens until they finally thin out completely in the uppermost Cretaceous horizons. As the quartz-tourmaline bodies are followed upward, cassiterite-sulfide-bearing quartz veins become more abundant, replacing the quartz-tourmaline material. Further, the sulfides increase, in particular the chalcopyrite.

These sulfides are associated with cassiterite in both these quartz veins and in quartz-tourmaline replacement bodies; in both places, the sulfides form large nests, thick branching veins, and complex-shaped lenses as well as dense networks of thin veinlets and as fine segregations. The major sulfides are chalcopyrite, arsenopyrite, galena, and stannite with lesser pyrrhotite, pyrite, and marcasite. Microscopic amounts of tetrahedrite-tennantite, jamesonite, boulangerite, bismuthinite, native bismuth, valleriite, cobaltite, and franckeite have been observed.

In the Festival'noe quartz veins in the horizons above the unconformity, the early quartz of the quartz-cassiterite-arsenopyrite mineralization was quite coarse. These early quartz crystals changed considerably from weakly elongate to equant forms in the lower horizons (above the unconformity) to markedly elongate and acicular types in the upper horizons. The deposition of the cassiterite in the lower horizons took place between 317° and 290°C and the quartz associated with it between 360° and 291°C. In the middle and upper horizons, the cassiterite was precipitated between 294° and 285°C. The later quartz-cassiterite-sulfide association was determined to have been deposited between 268° and 230°C (Materikov, 1977).

The cassiterite closely related in time to the sulfides was much more abundant than the earlier (and coarser) cassiterite. This cassiterite took the form of veinlets and nests in the sulphides (arsenopyrite, chalcopyrite, and sphalerite); in places, however, the cassiterite is cut by veinlets of chalcopyrite. The early cassiterite in essentially all places is surrounded by reaction rims of stannite; this is not true of the later SnO_2 around which the stannite rims are uncommon, but, where stannite occurs with the late cassiterite, it most usually appears along contacts between sulfides and quartz and carbonates. In a few places, franckeite was found in aggregates of chalcopyrite, and, in the franckeite, were enclosed small grains of SnO_2. Gangue minerals include chlorite (thuringite), siderite, and fluorite; these normally are in quartz veins with abundant sulfides.

In the Festival'noe deposits, in contrast to those of Solnechnoe, the SnO_2 is much more enriched in In, W, and Fe but it is lower in Sc. The sulfide minerals at Festival'noe contain larger amounts of trace elements than do those at Solnechnoe. Near the surface of the unconformity between Jurassic and Cretaceous rocks, the trace elements in the Festival'noe ores depart appreciably from the normal averages with the usual case being for these elements to be higher than the average for the entire deposit.

The igneous rocks with which the ores are spatially, and probably genetically, associated quite definitely were introduced in the late Mesozoic, almost certainly dating the ore cassiterite-(tungsten)-sulfide ores as of that same time.

The association of cassiterite with arsenopyrite, wolframite, and scheelite and the temperatures of deposition determined from fluid inclusion studies set the cassiterite stages of the two deposit (Solnechnoe and Festival'noe as either hypothermal in non-calcareous rocks or xenothermal. The only question not clearly solved is which of these two categories is the correct one. The abundance of extrusive and near-surface igneous rocks would seem to point toward xentothermal as the correct category to which to assign these deposits. The depths to which mineralization extends appears to be from a few hundred to a 1000 to 1500 m seems a depth much more compatible with the xenothermal category than the hypothermal one. The probable near-surface environment in which the ores were deposited also appears to conform to the requirements of the xenothermal subdivision of the modified Lindgren classification. The SnO_2 deposits, therefore, are categorized as xenothermal.

The temperatures at which the quartz containing the sulfides of the later stage of mineralization was emplaced ranges between about 270° and 230°C, a range that places these sulfides in the kryptothermal range, and they are so classified here.

Although the Komsomol'sk ore district was early known as a tin district, it is now regarded as a complex ore region as most of the ores are now known to contain lead, zinc, silver, bismuth, cobalt, cadmium, and indium. The complexity of the mineralization of the Festival'noe deposit resembles, on a small scale, the Cornish deposits. Much of the Festival'noe ore is far from granite outcrops, but the mineralization descends to considerable depths with vertical zoning being quite obvious.

The ores are quite fine grained with cassiterite and chalcopyrite being quite closely associated. Kokorin and others (1979) assume that the ore at the lowest levels of mineralization will be in stockwork bodies in which the major recoverable elements will be molybdenum, gold, and bismuth (?).

These authors believe that mineralization at Festival'noe is controlled by a north-south striking fault in strongly folded Jurassic shales over which Cretaceous volcanic rocks have been thrust. The wall rocks have undergone appreciable alteration, mainly the development of epidote, plus chlorite and carbonate. The main minerals in the deposit are: cassiterite, arsenopyrite, chalcopyrite, wolframite, scheelite, pyrrhotite, magnetite, and pyrite. Subordinate minerals are: sphalerite, galena, stannite, bismuthinite, native bismuth, boulangerite, marcasite, and loellingite. Rare minerals are: bournonite, cobaltite, stibnite, franckeite, teallite, matildite, native silver, and native gold.

At depth, the ore zone divides into several small branches with the wall rocks strongly tourmalinized; quartz and tourmaline are the other main gangue minerals with which are associate chlorite, calcite, and epidote. This mineral suite was brecciated and further tourmalinization took place. The third stage was the quartz-cassiterite stage with wolframite, scheelite, arsenopyrite, and tourmaline in large veins.

DZHIDA, BURYAT ASSR

Middle Mesozoic Recent (placers)	Tungsten, Molybdenum	Hypothermal-1 Placers, 1B, 1C

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Notes

The Dzhida ore fields includes a variety of tungsten and molybdenum deposits that appear all to be genetically as well as spatially related. The center of these deposits is at 50°20'N,103°20'E, a spot almost on the border between the USSR and Outer Mongolia. The mines appear to lie just to the east of the city of Zakamensk and to be nearly 600 km SSW of Irkutsk on a western-extending arm near the southern end of Lake Baikal. The mining area appears to be on the south side of the Dzhida river near its source and on the northern slope of the Dzhidinskii mountain range. A small southward extending loop of the border puts the mines in the RSFSR and not in Mongolia.

The mines are located largely in the igneous rocks of a rather complex group of intrusions: leucogranites and granite prophyries of the Pervomaisk and Gorka massifs, the Inkura stock, assorted dikes of bostonite, granite porphyries, quartz pegmatites, aplite porphyries, syenite, quartz syenite porphyries, and micrograntie. These, and other various igneous and metamorphic rocks have been divided into four complexes that range in age from lower Paleozoic to Triassic-lower Jurassic; the youngest igneous rock is 140 m.y. old (or uppermost Jurassic).

The first of these complexes is the Tsakir that consists of ultramafic rocks (serpentinites, pyroxenites, and amphibolites). These rocks cut a sedimentary-volcanic sequence (the Khokhyurtovka group), and a combined igneous and sedimenatry-volcano complex was been folded during the Caledonian orogeny. Later, this complex was cut by Paleozoic quartz diorites. The second complex is the Dzhida of late Paleozoic age that is made up of masses of quartz diorites, quartz gabbro-diorites, microdiorites, and related veins, plus dikes of leuco-granites, plagiogranites, microdiorites, diorite porphyries, spessartites, and quartz veins. The third complex is the Malo-Kunalei that is composed of subalkaline syenites, quartz syenites, and dikes of syenite porphyries, quartz syenite porphyries, and bostonites. The fourth complex is the Gudzhir that is Triassic to upper Jurassic and consists of leucogranites and granite porphyries (the Gorka and Pervomaisk massifs). These rocks are positioned at the intersection of the zone of a NW-SE fault with a contact zone of quartz diorites with sedimentary-volcanic rocks of complex (1) - the Khokhyurtovka. Into this melange also are included dikes that have been separated into: (2) pre-ore rocks (granite porphyries, quartz pegmatites, and aplite porphyries); (b) syn-ore dikes of aplite porphyries (containing segregations of molybdenite) and inter-mineralization dikes (syenites, quartz-syenite porphyries, granite porphyries, and microgranites). The molybdenum and tungsten mineralizations are thought to be genetically co-joined with the Pervomaisk granite porphyries, these last porphyries are the ones that are 140 m.y. old (uppermost Jurassic).

To sort these igneous bodies out on a more ore-related basis, the key to the problem is Inkura tungsten stock that has the form of an irregular letter "C" with the open part of the "C" facing east. In the northwestern third of this opening is the Pervomaisk molybdenum deposit. The rest of the open space in the "C" is filled by the remaining (and less mineralized) portion of the Pervomaisk stock. Running irregularly north-south just to the east of the "C" of igneous rocks just described is a dike-like mass of early Cambrian altered ultramafic intrusives; a northern part of this dike has been intruded by granite porphyries of the Gorka stock (in which there probably is some tungsten mineralization. Numerous tungsten-containing veins (the Kholtoson quartz-sulfide-hübnerite) are superimposed on most of the rocks; the veins trend roughly east-west and provide the farthest east and farthest west mineralization in this mixture of complexes; the introduction of these veins appears to have been the last event (except for recent erosion) to have affected the Dzhida ore field.

Polivaitis (1975) and Ignatovich (1975) appear to agree that the sequence of mineralization in the Dzhida field occurred in five stages: (1) pre-ore, alkaline metasomatism and the introduction of quartz-microcline and aplite dikes; (2) molybdenum ore introduction, quartz-molybdenite veins in aplite porphyries; (3) non-ore stage, quartz-pyrite and microcline-quartz veins related to the gray syenite porphyries and the quartz-syenite porphyries; (4) quartz-tungsten mineralization - hübnerite-quartz veins with which are associated pegmatoid-microcline and microcline, plus the Gorka granite porphyries; and (5) tungsten-sulfide mineralization in quartz-sulfide-hübnerite veins associated with cherty quartz.

The Inkura stockwork tungsten deposit makes up the irregularly shaped "C" mentioned above. It contains numerous quartz-sulfide-hünerite veins of various sizes, but there are at least as many major W veins outside the Inkura stock as there are in it. Many, probably most, of the major W veins outside the Inkura stock are located in a major fault south of the stock and extending west-northwest and collectively are known as the Kholtoson tungsten deposit. Representatives of this vein-type also are found in the Pervomaisk Mo stock, both portions of it. The more mineralized part of the Pervomaisk stock (to the west) is a typical stockwork, made up of a steeply dipping network of thin, randomly oriented veinlets: containing quartz, quartz-feldspar, quartz-muscovite, feldspar, and quartz sulfide. Essentially all of these contain hübnerite and scheelite and are confined to zones of intense crushing of the quartz diorites, the bostonites, and the schists. Of these veinlets, about 50 per cent are less than 0.5 cm which, 40 per cent are from 0.5 to 5 cm, and 10 per cent are more than 5 cm wide.

The character of the wallrock appears to exert much, if not complete control on the mineral content of the veins in the Inkura stockwork. Where the enclosing rocks are schists and hornfelses, the veins and veinlets are filled with feldspar and carbonate; where the containing rocks granite porphyry dikes, the veins are filled with quartz and hübnerite and do not contain feldspar.

The veins, in relation to the massif, are either radial or concentric. Commercial tungsten mineralization has been developed in thick bands of complex form and separated from each other by barren or non-commercial vein filling. In the center of the stockwork, most of the veins strike more or less east-west and dip 70° to 85°S. These veins are contained in a band 400 to 500 m wide where they are less clearly outlined than in the northern and southern parts of the stockwork and are at a much greater depth below the surface.

The best-grade ores in the northern and southern sectors of the Inkura stockwork are in bands that are branches of a single stockwork segregation, the form of which depends on ore-controlling structures that strike north-south or NW-SE.

The ores show little change in grade in plan or in section; mineable ore has been encountered in drilling at depths of 450 to 600 m, and the veins are of the same grade and thickness even at these depths. At depth, however, hübnerite has been replaced by scheelite, this latter mineral providing 18 per cent of the WO_3 in the ore.

Recent placers rest on the surface of the Inkura stockwork; they contain commercial amounts of hübnerite over considerable areas.

The Kholtoson deposit of tungsten veins is situated in the southwestern portion of the Dzhida ore area; it is made up of numerous quartz-sulfide-hübnerite veins enclosed in quartz-diorite rocks. Additional individual veins are present in the eastern part of the Pervomaisk granite porphyries and schists. The veins are localized principally by a NW-SE-trending fault; this fault contains such veins for 2.5 km. Further veins occur in the footwall of this fault, where they are emplaced in feather-like offshoots of strike-slip faults. Most of these veins are at angles of 15° to 25° from the strike of the main fault; rarely this angle of departure may be as much as 90°. In several places, such off-shoot veins may provide connections between larger veins.

In the northwestern and southeastern portions of the main fault, more or less east-west-striking veins are separated from each other by a largely barren block of rocks bounded by the east-west fault and by NE-SW-striking veins. Within this block, the channels for the passage of the ore fluids must have provided far less ready access than was true of the more broken rocks on either side of the main fault zone. Materikov suggests (1978) that this condition results from stresses in the center section being relieved by NE-striking vein-joints whereas, to the north and south of this central block, stress relief was provided by the more or less east-west faults, producing structures that gave the ore fluids ready access to these north and south border areas.

Shcheglov reports that there are 140 veins of the Kholtoson type in that deposit, and 70 of these are workable commercially. These quartz-hübnerite veins occupy east-west faults, and their usual dips are 30° to 50°S; locally these dips may be as much as 55° to 85°S. NW-SE and NE-SW veins with steep dips to the northeast and southeast, respectively, are uncommon as sources of mineable ore.

Most of the vein-joints are the results of shearing, but steeply dipping veins in the granite porphyries and the apophyses of the main veins are filled tension fractures; this last type of vein is minor both in strike and dip length. On entering gray syenite dikes, the veins split into thin veinlets. Along dip, the tension veins have twice the length they have along strike. Joint veins appear in many places to be limited to dike contacts.

During tectonic movement, bends in joint-type veins became loci for open space formation and, alternatively, for crushing. Most of the ore material in these veins was deposited by filling of open space, but a lesser amount was emplaced by replacement of the wall rocks and of the earlier feldspar veins; this results in these veins having quite complex forms. Although the average thickness of the veins is about 0.8 m, tectonically induced bulges may have thicknesses from 2 to 12 m; these bulges, of course, alternate with thinned vein volumes where the vein material may be no more than a few centimeters in width. Normally, these thinned volumes are not commercially mineable, in many places being composed of nothing more than hydrothermally altered crushed rock and breccia mineralized with only fluorite and sericite. What tungsten is in these crushed or breccia zones is as scheelite.

The volume of these pinched-down portions of the veins makes up from 18 to 51 per cent of the total vein volume. The richest tungsten in the veins is concentrated in the 200 m directly beneath the surface; below that elevation, tungsten gradually decreases in quantity, and, in the deepest ore levels, tungsten is not half what it is in the upper portions of the veins.

In the Kholtoson veins, there are six stages of mineralization: (1) pegmatoid-microcline, (2) quartz-hübnerite, (3) microcline, (4) quartz-sulfide-hübnerite, (5) quartz-rhodochrosite-hübnerite, and (6) cherty, quartz. The bulk of the tungsten ore was deposited in stages (4) and (5), the amount of tungsten contributed in stage (2) was minor.

In detail, the pegmatoid-microcline phase consisted of uncommon feldspar veins that were, nevertheless, as much as 3 m wide. The stage (2) quartz-hübnerite phase was also of minor extent, the veins being made up of coarse gray quartz that enclosed fluorite, gilbertite, and sericite with insignificant pyrite and hübnerite. Stage (3) was another minor development, in this case thin veins of microcline that included small quantities of plagioclase, fluorite, and sericite; where microcline was replaced by quartz, minor hübnerite was also introduced. Stage (4) is the broadly distributed quartz-sulfide-hübnerite phase; this phase is present in all of the commercially valuable concentrations of hübnerite. These veins are abundant and may contain impressive amounts of tetrahedrite-tennantite, plus appreciable galena, chalcopyrite, and sphalerite; the gangue minerals include microcline, gilbertite, sericite, and fluorite in major quantities and less triplite and ankerite. Additional ore minerals include some pyrite, scheelite, galena, and aikinite. The hübnerite has from 75.24 to 76.63 per cent WO_3, 19.67 to 23.65 per cent MnO, and 0.14 to 2.6 per cent Fe_2O_3. The amount of replacement of hübnerite by scheelite is minimal. Minor tungstite [$WO_3 \cdot H_2O$] has been reported. Stage (5), for the first time, contains rhodochrosite in addition to quartz and hübnerite; sulfides and fluorite are more common than in stage (4). Comb quartz is the common variety of SiO_2 and it contains the gangue minerals. In addition to the hübnerite already mentioned, this stage contains galena, pyrite, sphalerite, chalcopyrite, tetrahedrite, and rare scheelite, aikinite, hematite, and chalcocite.

The Kholtoson veins are given a banded structure by the alternation of rhodochrosite, sulfides, and quartz with hübnerite being present in both the rhodochrosite and the quartz. The veins cut those of the earlier stages.

The cherty quartz assigned to stage (6) is rare in the Kholtoson veins, being largely confined to the Pervomaisk molybdenum deposit in which, however, are major examples of the Kholtoson veins. The cherty quartz also can contain, in minor amounts, hübnerite, pyrite, chalcopyrite, and sericite.

The wall-rock alteration in these Dzhida deposits has converted the igneous wall rocks to beresites (rocks originally prominent feldspar with intergrown quartz and usually much kaolinized), sericite-quartz, gilberite, gilbertite-fluorite, and (in the granite porphyries) greisen. These alterations are essentially contemporaneous with the various stages of mineralization and have thicknesses from a few centimeters to as much as 0.5 m. The contacts of the alteration with the various veins are sharp.

The Pervomaisk molybdenum stockwork is a small, centrally located body of granite and aplite porphyries and granites that was mineralized early [in stage (2)]. It appears to have been mineralized before essentially any tungsten had been introduced into the rocks of the Dzhida region, being cut by Kholtoson tungsten veins and surrounded on three sides by the Inkura tungsten stockwork and on the fourth side by relatively unmineralized igneous rocks of the same type as those constituting the Pervomaisk Mo stockwork. The quartz-molybdenite veins appear to be typical of those in such deposits as Climax, being steeply dipping, thin randomly oriented veins, the filling of which is mostly quartz plus molybdenite, but feldspar and muscovite, also may be present.

So far as I can discover, the sulfides that are so conspicuous in the various Dzhida region ores are not recovered during mineral dressing, but this may be done without mention having been made of it in the literature I have seen.

Both the earlier molybdenite veins and the later tungsten ones almost certainly were emplaced under high-temperature conditions and probably far enough beneath the then-obtaining surface of the earth to be classified as hypothermal in non-calcareous rocks (i.e., Hypothermal-1).

The age of the Pervomaisk granite porphyries is reported (Materikov, 1977) to be 140 m.y. As the mineralization appears to have been established as of the same general geologic age as the Pervomaisk granite porphyries and these porphyries are 140 m.y. old, the age of the ores is late Jurassic or Middle Mesozoic. It must be remembered, however, that the mineralization was spread over an appreciable amount of time within the late Jurassic.

OZERNOE, BURYAT ASSR

Early Paleozoic	Zinc, Lead, Pyrites	Mesothermal to Leptothermal, after Sedimentary-IIa3

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Ripp, G. S., 1979, Towards the evaluation of conditions of formation of lead-zinc deposits of Buryat: Basic Parameters of Natural Processes of Endogenous Ore Formation: Nauka Press, Novosibirsk, v. 2, p. 109-118

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Notes

The text of the article of the Osernoe deposit in the Russian original of Smirnov's "Ore Deposits of the USSR", Volume II, says that the deposit is in the Buryat ASSR. The translator of Smirnov's volumes gives the coordinates of Ozernoe as ?ca53°00'N,114°00'E, which places the deposit not quite 50 km inside (to the west of) the border between the Buryat ASSR and the Chitinskaya Oblast of the RSFSR. Ozernoe, therefore, lies in the drainage area of the northward-flowing Konda river, a tributary of the Vitim river that for hundreds of kilometers forms the eastern boundary of the Buryat ASSR, and on the northwest slope of the Yablonovyy mountain range. The nearest city of any size is Chita (52°03'N,113°35'E) in the Chitinskaya Oblast, some 250 km slightly west of south of the mine and the nearest point on the Trans-Siberian railway; not the end of the world, but close to it.

Considering the location of Ozernoe, it is not surprising that it was not discovered until 1963 and was under exploration from 1964 to 1969. Presumably, mining would have begun there not long after 1969, but the isolated location of the deposit and remarks by Distinov suggest that mining has not begun. Mining is reported to have been scheduled to begin in 1982; when it reaches capacity, it is expected to produce 5,000,000 tons of ore a year.

Geologically, the deposit is located on the southerwestern folded margin of the Silberian plateau, the local subdivision of which is the Vitim plateau. The sedimentary beds in which the ores are contained are lower Cambrian in age and make up the central part of the Uda-Vitim eugeosynclinal zone of the Caledonian geosyncline of the Sayan-Baikal mountain range.

The ore are entirely contained in the huge complex of terrigenous-volcanic-carbonate formations in which the dominant form is tuffites; the total thickness of these beds is 5 to 6 km. The overlying middle Cambrian formations are in extreme contrast to those of the lower Cambrian, being of the red-bed-molasse type. The lower Cambrian formations have been thrown into steep folds with NE-SW axes, and this structure has been further complicated by a considerable series of faults of more or less east-west and NE-SW strike.

Distinov (1975) who, as the principal student of Ozernoe geology, has concluded that the textures of the primary ores show the normal series of sedimentary and diagenetic textures that demonstrate sedimentary deposition, lithification, diagenesis, to which are added appreciable evidence of later epigenetic events that resulted in the formation of zones of hydrothermal metasomatic ores and vein fillings. The problem of interpreting these textures is further complicated by local dynamometamorphism related to subvolcanic intrusions and dikes. This igneous activity has caused major recrystallization in which the primary textures involved have been completely destroyed. Distinov considers that much of the sulfide material was deposited as sulfide-bearing clays under calm-water conditions in conjunction with deposition of tuff-lime sediments; this melange of sediments solidified together to produce: (1) banded or brecciated sedimentary textures, (2) the sorting of the sulfides into indistinct strata based on grade-size changes from coarse-grained at the bottom of the sequence in question through globular to fine- (micro-) crystalline textures at the top of the series; (3) rhythmic sediments, some horizones of which are enriched in sphalerite, or pyrite, or siderite more than others with the ore-bearing rhthyms being concentrated in the upper parts of a given sequence. Sphalerite, however, was consistently the earliest sulfide to be precipitated; (4) globular and crystal-globular pyrite textures that have the appearance of metacolloids and aggregates of sphalerite-galena-pyrite in banded arrangements. Distinov thinks that such textures were developed by precipitation from gel-like sulfide-carbonate systems for which concept there appears to be no solid evidence; (5) diagenetic textures produced by the recrystallization of all the sulfides except pyrite that preserves its primary collomorphic textures. Distinov believes that the sedimentary muds (composed of sphalerite, galena, tuff, and carbonates) were recrystallized with quartz and carbonates being concentrated into distinct and separate aggregates that retained, however, their original banded character.

This polymetallic mineralization is contained in a synclinal structure (the Ozernoe syncline); all the sediments directly related to the ore in the lower Cambrian have been placed in the Oldynda group. On the basis of lithology and faunal remains, the Oldynda group has been subdivided in lower and upper subgroups. The dominant materials in the lower of these subgroups are tuffs and tuffites; this subgroup also includes subordinate amounts of carbonate beds. The upper subgroup, however, includes an appreciable representation of carbonate sediments in which fossil remains of archaeocyathids and algae are present. In each of these sedimentary subgroups, one or two members can be identified at the bases of which definite sedimentary rhythms can be determined; these are clearly associated with the pulse-like nature of both the volcanic and tectonic activity.

In the case of these members, their lower parts are composed principally of tuffs and lavas that are andesitic-dacitic and dacitic in composition; the upper parts are made up of tuffs, tuffites, and limestones. Although the sequence contains flows of lavas (largely dacites) that may be several tens of meters in thickness their proportion of the entire extent of the lower and upper Oldynda group is minor. The extrusive rocks are interbedded with tuffaceous formations of the Ozernaya member (the only member of the upper subgroup of the Oldynda group). The major portion of the sequence is composed of tuffaceous-carbonate sediments (calcareous tuffites) that include various amounts of tuffaceous andesite-dacite and dacite, carbonate, and terrigenous-pelitic debris. Decomposition of these beds has produced reef limestones and calcareous breccias that have a tuffaceous-carbonate matrix. Distinov holds that the entire sequence was saturated, to a greater or lesser extent, with fine disseminated sulfides of which pyrite (with a globular texture was the most important). This dispersed sulfide was most abundantly developed in what are now the ore horizons.

The Ozernaya member (or the upper subgroup of the Oldynda group) is divided into five horizons. Of these five horizons, the third and fourth from the bottom contain the four major Ozrenoe ore bodies; two of these ore bodies are in the third and two in the fourth horizon. The two ore-bearing horizons are clearly recognized (by Distanov) in the structure of the lower (and southeastern) limb of the Ozernoe syncline. In each of the two horizons, two subhorizons are readily identified. In the lower of the two horizons, the lower subhorizon is composed of calcareous breccias, tuffaceous gravels, tuffites, pyritic ores, and reefal limestones. The upper subhorizon of the lower horizon contains rhythmically stratified tuffites, dacite tuffs and pyritic ores. The lower subhorizon of the upper horizon is made up of limestones and such limestones as breccias that contain a pyritic ore body. The upper subhorizon of the upper horizon is composed of rhythmic interlayering of calcareous gravel tuffs, tuffites, and pyritic ore bodies. In the lower and southeastern limb of the Ozernoe syncline, the two horizons are easily distinguished; in this fold, the rocks have been (after sedimentation and probably lithification) converted to siderite. In the northwestern limb, the horizon consists of limestone, limestone breccias, gravel tuffs with bedded and brecciated ores that, in most instances, regularly alternate to exhibit definite sedimentary rhythms. The thickness of this productive sequence (third and fourth horizons) is between 140 and 230 m. The thickness of the Oldynda group is 1500 m out of the total of 5 to 6 km in the entire lower Cambrian stratigraphic column.

Three magmatic complexes are known to intrude the volcano-sedimentary beds of the ore field; these are: (1) subvolcanic intrusions, comagmatic with the automagmatic breccias of rhyolite-dacite character, dacite porphyries, diabases and diabase porphyries, quartz diabases, and quartz-diabase porphyries; thus, these igneous rocks also are lower Cambrian; (2) trachydacite-quartz porphyries and granosyenite porphyries that are assumed to be of Permo-Triassic age; and (3) Triassic-Jurassic alkaline basaltoids (dolerites and trachydolerites). All of the intrusive rocks of Cambrian and younger age cut through ore masses. Isotopic analyses of ore leads from the Ozernoe deposits give isotope ages that, when averaged, indicate also that the age of mineralization was Cambrian.

The major structures of the area (folds and faults) were active during a long period of geologic time, at least up to and including the Triassic-Jurassic. Some tectonic movements appear to have been contemporaneous with the volcano-detrital sedimentation. The major folding probably was early Caledonian; later earth movements resulted mainly in block faulting.

The northeast-southwest-striking Ozernaya syncline was broken by the Central and Southwestern faults into three tectonic blocks, not unreasonably designated as the Northeastern, the Central, and the Southwestern. The syncline itself is asymmetrical and has a curvilinear axis. In the northeast, the syncline has a periclinal closure that involves the strata of the first productive horizon of the Ozernoya member. To the southwest, the syncline plunges gently, the plunge locally being reverse by transverse flexures. The dips of the limbs of the folds and the shape of the closure differs considerably from one cross-section to another.

Distanov is convinced that the Ozernoe deposit is made up of layered accumulations of massive and disseminated pyritic-polymetallic ores, many times repeated. Most of the metallic minerals are contained in sedimentary structures, although appreciable numbers of epigenetic veins and veinlets cut the layered sulfides. Distanov believes that the veinlets and veins were developed through diagenetic processes and post-lithification hydrothermal open-space and replacement deposition. In the sequence of rocks in the productive layers of the Ozernaya member, he recognizes 10 stratified segregations of ore-rich layers. The uppermost of the ore-rich layers has been exposed by erosion and the lowest cut by drilling at 1300 m depth.

In association with the sulfide-rich layers, complexely shaped segregations of siderite (intersecting lenses) have been produced by replacement of carbonate, tuffaceous-carbonate, and sulfide deposits. These lenses of siderite are especially well-developed in the southeastern limb of the Ozernoye (ore-bearing) syncline.

The sulfide-rich ore layers are localized at several stratigraphic levels in the sedimentary sequence with ore layers being separated by barren volcaniclastic and sedimentary beds that are from 5 to 30 m thick. Where the ore layers occur near limestone reefs, their thickness is markedly different, and they cut off sharply against the surrounding sediments.

Distinov considers the layered pyritic ore bodies to be a normal part of the sedimentary sequence that lie between, and separate, the underlying, finely clastic portions of the sediments from the coarsely clastic material that overlies the ore layers. The lower boundaries of the ore layers are quite distinct, and the upper boundaries usually are much less so, grading from rich bedded and brecciated ores through a zone of decreasing content of disseminated sulfides to weakly or almost-unmineralized rocks. Where tuffite layers occur in intra-ore positions, the contacts are quite definite and

and exceptionally persistent. Internally, the pyritic ore layers are quite heterogeneous, being composed of alternating layers of pyritic ores of different grain sizes, barren seams, mineralized calcareous breccias, and siderite bands.

Where the Ozrenaya sequence has been folded, the ore layers are conformable with the barren rock layers, and the folds have NE-SW strikes, are normally asymmetrical, and have broad crests. Many late faults with both about east-west and NW-SE strikes and throws of 5 to 10 m, cut the sequence into a comples block structure.

Distinov recognizes two types of siderite ores, those that he considers are truly sedimentary and those that were formed by replacement; both types in many instances are spatially closely associated.

The veins, veinlets, and zones of veinlets containing polymetallic-sulfide and barite-sulfide fillings are contained in fractures parallel to the east-west and NW-SE faults; only locally do these vein structures have sulfide contents that are more than a few meters in width.

Distinov is convinced that the major fraction of the polymetallic sulfides was deposited by hydrothermal-sedimentary means, that is, the sulfides were added from hydrothermal sources at the same time that normal sedimentary processes were providing the chemical and detrital material intermixed with the ores. Obviously, the greater the influx of hydrothermal fluids at any given time, given the relative constancy of detrital and chemical sedimentation, the richer the layer in question would be in sulfides. This author also is convinced that the bulk of the vein-like mineralization was introduced at essentially the same time as the hydrothermal sedimentation occurred; the replacement phenomena involved presumably taking place in at least already partly lithified sediments in which fracture planes could be maintained. From this it follows that the introduction of hydrothermal fluids was turned on and cut off periodically during the sedimentary cycle. The remarkable thing is that the vein-like structures are confined to the ore-rich layers. If they are deposited from much the same hydrothermal fluids as those that deposit the ore minerals in the ore-rich layers, why should they not, in moving upward, precipitate ore sulfides in the barren rock through which they must move to reach the ore-rich layers in which they do form vein-like deposits? The problem does not appear to have occurred to Distinov nor to Smirnov and Gorzhevsky (1977).

Distinov reports that a study of the temperatures of formation of massive fine-grained banded ores by the 'thermobaric' method of pyrites and sphalerites has shown the absence of any decrepitation effect when they are heated, which suggests that the temperature of formation of this type is below 100°C. That this method of temperature-determination is subject to considerable question throws some doubt on these results. Distinov does, however, say that the ores in discordant ore zones are shown by homogenization methods to range between 160° and 200°C and the epigenetic veinlets in pyritic ores to have formed between 100° and 170°C. Finally, the discordant veinlets of carbonate-quartz-sphalerite were formed over a range from 220° to 360°C. He considers that the development of these last took place during the regeneration of ore material during the injection of post-ore magmatic formations. Of course, this last is the standard approach of sedimentationists to the problem of high-temperature minerals in deposits that they consider to have been produced almost entirely as low temperatures.

The principal minerals in the layered deposits are pyrite, sphalerite, and, to a lesser extent, galena. The main gangue minerals, either formed with the ore sulfides or in rhythmic alternations with them, are siderite, calcite, ankerite, dolomite, and a little quartz.

In the minerals that are formed by vein filling or by replacement centered around vein structures are larger in number but occupy far less space than do the bedded or layered minerals. These minerals are: sphalerite, pyrite, galena, native gold, native silver, and silver sulfosalts. The sideritic concentrations were developed after the ore in ore-rich layers, but also were introduced into in layers that contain no ores at all. Locally, in these concentrations of siderite, minor masses of barite with some galena and sphalerite contained in them are present.

Considerable numbers of other minerals, though in generally minor amounts have been developed, Distinov believes, through such processes are diagenetic, epigenetic, and metamorphic transformations. Along the borders of dikes appreciably younger than the ores, some contact metamorphic action has taken place. All of the silicate-rich rocks of the Oldynda group, volcanic and pyroclastic rocks, have been more or less sericitized, converted to K-feldspar, altered to hornfels and, to a lesser degree, to hydrobiotite. Still later, all the rocks of the region have been subjected to potassium metasomatism.

Distanov summaries what he considers the distinctive features of the Ozernoe ore as follows: (1) the major part of the ores have bedded structures; (2) all intrusive rocks of all ages represented (Cambrian and younger) cut the ore layers, veins, and segregations; (3) copper is almost completely absent from the Ozernoe layered ores; chalcopyrite is present only in the late hydrothermal-metasomatic ore-mineral deposits; (4) the trace-element content of the Ozernoe ores is low; the most typical elements of this type are: arsenic, antimony, cadmium, silver, germanium, thallium, and manganese; (5) average isotoic ratios of the various leads, 206, 207 and 208 to 204, indicate that the ores are Cambrian in age.

A good case is made by Distanov for accepting the idea that the bulk of the sulfides and of the two productive horizons were deposited by hydrothermal-sedimentary means. On this basis, the bulk of the sulfides should be categorized as deposits chemically concentrated in quiet waters by the introduction of fluid igneous emanations or water-rich fluids (some of the gangue minerals may have been brought in as detritus) and should be classified as Sedimentary-IIA3. The later vein-like deposits and the disseminated metasomatic siderite appear to have been deposited after essential lithification of the first and second productive horizons and should be classified as deposited by hydrothermal solutions under mesothermal to leptothermal conditions rather than under kryptothermal to epithermal, indicating that believe that this part of the ore in the Ozernoe deposits was introduced at an appreciable time after the lithification of the strata overlying the Oldynda group.

DZHEZKAZGAN, KAZAKHSTAN SSR

Late Paleozoic	Copper, Lead, Zinc, minor Silver, Rhenium	Mesothermal to Telethermal

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Yesenov, Sh. E. and Zaitseva, U. A., Editors, 1975, Geology and mineral

Notes

The town of Dzhezkazgan is located at 46°58'N,74°59'E in more or less the center of the Kazakhstan SSR; the nearest city of any size is Karaganda, (49°53'N,73°07'E) which is close to 400 km to the northeast. Karaganda is connected to Dzhezkazgan by a second-class railroad. The Dzhezkazgan concentrates, however, are shipped to the huge smelter at Balkhash. The Gory Ulutau (Ulutau mountains) form a barrier to the west of Dzhezkazgan, a barrier made all the more impressive by the desert character of the country that is emphasized by the apparent seasonal nature of the upper reaches of the Sarysu river on which Dzhezkazgan is situated. Production from the Dzhezkazgan ore field is between 17 and 18 million tons a year.

Although the Dzhezkazgan mine is the largest of the stratiform ore bodies in the area, there are others of commercial importance such as Itauz, Saryoba, and perhaps Shipisai, Akchii, Spasskii, and Karashoshak.

The first exploratory work in the Dzhezkazgan district took place in the first decade of the 20th century with a Russian geologist and an American one being the first in the area. At that time, Ball and Kozurev considered that the ores were epigenetic, but two other geologists, again one English and one Russian, said they were syngenetic. The controversy has continued to the present day, but the syngenetisists seem to be pulling ahead.

The Dzhezkazgan ore bodies are contained in an Hercynian structure, the Dzhezkazgan-Sarsyu trough. This structure was superimposed on an early Caledonian basement. The rocks in the trough begin with a lower sequence of rocks of volcanic origin of various compositions that are intercalated with red sandstones. Higher in the stratigraphy, the middle Devonian is represented by red sandstones and conglomerates; farther up are lower Carboniferous silicified limestones, sandstones, and marls; these beds have a total thickness of 1500 m. Above these beds are conformable and rhythmically alternating red and gray sandstones and siltstones with minor beds of conglomerate. These rocks are middle to upper Carboniferous and are some 700 m thick, and they contain the productive beds of the Dzhezkazgan group. In turn, these rocks are overlain by bright red sandstones, argillites, limestones and marls that are of early Permian age. This long sequence of Paleozoic rocks is covered in part by unconformable sequence of weakly cemented sandstones, clays, sands, and pebble beds that range from Paleogene to Miocene in age.

Within the Dzhezkazgan-Sarysu trough is an alternation of broad synclines with gentle anticlines and domes, the crests of which are, in many places, complicated by additional flexures. Locally, where the strata are steeply folded, reverse and overthrust faults developed. A number of deep-seated faults, striking more or less east-west and north-south have been located by geophysical means; these faults merge directly south of Dzhezkazgan.

The northern portion of this Dzhezkazgan-Saryso trough contains two anticlines (Zhanai and Kengir) between which lies the Dzhezkazgan syncline. In the vicinity of the area of periclinal closure of these two anticlines are the bulk of the Dzhezkazgan ore bodies. Within the Dzhezkazgan ore field are a number of domes and gentle anticlines; the domes are designated as the Porovo, Krestovo, Nikol'sk-Annen, and Akchiya with synclines between them. These structures are further complicated by less prominent folds. The Zhanai and Kengir anticlines and the Dzhezkazgan syncline have axes that strike NE-SW and have limbs that dip from 8° to 25°; the crests of these anticlines are from 900 to 1300 m wide with the beds exposed there being essentially flat. The limbs of the structures were locally subjected to minor folding by which the dips of the major structures are increased to as much as 50°. Further, the major structures are cut by faults in which the throws are from 10 to 30 m and increasing locally to as much as 80°. The intensity of the folding in this area gradually decreases with depth.

The ore bodies of the trough area are contained in the rocks of the Dzhezkazgan group that are divided into 51 rhythmically alternating layers of gray and red sandstone, siltstones, and conglomerates. The proportions of red and gray rocks are about equal. Although the colors of these beds are determined by the valence state of the iron they contain, their mineral content is essentially the same. The thickness of these siliceous layers of red rocks ranges between 2 and 28 m; those of gray rocks between 7 and 42 m. Siltstones are the main rock type in the red rocks and among the gray rocks sandstones. The clasts in the gray rocks are mainly quartz and feldspar, and the matrices normally are silica and carbonate.

Where the sandstones have been mineralized, they are dark-gray, and the abundance of sulfides in them gives them a metallic appearance. The gray sandstones have a definite parallel stratification that is made the more prominent by intercalations and lenses of the red sandstone. The lower boundaries of these beds are quite sharp, but they grade gradually into the layers that overlie them.

The red sandstones have somewhat smaller grains and a clay-carbonate matrix. These sandstones are typically cross bedded. The siltstones are compact rocks green-gray in color and conchoidal in fracture and composed of fine-grained clay minerals. The red siltstones, much more abundant than the gray ones, make up the bulk of the barren layers; in all other respects they are the same as the gray variety.

There are two types of conglomerates: (1) intraformational and (2) in-interformational. They range in thickness from 0.8 to 2.0 m and may grade, along strike, into sandstones and less frequently, into siltstones; the intraformational type may contain small fragments of rocks in the Dzhezkazgan sequence, and their matrix is gray sandstone. The interformational varieties have pebbles that came from rocks not in the Dzhezkazgan group; ordinarily they are thicker than the intraformational type; they occur only in three layers in the middle of the group.

The Dzhezkazgan group has been divided (Yedresov and Yesenov, 1969) into lower, middle, and upper subgroups. The lower has the highest amount of siliceous material in its sandstone matrix and has a marker bed of silicified limestone at its top - this subgroup is 230 m thick. The middle group contains the three layers of interformational conglomerate; it is 200 m thick. The middle subgroup turns gradually into the upper, has a large amount of carbonate in its sandstone matrix, and is 215 m thick.

The Dzhezkazgan ores always are in gray rocks, no matter what their lithology may be, with the quantities of copper, lead, and zinc being 4 to 5 times larger than in the red rocks. Commercial mineralization is found as far as 600 m below the top of the group; nothing better than very weak mineralization is found below this 600-m level.

Of the total of 51 recognizable layers in the Dzhezkazgan group, 26 (all in gray rocks) are mineralized, but only 19 of the 26 are of commercial grade. For mining purposes, these mineralized layers have been combined into 9 ore-bearing horizons, separated by red layers; each of the ore horizons contains more than 1 ore layer and may have as many as five.

The lower subgroup is composed of two ore-bearing horizons, the Tashuduk and Zlatoustovsk; the middle four such horizons, the Pokrovo and the three Raimundovsk; the upper subgroup has three ore horizons Krestovo, Akchiya, and Annen. Any single stratigraphic horizon is not entirely ore, being made up of several discrete ore bodies quite close to each other. A number of these ore bodies (2 to 30) are combined in plan into "segregations"; the entire Dzhezkazgan area contains more than 300 ore bodies and over 100 "segregations".

Each segregation in a given mineralized zone forms an elongate oval in plan. In the upper group, these ovals have their long axes running in a NE-SW direction and make up an almost continuous band (1 km wide) from the eastern limb of the Zhanai to the closed portions of the Kengir anticlines. In the middle subgroup, the segregations form as narrower band (0.4 km wide) that circles around the nose of the Kengir anticline; these segregations cut the folds of this anticline and the faults associated with them at almost right angles. In the lower subgroup, the ore segregations run NW-SE.

The form of the ore bodies is layered, the layering being that of the beds in which they are contained; usually the bodies are oval in shape with the length being about 5 times the width, though some may be essentially equidimensional. In the higher horizons, however, the ore bodies may be much larger with lengths that may be several kilometers long and widths that run from 50 to 100 m or even more. These ore bodies are separated into three categories: large, medium, and small. Large ore bodies are those with lengths and widths mentioned above: small bodies have their long dimension in hundreds of meters; medium bodies lie between the two extremes. As is true of the segregations (see above), the long dimensions of the large ore bodies trend NE-SW; the medium ore bodies are elongated from NE in the north of the ore area to NW in the south; again, in the small ore bodies, the elongation is NE-SW (Samonov and Pozharisky, 1977).

No geologic feature marks the boundaries of these ore bodies in the upper horizons; they change from ore to non-commercial over rather short distances. The major commercial elements (Cu, Zn, Pb) normally are quite evenly distributed over an entire ore body; in many instances, however, the copper content differs appreciably from one part of the ore body to another. In some place, within mineable rock volumes, areas of marginal ores may be encountered.

The crests and limbs of the anticlines and the minor folds that complicate them contain the best ores; in these situations, the ore normally reaches its maximum dimensions. Yesenov and others (1971) calculated that 40 per cent of the total reserves in the Dzhezkazgan district are contained in the arched portions of anticlines or domes, 35 per cent in the limbs of these structures and in the minor flexures; the synclines include only about 25 per cent of the total ore.

In addition to the Cu, Zn, and Pb, Ag and Re are present as traces, but As, Cd, Bi, Co, Hg, Au, Ni, and Mo are even less in amount. The mineral content of the ores is quite simple. Almost all of the copper is contained in three minerals: chalcopyrite, bornite, and chalcocite, in the higher levels. At intermediate depths, bornite dominates, and chalcocite is most important in the deepest portions. This arrangements of minerals is true not only for the ore region as a whole but for the individual ore bodies as well. In such individual ore bodies, chalcocite is the principal ore mineral in the deposit center, farther from the center, bornite takes over, and on the margins, chalcopyrite. In the marginal volumes lead and zinc increase relative to the more internal volumes of the ore bodies. Galena is the main lead mineral, Sphalerite (mainly the whitish form - cleiophane) is the principal zinc mineral. Such silver as is present is there in various silver minerals, including native silver, but it also may be present in the lattices of the various ore minerals; chalcocite contains the most silver with rhenium in the copper minerals, mainly bornite. Dzhezkazanite is questionably a lead-rhenium sulfide. The minerals present in traces are: pyrite, marcasite, arsenopyrite, betekhtinite [$Cu_{10}(Fe,Pb)S_6$], various arsenides of copper and cobalt, and tennantite-tetrahedrite.

The ore minerals in the deposits at Dzhezkazgan are most abundantly present as replacements of the matrices of the host rocks and, less commonly of feldspar and quartz. The sulfides inherit the banding of the primary detrital and sedimentary minerals. Where mineralization is weak, the ore minerals preferentially inhabit the seams with the coarsest grain. In the richest ore, replacement has been so complete that the ore comes close to being classifiable as massive, but even this "massive" ore retains traces of the primary banding. In conglomerates, pebbles are hardly attacked by the ore minerals; these last are confined to the matrices.

There has been wide-spread development of vein-type mineralization, but its total amount is small and has essentially no commercial worth. These veins-type ores normally are confined to single layers, but some veins pass uninterruptedly from gray to red rocks. Normally, the veins contain the same minerals as the bedded deposits, but, in places, veins of galena may cut bedded copper minerals. The gangue minerals of veins of this type are quartz, barite, and calcite; some veins may be confined to brecciated rock volumes.

Although the main worth of the Dzhezkazgan ores comes from the copper minerals, four commercial types are recognized: (1) copper; (2) complex (copper-zinc, copper-lead-zinc, and copper lead): (3) zinc and lead-zinc, and (4) lead. Some ores are largely or partly oxidized, but these comprise less than 2 per cent of the total Dzhezkazgan ores. The complex and lead ores may make up distinct ore bodies at the borders of the copper ore segregations.

All the ore bodies that outcrop have been oxidized to produce the usual arrangement of oxidized, leached, and secondary sulfide ores (from the surface downward). These processes have not changed the quantities of copper appreciably, probably because of the large amounts of carbonate matrix in most of the ore sandstones. Normally, the zones of oxidation and leaching are from 8 to 12 m thick, although, in the vicinity of faults, these processes extend to considerably greater depths. The zone of secondary sulfide enrichment lies from 10 to 70 m below the surface. The main oxidation minerals are: hydroxides of iron and manganese, native copper, malchite, azurite, brochantite, ehlite (unknown to me), and chrysacolla. The secondary sulfides are: covellite and secondary chalcocite and bornite.

The controversy between syngenetic and epigenetic manner of formation for the Dzhezkazgan ores has raged almost from the first two papers published on the ore body.

Samonov and Pozharisky (1977) summarize the arguments for each concept; their ideas are further summarized here. The syngenetic concept is favored by the following facts: (1) the position of the ore bodies definitely is controlled by the stratigraphy; (2) the ore bodies are confined to near-shore-marine deltaic sediments; (3) the ores widely distributed; (4) the economic components of the ore bodies are relatively widely distributed (although this distribution is somewhat less than entirely even as is shown by the discussion just given above); (5) no magmatic rocks are associated with the mineralization; and (6) the amount of organic matter in the ore bodies is appreciably greater than in the surrounding rocks.

The epigenetic concept is favored by the following facts: (1) the ores are confined to areas immediately adjacent to deep-penetrating faults, making highly possible the tapping of sources of magmatic ore fluids (however, mineralization of the ore type is essentially lacking in such faults); (2) the ore body locations definitely are controlled by structures, particularly the upper limbs and the crests of various anticlines; (3) the ore are polymetallic in composition and typical hydrothermal minerals are abundant in the deposits; (3) the ores were, in many places, developed in more than one phase; (4) the ores were not emplaced by the filling of open space but by the replacement, mainly of gangue minerals; and (5) the wall rocks of the ore show alteration that is more probably related to hydrothermal ore deposition than to syngenetic-diagenetic modifications.

The argument against the validity of the syngenetic criteria takes the following form: (1) the stratigraphic control of the ore bodies can as easily have been exerted on hydrothermal fluids moving up through the formations as of surface waters moving through them - the emplacement of the ores in the various folds strongly suggests that the rocks had been sufficiently lithified to fold as competent beds before the ores were introduced into them; (2) the confinement of the ores to near-shore facies simply may mean that these facies were the most permeable to hydrothermal fluids and most favorable for deposition from such fluids; (3) the wide and real distribution of the ores merely may mean that the ore fluids were provided to the area in large enough amount and from sufficient source channels to insure this wide distribution; (4) the distribution of economic components in the ores is far from completely even and a definite zoning is exhibited by them, which is as explicable by deposition from hydrothermal fluids as from volcanic emanations or surface waters; (5) the lack of known magmatic rocks is to be expected in deposits that show moderate to low temperatures of deposition of the ore minerals; and (6) organic matter can have been a relic of original deposition and would have as much effect on hydrothermal solutions as on surface waters or magmatic emanations.

On the contrary, the arguments for epigenesis can be refuted perhaps equally as well: (1) deep-seated faults may have been the channels for the entry of volcanic emanations to reach the sea floor as for hydrothermal fluids to reach the folded rocks; (2) the structural control argument is somewhat more difficult to refute but the concentration of elements in such structures may have been due to remobilization of primary ores; (3) the stages of ore mineral deposition and of the polymetallic composition of the ores may have been the result of variations in emanation composition and, to some extent, to later remobilization of the primary ores; (4) the alteration of the country rocks may have been due to fluids developed during metamorphism later in the cycle of geologic events, weak though that metamorphism may have been; (5) the confining of the ores to specific horizons may better be explained as the result of cyclic sedimentation that of hydrothermal fluids being considered to have been so particular as to the rock layers in which they deposited their mineral loads.

On balance, it seems to me that the points in favor or an epigenetic origin and against a syngenetic one are sounder than those for the reverse view, but the argument certainly has not been settled.

I would classify the ores as mesothermal to telethermal with the caveat that they may have been formed by sedimentary processes in which volcanic emanations or surface waters may have been the ore fluids; such a situation should be classified as "deposits chemically concentrated in quiet waters by inorganic reactions or by the introduction of fluid igneous emanations", that is: Sedimentary-A1a and/or Sedimentary-A3, the latter being preferred.

If the deposits were basically sedimentary, their age would be that of the rocks in which they are contained, that is middle to late Carboniferous or late Paleozoic.

If they were formed by deposition from hydrothermal fluids, the age of that event probably would have been that of the Hercynian orogeny or, again Late Paleozoic.

KOUNRAD, KAZAKHSTAN SSR

Late Paleozoic	Copper, Molybdenum	Hypothermal-1(?) to Mesothermal (Primary) Grd-Water Circulation-IIB2 (Secondary)

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Pavlova, I. G., 1978, Copper-porphyry deposits (rules of distribution and criteria of prognosis): Nedra Press, Leningrad, 273 p.

Samonov, I. Z., and Pozharisky, I. F., 1977, The Kounrad deposit, in Smirnov, V. I., Editor, Ore Deposits of the USSR, Vol. II, Pitman Pub. Co., London, p. 120-127

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Notes

Although the mineralization at Kounrad had been known since at least the early nineteen hundreds, the actual discovery dates from the work of M. Rusakov in 1928. The original name of the deposit, if the 1916 map of the deposit by A. Tim is to be believed, was "Konrad". This ore body was the first porphyry-copper deposit to be found in the Soviet Union; the deposit is located about 19 km nearly due north of the city of Balkhash (46°50'N, 74°57'E) (the coordinates of Kounrad are 46°58'N,74°59'E). Balkhash lies on the shore of a major lake, Ozero Balkhash, that extends in an irregular half circle (convex northward) for nearly 500 km. Balkhash is about 200 km from its southwestern end and is the site of the large Balkhash smelter. The Kounrad concentrates provide about 40 per cent of the feed for this operation, 50 per cent of the concentrates processed here come by rail from Dzhezkazgan, and 10 per cent various neighboring small deposits. Kounrad produced 14,000,00 tons of ore per year with a grade of 0.8 to 0.9 per cent copper average, though the grade ranges between 1.1 per cent and 0.7 per cent. Gold production amounts to about 8.4 tons per year, and molybdenum and rhenium are by-products.

The first work on the open pit from which the Kounrad ore was to be extracted began in 1934; by 1954, the depth of the pit had reached 150 m, and it appeared that the end of the known reserves would be reached in a short time. Additional reserves, however, were found in the mid-1950's and equipment was modernized and the pit expanded to 1600 x 1200 m with the plans calling for the final depth to be 400 to 500 m. During the period from 1957 through 1964, overburden removal fell 20 per cent behind the planned tonnage, causing considerable difficulty in producing ore in the early and mid 1950's. However, everything was back on track by 1968, permitting the planned tonnage to be reached by 1968.

The Kounrad deposit is contained in the northwest corner of a massif of secondary quartzite and altered grandiorite porphyries that stands 150 to 200 m above the surrounding featureless plain below it. The massif is composed of a central ring-like mass of the granodiorite porphyries that stand lower than the surrounding ridges and isolated hills of the secondary quartzites. The granite porphyries contain the ore bodies, whereas the quartzites are replacements of eruptive rocks.

Two general explanations have been offered for the situation that obtains at Kounrad. Such authors as M. Rusakov and K. Gazizova considered that the marginal parts of the hill (what are now thought to be secondary quartzites) were felsites that were intrusive ring dikes or the roots of felsite flows that lay on the eroded surface of the granodiorite.

The second concept was put forward by N. Nakonvik (1937) and has been rather generally supported since. He demonstrated that the silicic altered volcanics were cut by the porphyrites; thus, the silicic rocks could not be the source of the ore fluids. Nakonvik's hypothesis, and it has strong support today, was that the silicic rocks (the remains of silicic extrusives) are all that remains of a volcanic structure of the central variety and that the granodiorites are a sub-volcanic intrusive body located in the old volcanic vent.

The Kounrad deposit is located in the eastern part of the Tokrau basin in central Kazakhstan; this large structure was formed in Hercynian time and was imposed on a basement folded in Caledonian time and composed of heterogeneous rocks of Devonian and older (?) ages. The uplifted oval is 40 x 15 km in size; almost the entire area is made up of intrusive rocks that are the southern part of the Kounrad-Bedtauta pluton, one of the largest of such igneous massifs in Kazakhstan - sedimentary rocks compose only about 10 per cent of the total volume of the pluton. The oldest of these sedimentary rocks are deposits of Famennian (uppermost Devonian) age; these rocks consist of siltstones and argillites that include seams of sandstones, lenses of cemented gravels, and pebbly conglomerates, and are west and northwest of the copper deposit.

Tournaisian (beds immediately above the Famennian) occur around the margins of the ore body and consist of two types of extrusive rocks, mafic (older) and silicic (younger). These mafic extrusives were formed by an alternation of andesite hornblende porphyrites, diabases, amygdaloidal porphyrites, andesite-dacites, and several tuff formations. The silicic extrusives cover a much wider area than do the mafics and occur as hills around the ore body and make up an appreciable part of the upper levels of the open pit; most of these eruptives are now secondary quartzites. The relative ages of these two eruptives are shown by the silicics lying on the mafics and fragments of mafics being found in the silicic type.

A few probably Cretaceous pebble beds (cemented by sand-clay) are present on the border of the secondary quartzites. Twenty to 10 m of rocks produced by Mesozoic-Tertiary weathering originally overlay the ore deposits but were removed, of course, prior to open-pit mining.

Directly to the south of the ore deposit is a large massif of grandiorites, quartz, diorites, and diorite porphyries; these are thought to belong to the widespread Topar complex of middle (?) Carboniferous age. A lower Permian complex of diorite-porphyry dikes, quartz diorites, granodiorites, and granites forms a concentric belt around the secondary quartzites on the north and northwest. On the east, leucocratic granites of the Akchatau complex border the quartzites; they are Permian in age.

The ore bodies at Kounrad are contained largely in the apical part of a gronodiorite-porphyry stock that is the end stage of the Topar complex. The mineable ore has an irregular oval shape, being somewhat longer in a NE-SW direction than in a NW-SE one. Geophysical studies have shown the ore to be conical in shape with the northwestern and northern sides being the steeper. At 2.0 to 2.5 km below the surface, large negative gravity anomalies have been discovered, suggesting the presence of an unmineralized granitoid mass. In detail, however, apophyses extending out from the granodiorite porphyry stock, and blocks of sunken eruptives complicate the pattern of the ore body.

Practically all of the host rock of the ore body is granite porphry altered to secondary quartzite that contains well-defined textural relics of the porphry. The rocks are broken by joints with a variety of attitudes, essentially all of which are filled with: (1) quartz; (2) quartz with disseminated pyrite; (3) to an appreciably lesser degree, sulfides and quartz. The types of ore veins are: (1) monomineralic - chalcocite or chalcopyrite or molybdenite and (2) polymineralic pyrite-molybdenite; tennantite-tetrahedrite-pyrite-chalcopyrite; and chalcopyrite-tennantite-tetrahedrite-galena-sphalerite.

In the central portions of the ore body, chalcopyrite and bornite make up the major part of the ore minerals with a gangue consisting of quartz, kaolinite, andalusite, sericite, tourmaline, alunite, pyrophyllite, epidote, and diaspore. The veins may be as thick as 10 cm, but they range downward from that width to fractions of a millimeter; these veinlets average 1 to 2 cm in width. The lengths of the veins range from 10's of centimeters to a few meters; unusually, they may be tens of meters long. The amount of rock separating the veinlets may be from 0.1 m to as much as 40 to 60 m; the average separation distance is 5 to 10 m.

The essentially barren quartz and quartz-pyrite veinlets are much smaller than the other types, but they are much more abundant. Locally, these quartz-pyrite veinlets are so common that they constitute a uniformly veined mass with only small relics of the original rocks between them. Specifically dominant directions of veins never have been established. With depth, however, the number of veins decreases but their width increases. Although these veinlets are the most visible type of mineralization at Kounrad, the most of ore minerals are present as rather uniformly distributed, but submicroscopic, segregations. These segregations range from a millimeter in diameter up to several millimeters, but the average is 0.2 to 0.5 mm; they contain much the same minerals as do the veins. In places almost barren blocks of rock (10-15 m on a side) are present.

Around the ore-bearing stock, the secondary quartzites are compact, very hard rocks of various pale colors; they may have a somewhat banded appearance.

The rocks on both sides of the contact of the granodiorite with the eruptives were metasomatically converted to what are referred to by Nakovnik (1937) and Yudin (1969) as "monoquartzites", that is, rocks from which essentially all constitutents except silica have been removed; this zone is from 10 to 60 m thick. In both directions from this "monoquartzite" zone, the alteration became less drastic. Here quartz-sericite rocks have been developed from the granodiorites; when these quartz-sericite rocks are followed down to depths of 600 to 650 m, they change to propylitized and partly argillitized rocks. On the outer side of the contact, as the quartzites are traced farther out, they become andalusite- and then sericite-quartzites. The thickness of the andalusite quartzite is from 100 to 300 m and of the sericite quartzites 200 to 400 m. Out beyond the sericite-quartzites is a 200- to 300-m thick zone of propylites.

This entire sequence of altered rocks (granodiorites and extrusives) have been intersected by two types of dikes: (1) fresh-appearing quartz diorites and diorite porphyries that almost certainly are post-ore; these dikes are assigned to the Zhaksytogala complex and (2) sparsely distributed dikes that certainly are post-ore. If the relative ages of these two sets of dikes to the ores is correct, then the age of the copper mineralization is definitely late Carboniferous, that is late Paleozoic.

Along the southern border of the open pit, unusual explosion breccias are present; the breccia fragments range in size from almost neglible to 0.5 m and their shapes have all sorts of irregular forms. Although most of the explosion breccias are irregular in form, some are vein-like with widths of 0.5 to 1.0 m and lengths of as much as 100 m.

The Kounrad ore body is a stockwork-type deposit; its more or less equidimensional plan and its bowl-like shape in section was determined by a "conical-circular" set of joints (Gazizova,1957). The grade of the mineralization changes so gradually and is so difficult to judge by the naked eye, that the boundaries between ore and waste can be found only by sampling. The two sets of joints that make up the stock work were the earlier concentric variety and the later vertical variety.

The ore in the Kounrad stockwork can be divided into five zones as follows (Gazizova, 1957):

Leaching zone	0 to 80 m thick	average thickness 25 to 30 m
Oxidation zone	0.5 to 60 m thick	average thickness 2 to 25 m
Mixed ore zone	0 to 50 m thick	average thickness 10 to 15 m
Secondary enrichment sulfide zone	15 to 270 m thick	average thickness 130 to 140 m
Primary zone	- -	average thickness over 400 m

As might be expected, the leaching zone almost never is ore, the quantity of copper left in it being rarely as much as 0.2 per cent. This zone is most varied in thickness and is absent in some places and may, in others, extend to 80 m below the surface. It is normally rather pale in color, locally being almost white.

In the oxidation zone, more than half the copper content is in oxide minerals; it is extremely varied in thickness and mineral composition with the maximum copper grade being as much as 1.5 to 2.0 per cent. Oxide ore may be separated by essentially barren volumes. This zone at Kounrad is now exhausted. The ore minerals present included hematite, limonite, jarosite, malachite, azurite, brochantite, atamcamite, chrysocolla, native copper, cupirte, and (appearently relict) chalcopyrite.

In the mixed zone, copper is 25 to 50 per cent in the oxidized form and 50 per cent or more is present as sulfides; it grades gradually into the overlying and underlying zones.

The secondary sulfide zone has provided, and will provide, most of the Kounrad copper. The major minerals in this zone are, in order of lessening abundance: chalcocite, pyrite, chalcopyrite, bornite, covellite, sphalerite, enargite, arsenopyrite, tennantite-tetrahedrite, and molybdenite. This belt of ore is a manto-like body with a well-marked depression in both the northeastern and southwestern portions, a structure inherited from the form of the primary ore or produced by a concentration of water-carrying structures in those sections so depressed. The top of this zone slopes to the southeast, dropping 30 to 40 m in crossing the ore mass. The vertical thickness of this ore ranges between 100 and 270 m with the thickest ore being in the southwest and the thinnest in the northeast. The ground waters circulating through the fractured (and altered) granodiorite appear to have been largely confined to that rock volume by the shape of the basin and the pattern of its fractures. The greatest depths to which secondary enrichment reached were determined by the depths to which the fracture zone permitted the ground waters to extend. The strong flow of ground water out of a discharge channel at the southeast end of the fractured structure disrupted any secondary enrichment that had earlier formed there. Where the pattern of fractures in the ore zone slowed, but did not prevent, the movement of ground water was a favored volume for the production of chalcocite.

With the removal of much of the ground water circulation from the lower parts of the ore body and the lack of the ions necessary for the reduction of copper brought the process of secondary enrichment to an end, leaving the remaining primary ore largely unaffected by any nearer-surface processes. The major primary minerals are; pyrite, chalcopyrite, molybdenite, enargite, and tennantite-tetrahedrite; minor minerals are sphalerite, galena, bornite, and magnetite and very rare minerals are arsenopyrite, marcasite, pyrrhotite, idaite [Cu_3FeS_4], and native gold and silver.

Most of the molybdenite is contained within the volume occupied by the copper ores, but molybdenite is found over a wider area than are the copper minerals. The form of both mineral volumes in the of a funnel, small end down. Where there are major concentrations of molybdenite, they are located on the far margins of the copper ore body.

Most of the copper ore is disseminated in the altered granodiorite with but a minor fraction of that ore confined to veins; a small amount of the ore is as breccia-fillings. Much of the ore, both primary and the later secondary was emplaced by the replacement of earlier sulfides.

The hypogene mineralization, on which, of course, the secondary (or hypergene) depends has been divided into various stages by different students of the deposit. Yudin's (1969) scheme appears to be preferred in which he sees all the hypogene minerals as having been formed in a single episode. This suggestion is favored because paragenetic sequences are similar throughout the deposit and because the range of temperatures of mineral formation are quite constant between 460° and 110°C with the bulk of the primary sulfides having been deposited in the upper portions of that range.

Although the literature I have seen does not specifically state that the chalcocite is all the result of secondary enrichment, the list of primary sulfide minerals in the deposits (given by Samonov and Pozharisky, 1977) does not include chalcocite. Yet these same authors, in defining the two types of veins in the ore body, says these are; (1) monomineralic in which the one mineral can be chalcocite and (2) polymineralic with chalcocite not given as a possible constituent. Of course, chalcocite could substitute for chalcopyrite in the monomineralic veins by the replacement of chalcopyrite, but I see no reason why, in the zone of secondary enrichment, chalcocite could not be emplaced in the polymineralic veins by replacing anyone of the wide variety of sulfides and sulfosalts available for that process. At any event, I am assuming for the purposes of genetic classification that chalcocite is not a primary mineral or, if it is, one of very minor importance.

From the fact that the minerals making up the primary assemblage are typical of the mesothermal range, but the possibility exists that some of them were deposited in the leptothermal range (the sulfosalts mainly) and that the early molybdenite might have been precipitated in the lower portion of the hypothermal. Because the molybdenite is a mineral of economic importance, I am including with the Mesothermal designation for the Kounrad deposits, "Hypothermal-1 (?)"; I am omitting "Leptothermal" although some of the native gold may have been deposited in that range and is recovered in the processing of the ore.

The oxidized and secondary ore deposits were formed by the circulation of ground water and are classified as Ground Water-B2.

MIRGALIMSAI, KAZAKHSTAN SSR

Late Paleozoic (or Middle) Lead, Zinc, Barium Mesothermal

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Notes

The Mirgalimsai mine is located at 43°32'N,68°36'E on the southwestern slope of the Karatau range in the southern part of central Kazakhstan about 275 km slightly west of north of Tashkent (41°16'N,69°13'E), the capital city of Uzbekistan. Mirgalimsai is in the immediate vicinity of the town of Kentau (43°28'N,68°36'E) and is connected from Kentau by rail with both Tashkent and the Kazakhstan administrative center of Dzhambul, the railway to the latter city passes through the smaller city of Chimkent, the site of a major smelter.

The Mirgalimsai deposit was discovered in 1928, was prospected from 1929 to 1936, but was not put into production until 1942 when World War II needs for lead and zinc dictated that this should be done.

The Mirgalimsai ore region is located in carbonate rocks of the Famennian stage (uppermost Devonian) and of the immediately overlying Tournaisian beds of the lowermost Carboniferous. The strata in the Famennian show alternations of limestone and dolomite beds that differ in structure from massive to ribbon-like to lumpy. The Famennian is separated into 11 horizons that range in thickness from 5 to 70 m. The Mirgalimsai ore horizon is the 7th from the base of the Famennian beds. The Famennian beds are, from bottom to top: (1) Argillitic, 90-300 m of clay-tuff clastic rocks with intercalated calcareous sandstones and dolomitic limestones; (2) Transitional, 35-60 m of limestones with seams of dolomite; (3) Dolomitic, 40-60 m of dolomites; (4) lower Lumpy, 50-70 m of coarse-bedded limestones and dolomites; (5) lower Platy, 10-15 m of limestones and dolomites; (6) Cellular, 5-10 m of dolomites and limestones; (7) Second Ribbon-Like (where is the first?), 10-15 m of dolomitized limestone, calcareous dolomites, and limestones that are thin-bedded-massive, and lumpy - the Ore Horizon; (8) middle Lumpy, 40-70 m of coarse-bedded limestones; (9) Platy, 20-35 m of limestones and dolomitized limestones; (10) upper Lumpy, 30-50 m of limestones and dolomites; and (11) upper Ribbon-Like, 15-20 m of dolomites.

Conformably above the Famennian are the Tournaisian beds that are divided into five horizons that range in thickness from 20 to 280 m. These horizons (from bottom to top) are: (1) Basal, 15-135 m of limestone breccias with fragments and blocks of dolomite; (2) Sparkling, 20-70 m of massive, locally bedded domolites; (3) First Interbedded, 100-160 m of interbedded limestones and sedimentary limestone and dolomitic breccias; (4) Second Interbedded, 160-200 of interbedded limestones and limestone breccias; and (5) Banded, 280 m of fine interstratified light-gray and gray limestones.

Characteristically, the ore horizon (Second Ribbon-Like) contains appreciable quantities of organic remains with the amount of organic carbon being 4 to 5 per cent and the soluble bitumens 0.77 per cent. The bitumen contains paraffins (7.6 per cent), oils (27.0 per cent), and resins (49.2 per cent). The portion of the organic matter that is soluble is highly metamorphosed bitumens of the kerogen type. The thickness of the formation has a wider range than the stratigraphic table indicates, being from 2 to 4 m up to 24 to 28 m (Smirnov and Gorzhevsky, 1977).

The Devonian and Carboniferous rocks just described (Famennian and Tournaisian) have been folded into a series of brachyanticlines and synclines, the axes of which normally strike NW-SW. The major difference between the anticlines and synclines is that the latter normally are wide and trough-like whereas the anticlines are appreciably more complex and exhibit cores of Famennian rocks. A widespread system of faults characterizes the ore field; these are NW-SE and nearly east-west-striking structures that are prominently brecciated and crumpled and are accompanied by zones of variously oriented jointing that result in fault zones, the thickness of which ranges from 10 to 20 m or even more. These faults in most instances are reverse or overthrust in type. The displacements on these faults, are between a few 10's of meters up to 1.5 km. The largest overthrusts, especially the Main and Southern, aided by a major number of other, lesser faults cut the beds up into a pattern of blocks. The crests of most of the folds, mainly in the southern part of the ore field, are broken into overthrust faults, causing much of the northern or north-westerly displacement of the beds; all of these faults are pre-ore.

The rocks of the Mirgalimsai ore field range from dolomites to weakly dolomitic limestones and show almost every variation between these two extremes. In the center of the ore field (Zorin, 1969) the most dolomitized rocks are centrally located and contain less and less magnesium as the periphery of the ore zone is approached. The farther down the ore zone is followed, the more widespread is the complete dolomitization; the nearer to the surface the rocks may be positioned, the smaller the central core of dolomitization. The less magnesium in the ore-field rocks the more uniform their texture; the fully dolomitized rocks are much more complicated in texture.

The most widely developed textures in the beds is micrite in both lime-bearing dolomites and dolomites with the grain sizes in the matrices less than 0.01 mm; various types of sparite (dolomite and calcite matrices) also exist in which the grains are larger than 0.01 mm. These micrites and sparites have been developed by recrystallization of pelitomorphic (clay-sized) and other of the various matrix components of the primary sediments.

The process of dolomitization and baritization of the ore horizon originally was thought to have occurred after the rocks of this horizon had been lithified, but Zorin (1969 and later) argues that the various metals (lead, zinc, barium, and those of lesser abundance) were added at the same time as the introduction of magnesium, that is, during the primary sedimentation and diagenesis, as well as during catagenesis. Thus, the relationships among the various minerals that now exist were produced over a considerable period of time, from when the rocks were laid down until they were lithified but still were under a thin covering layer (catagenesis). The contrary opinions of Yushko (1969) will be discussed below.

The ore minerals in the ore-bearing horizon are disseminated or massive enough to be considered layered; also present are small ore-mineral veinlets and small masses of uniform sulfides. In an opinion credited to Kudenko and Stetsenko, two varieties of ore are recognized at Mirgalimsai: (1) dolomitized and baritized rocks with layered and veinlet mineralization and (2) weakly baritized rocks with veinlet mineralization. On the basis of composition, however, three types of mineralization have been identified: (1) lead, (2) lead-barite, and (3) barite. The major minerals in the deposit are pyrite, galena, and sphalerite and dolomite, calcite, barite, ankerite, and quartz. Minor minerals include: chalcopyrite, tennantite-tetrahedrite, including freibergite, chalcocite, native lead, argentite, sulvanite [Cu_3VS_4], native silver arsenopyrite, magnetite, hematite, witherite, and fluorite. Zone of oxidation minerals are: anglesite, smithsonite, hydrozincite, calamine, wulfenite, chalcocite, covellite, argentite, native silver, hydrohematite, goethite, psilomelane, pyrolusite, plumbojarosite, malachite, azurite, opal and gypsum.

The normal depth of the zone of oxidation is as much as 50 m, but locally oxidation may extend down to 200 m.

A further break down of the three types of mineralization listed above results in six mineral associations identified at Mirgalimsai: (1) dolomite, quartz, barite, galena, pyrite, sphalerite, and chalcopyrite; (2) dolomite, calcite, galena, pyrite, and sphalerite; (3) pyrite, sphalerite, galena, and calcite; (4) ankerite, barite, quartz, galena, sphalerite, chalcopyrite, chalcocite, tennantite-freibergite, argentite, dolomite, and quartz; (5) fluorite, calcite, quartz, galena, sphalerite, and pyrite; and (6) quartz, chalcopyrite, sulvanite, Co-bearing pyrite, arsenopyrite, magnetite, and goethite.

The suggestion has been made by Yushko, (1969) that sequence (1) is syngenetic to the country rocks, (2) is developed diagenetically, (3) is of little commercial importance, and (4), (5), and (6), probably were the result of precipitation from deposition of later hot solutions of deep-seated genesis. It is equally reasonable, at least, to assume that one ore fluid, differing somewhat from one place to another in the large area of the Mirgalimsai ore field, could produce the various types of mineralization listed above without any recourse to distinct and separate ore fluids, one to produce each of the six ore types listed above. Instead, one ore fluid, developing at one time and differing only as a result moderate variations in composition and of the somewhat dissimilar reactions in which it participated in the various paths it followed in its upward journey could accomplish the production of these only slightly varied ore types.

The principal argument in favor of a syngenetic manner of genesis for the Mirgalimsai ores is the apparent lack of any igneous rock and the absence even of tuffs except in the lowest of the Fammenian beds. On the contrary, it can be put forward that the widespread faulting and brecciation within the rocks of the ore field provide channelways for upward solutions generated at depth to reach the horizon where the ores were deposited. The precipitation of the ore and gangue minerals by the replacement of the carbonate matrix, a matrix almost certainly developed in its present texture after the ore horizon had been lithified, indicates that the ore sulfides and the barite were not syngenetic or even diagenetic.

Although lead and barite are the main products of mining at Mirgalimsai, zinc probably also is recovered, though most of the zinc is concentrated in the marginal rock volumes of the ore field and probably is not as readily available for exploitation as the galena-barite-rich portions more centrally located.

The trace elements found in the ore include: silver, germanium, thallium, gallium, copper, cobalt, bismuth, and cadmium.

Cerussite or lead as a substitute for Mg or Ca in the structure of the carbonates is miniscule in amount; the cerussite, if not the lead in the carbonate molecules, very probably is the result of oxidation, although I have seen no data on cerussite distribution. It certainly would be a most unlikely primary mineral.

The belt of lead concentration exists through the central part of the deposit trends northward from center to periphery with the actual quantity of lead decreasing away from the deposit center. Lead also decreases as the ore horizon is followed upward, the Pb content changing from an average of 1.17 per cent in the lower parts of the horizon to 0.58 in the central volumes to 0.36 in the upper ones. Zinc is minimal in the central portions of the deposit. The zinc, however, is not concentrated in a belt as in the galena but in a ring-like band around the lead.

The major barium mineral is, of course, barite, the quantity of which ranges from 0 per cent to 60 to 70. Other barium minerals are witherite, barytocalcite, and barytocelesite. The thicker the ore horizon at Mirgalimsai, the greater the amounts of barite with the quantity of barite becoming less the thinner the formation becomes. As is true of lead, the amount of barite decreases upward through the ore horizon, the average of 6.4 per cent in the lower parts dropping to 3.3 in the upper; locally, reversals of these trends of metal and mineral distribution occur in certain ore segregations and separate blocks.

Zorin (1969) reports that lead in the lower reaches of the horizon has its larger amounts, relative to the rocks above it, preferentially in calcareous dolomites; the amount of galena becomes less in both the calcite- and dolomite-rich matrices. The same relationship appears to obtain as regards zinc. On the contrary, the maximum amounts of barite are present where the host rocks are most dolomitic.

The less the textures of the matrix material have been enlarged by metamorphism, the higher the content of galena; the same is true of the disseminations of galena and the fine galena-bearing veinlets. The aggregates of galena become larger the more dolomitic the matrix; the size of the dolomite crystals also increases in the more dolomitic matrices. All this suggests that the fine-grained type of matrix is more easily replaced by galena, but, where the coarser-grained matrix is replaced by PbS, the larger the crystals of it that can be developed.

Careful examinations of the results of sampling suggest that four portions of the Mirgalimsai rocks are most favorable for the emplacement of galena. The first of these is immediately below the Second Ribbon-Like formation but the most of the galena (99 per cent of the total reserves) is in three layers in the Second-Ribbon-Like itself in which the relative proportions of lead to zinc to barium are: 1:0.1:6.

The vertical zonation observed in the Mirgalimsai ore field is subject to variation because of changes in the dolomite content of the rocks, to the thickness of the ore horizon, and to the distance of the ore in question from the center of the ore body.

On balance, I think that, despite the strata-bound character of the Mirgalimsai ores and the lack of any evidence of any igneous activity to which the formation of an ore fluid might be related, the deposit probably were emplaced after the host rocks had been solidified and even after they had undergone some metamorphism. The consistent emplacement of the ore minerals by replacement of the matrix carbonates strongly supports the idea that the ores were introduced after the host rocks had been lithified. The presence of what appears to be a higher trace-element content of silver than would be expected in a telethermal deposit, indicates that the ores should be categorized as Mesothermal, and this is done here. Certainly nothing in the mineralogy of the ores at Mirgalimsai would indicate that any higher regimen han mesothermal should be included in the classification.

Were the ores considered to be syngenetic, the age definitely would be efined as late Paleozoic. If they are epigenetic, there association with he Hercynian orogeny would place them in the same time span, so they here re dated as Late Paleozoic.

RIDDER-SOKOL'NOE, KAZAKHSTAN SSR

Late Paleozoic	Zinc, Copper, Lead, Barite Pyrite, Gold	Mesothermal

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Notes

The Ridder-Sokol'noe ore field, named for the original discoverer in the mid-1700's is now probably better known as the Leninogorsk ore field; its location is 50°23'N,83°32'E and it is only about 65 km southwest of the border of Kazakhstan with the RSFSR. When the deposit was worked from 1789 and 1861, only oxidized ore were exploited; production of the underlying primary ore began in 1885. In 1916, the mine was flooded and shut down. In the early 1920's, Lenin is said to have suggested that it be re-activated; the mine, therefore, was reopened in 1928, and a lead smelter was constructed; at the same time, the name of the deposit was changed to Leninogorsk, and the same name was given to the smelter. During this period, additional exploration was carried out in the area that found further lead-zinc deposits; these, plus more recent discoveries, made it possible for operations to be continued, probably to the present time.

The most recent discovery in the ore field was the Tisha deposit in 1958 on the outskirts of Leninogorsk (town); this deposit was first mined as an open pit, starting in the mid 1960's. In 1966, a zinc refinery was opened in the district, and in 1976, underground mining was begun at Tisha and, after some problems, the Combine began to meet its production quotas.

In the district, three types of ore are mined, oxidized ore, mixed ore, and sulfide ore. In 1965, 81 per cent of the ore processed at the Leninogorsk smelter was sulfide ore, 16 per cent mixed ore, and only 3 per cent oxidized ore. The gold content of the Leninogorsk ore ranges from 0.6 to 1.0 grams per ton (with the Tisha ore at the low end of this range); 2.7 tons of gold were recovered from the Leninogorsk ores in 1977.

From Leninogorsk, a second-class railroad runs southwest to Ust-Kamenogorsk and is continued WNW to Semipalatinsk.

The host rocks of the ores of the Leninogorsk field consist of lower Paleozoic and middle Devonian volcano-sedimentary beds. These lower Paleozoic rocks well may be as old as Ordovician or even older; they form the northern border of the Leninogorsk ore field in which area they have been thrust southward over Devonian sediments. Diamond drilling also has located the lower Paleozoic beds beneath a sequence of Devonian beds within the central part of the Leninogorsk ore field. The Devonian rocks include sericite-chlorite schists, phyllites, and chlorite-carbonate schists that may have been metamorphically produced from polymict and tuffaceous sandstones, argillaceous rocks, and tuffs.

Both the lower Paleozoic beds and the middle Devonian beds that overlie them to the north of this ore field have been intruded by magmatic rocks of various ages and compositions.

In the ore field proper, the middle Devonian sediments belong to the Eifelian stage, and they have separated into four groups: (1) the Leninogorsk (400 to 600 m thick); (2) the Kryukovsk (200 to 500 m thick); (3) the Il'ina (100 to 600 m thick); and (4) the Sokol'naya (greater than 500 m thick) (Man'kov, 1969). The Kryukovsk group is made up of fine-grained siltstones, quartzites, calcareous siltstones, siliceous shales, lavas, and tuffs of albitized rhyolite porphyries. It is in these rocks that ores of the Ridder deposit are contained with the ores being located mainly in the sedimentary rocks (quartzites, microquartzites, and fine-grained siltstones). The Tisha deposit lies on both sides of the contact between Il'ina and the Sokol'naya groups. This location of these two ore bodies in two different horizons in ore bodies that are separated by only a few kilometers has an important bearing on the genesis of both as will be discussed below.

The Il'nia group is made up of extrusive pyroclastic rocks of intermediate and mafic character that contain minor members of lavas and tuffs of silicic composition. The Sokol'naya group contains a sequence of volcano-sedimentary rocks and is the last of the rocks of this type in this ore field. These rocks are principally argillites and siltstones with thin intercalations of silicic and mafic extrusives.

The intrusive igneous rocks in the ore field are: (1) quartz albitophyres; (2) porphyrites; (3) diabases, and (4) extrusive breccias.

In the southern part of the ore field, the quartz albitophyres are quite widely distributed with their layered masses being confined to the Sokol'naya group; these extrusives have a thickness from a few up to 500 m. Burov and Kurek (1939), however, considered these rocks to be intrusives and to be a facies of upper Paleozoic granitoids. Shcherba (1957), nevertheless, thought them to be extrusives of the Sokol'noe group, and Yakovlev (1959) believed that, largely because of the conversion of the Sokol'noe siltstones to hornfelses where they came in contact with the quartz albitophyres and because of the extrusive nature of their upper contact, these particular quartz albitophyres to be of a hypabyssal mode of emplacement.

The porphyrites occur as intrusive bodies along planes of stratification and as dikes that cut across the beds; the opinion of Smirnov and Pozharisky (1977) is that these last form the roots of the layered bodies of these same rocks. Single bodies of porphyrite may be as much as 50 m thick. Shcherba (1957) defined four varieties of this rock type based on slight differences in their compositions. The concensus of opinion appears to be that the porphyrites are hypabyssal bodies of Il'ina age.

The diabases are dikes that dip vertically, have thicknesses of 3 to 5 m, but can be traced on the surface for as much as 1.5 to 2 km. The diabases are the youngest rocks in the ore field (as is so often the case), including the Sokol'naya group, the quartz albitophyres, the porphyrites, and the eruptive breccias. Several investigators (Chreprasov, Pokrovskaya, Kovrigo) believe that the diabases are the youngest rocks in the ore field and also post-date the ores.

The eruptive breccias are made up of fragments of the Kryuovsk group and some of those of the Leninogorsk group; these fragments are cemented by fine particles of clastic material. They are concordant and discordant bodies included in the Kryukovsk and Il'ina groups; some of them are pipe-like. Some of the fragments in the matrix are quartz-barite, massive sulfides, and sulfides in vein quartz; some of these breccias are penetrated by quartz-carbonate-barite-galena veinlets. These relationships are thought to show that these breccias were formed during a stage in the ore-forming precess.

Although Smirnov and Gorzhevsky (1977) quote Kurek as having studied the hydrothermally altered rocks in detail, they give no reference to his work. His information, supplemented by other data, surely indicates widespread quartzification, sericitization, carbonatization, chloritization, and baritization affected the rocks of the Leninogorsk ore field. Shcherba (1957) says that these altered rocks are pre-ore with the composition of these alteration masses having been largely determined by the rocks they replaced (so what else is new). Chreprasov, and others (1978) believe that the quartz-rich content of some of the alteration products was due to their having been primary quartz-rich sediments. Chekvaidze (1971) thinks that the amount of hydrothermal alteration has been exaggerated; he also believes that what alteration there is was associated with volcanic processes. Chekvaidze does accept that certain minor alteration effects, closely associated with faults and joints, are truly hydrothermal. More on this below.

The structure of the Leninogorsk ore field is that of a graben that is bordered on two sides by faults and on the other two by fold structures. The Northern thrust forms the boundary in that direction; along this thrust, lower Paleozoic sedimentary rocks and granodiorites of the Sinyusha massif have been forced over Devonian beds. The southern boundary is the Ivanovo fault by which metamorphic slates and granitoids of lower Paleozoic age (the Ivanovo block) have been thrust over older Quaternary sediments that are unconformably above the Devonian beds of the ore field. The western boundary is provided by the Kedrovsk-Butachikha fold zone and the eastern by the Uspensk-Karelinsk zone of folds; both of these are part of the northeastern zone of folding of the Rudnyi Altai.

Within this graben, the middle Devonian sediments have been somewhat deformed and normally have a slight dip to the south. Detailed work (Burov and Kurek (1939) and Chreprasov (no citation) identified two horst-anticlinal folds, the axes of which strike NW-SE; these are the Ridder-Sokol'noe and Kryukovsk. These fold structures have been further complicated by a graben-anticline that separated them. The horst-anticline structures have been additionally confused by several, more or less equi-dimensional domes in the cores of which are quartz-barite rocks; the roofs are composed of Kryukovsk siltstones. The control that these structures exert over the locations of the ore deposits is shown by the Ridder-Sokol'noe deposit being contained in the anticline of that name with the various ore segregations of that deposit being located in the domes.

How these domes were formed is a matter in dispute. They may be ordinary folded structures or volcanically formed domes or the result of block movements in the basement of lower Paleozoic rocks, or produced by comtemporaneous tectonic and thermal activity. For this distance, with the data given, it is impossible to judge among these concepts.

The Ridder-Sokol'noe ore field contains five major faults; they complicate the structures of the two anticlines mentioned above and are responsible for changing them into horst-anticlines. Thus, the faults also are important in controlling the locations of the ore bodies. These faults strike NNW-SSE; the dips range between west and east, and the throws are up to several meters. These faults are thought first to have been developed during sedimentation but to have been reactivated at several later times. The ore field also contains a large number of minor fractures that probably connect to the major faults and appear to contribute to localizing the ores.

The Ridder-Sokol'noe deposit is composed of several ore segregations of which the main ones are: (1) Central, (2) 1st, (3) 2nd, (4) 3rd Southwestern, and (5) Bystrusha; all of them are quite complex in form. The Kryukovsk body is on the far east of the ore field and appears to be far more strata-bound than the others (with no genetic implication intended). These ore bodies normally are more or less concordant with the rocks that overlie them, but in their lower parts they appear to be quite discordant with apophyses extending downward so that the vertical range of the ore zone is more than 600 m.

The various Ridder-Sokol'noe authors identify three types of ore: (1) massive; (2) segregated (disseminated), and (3) veinlet (stockwork). The massive and disseminated ores are located in the limbs and bases of the domes (although the diagrams given suggests that the ores are on the limbs and crests). The ore bodies normally are lens-shaped. The stockwork (veinlet) ores are considered to be of two types - mesh and subparallel - the former have been localized on sheets of joints, the latter lie beneath the meshed or disseminated types and pass gradually into the type above them.

On a chemical basis, three ore types have been defined: (1) lead-zinc, (2) copper-zinc, and (3) gold-bearing. The major ore type is the lead-zinc; the ratio of lead to zinc to copper in these is 1:2.1:0.2; in the Cu-Zn-Pb ores these ratios are: 1:4.3:8.1.

The structures of the ores are (Isakovich, 1970) massive, banded, veinlet, secretion, cockade, segregated, and streaky, plus other less important types. The textures are subhedral and subgraphic.

The major minerals are sphalerite, pyrite, galena, chalcopyrite, quartz, calcite, hydromica, phengite (var. of muscovite), dolomite, and barite. Minor minerals are: tetrahedrite, marcasite, arsenopyrite, silver, chlorite, epidote, albite, ankerite, kaolinite, magnesite, siderite, chalcedony.

The number of stages of mineralization differs from one author to another. One suggestion is that there are two stages: (1) hydrothermal-sedimentary in which bedded pyritic-polymetallic ores were developed in which three phases were recognized by Pokrovskaya and Kovrigo (1969); these are: sulfo-pyritic, zinc-copper, and quartz-barite-polymetallic; and (2) hydrothermal-metamorphic, composed of two phases, sulfide-sericite and late veinlet.

On the contrary, Isakovich and Gibsher (1971) believe that there were four stages: (1) pyritic (pre-ore); (2) galena-chalcopyrite-sphalerite; (3) galena-sphalerite-barite; and (4) quartz-carbonate (post-ore). They consider that stage (2) was formed between 325° and 125°C and stage (3) between 345° and 145°C. Temperatures for stages (1) and (4) were not established.

The Ridder-Sokol'noe ores are vertically zoned with barite-polymetallic ores on the upper levels, lead-zinc ores in the central portions, and copper-zinc in the lower parts.

Isotopic studies on the leads from the Ridder-Sokol'noe and Tisha deposits indicates that the leads in the ores and in the country rocks are very much the same, so that the ores in both the ore bodies and the country rocks came from the same source. Whether this means that the ore leads were derived from the country rocks or that minor quantities of ore leads were disseminated into the country rocks from outward moving ore solutions cannot be determined with the data at hand.

I cannot believe, after seeing a cross section (from east to west) through the Ridder-Sokol'noe ore body and reading the information about the degree of wall-rock alteration (typical of mesothermal deposits) and of the intensity of metamorphism (low) that the ores are other than ones deposited from hydrothermal solutions after the rocks containing them had been lithified and deformed. The relations of the ore bodies to structures seem to me to insist that the ores were emplaced only after the structures had come into existence and, therefore, long after the host rocks had been lithified. The widespread nature of the various types of alteration has been used as an argument that this type of wall-rock alteration was pre-ore; if this is true, it simply puts the ore farther away in time from the original sedimentation and confirms that epigenetic emplacement of the ores. I have, therefore, classified these ores as Mesothermal.

The age of the ores must be at least late Devonian and more probably is that of the development of the Rudnyi Altai mountains or Hercynian and, therefore, Late Paleozoic.

Space is not available for a thorough discussion of the Tisha deposit that is in the same ore field (Leninogorsk) as the Ridder-Sokol'noe ore bodies, the Tisha being between the Ridder-Sokol'noe and the city of Leninogorsk. The Devonian rocks that contain the Tisha deposit are the same four groups of the Eifelian stage in which the Ridder-Sokol'noe deposit lies. The Tisha ores, however, are appreciably higher in the stratigraphic sequence, being contained in the uppermost Il'ina series and the lowermost Sokol'noe whereas Ridder-Sokol'noe ore are contained throughout the Kryukovsk. Thus a vertical stratigraphic distance of at least 600 meters separated the two ore bodies.

The host rocks of the Tisha deposit have been much more intensely metamorphosed than those of Ridder-Sokol'noe. The Tisha deposit is contained in the Kedrovo-Butachikha fold-fault zone that marks the western boundary of the Ridder-Sokol'noe deposit. The Il'ina and Sokol'noe beds that contain the Tiska deposit have been cut by bodies of rhyolitic porphyries that compose the Poznopalovsk intrusion and several small stocks and dikes. The close spatial and temporal relations of these intrusives to the Tisha deposit (the Tisha ores are in the eastern contact zone of the intrusive where it fingers out) indicate that they probably also are genetically related, although workers on the deposit hold a variety of other opinions about the meaning of the relationships of the porphyries and the Tisha ores. The most reasonable view, to me, is that the porphyries are a facies of the late Paleozoic Zmienogorsk complex and that the ore fluids that developed the ore came from the same general source as the magmas of the complex. These intrusions probably also aided in, if they did not entirely cause, the metamorphism of the host rocks of the Tisha ores. The principal structural control in the Tisha area is provided by two east-west faults that bound the ore on the north and the south, the Mar'insk and the Glazkovsk, respectively. These faults are overthrusts that dip north (Mar'insk) and south (Galzkovsk); the eastern and western boundaries of the Tisha area are furnished by the Tisha (3) and Sokol'nyi (W) faults that strike more or less north-south. Several smaller faults complicate the structure of the Tisha ore-bearing block, having feathered out from the Mar'insk fault and consisting of crush zones around the rhyolite porphyries. The Tisha block contains three zones of major shearing, the strike of which is more or less east west, more or less the same as those of the rocks of the Il'ina and Sokol'naya groups. The shears are from 40 to 150 m wide. Some NNW-SSE-striking shear zones also are present and dip steeply to the east. To this structural pattern have been added a series of folds with bowl- or box-like-shapes, the axial planes of which strike east-west to NE-SW. Still further structural complications are caused by flexures and drag folds.

It is suggested that the Central belt of shearing that is considered to be a closely spaced series of shearing joints feathering out from the Longitudinal fault that cuts through the center of the zone of shearing. The shearing zone surrounds the main ore bodies of the deposit and is confined to the contact zone between the Il'ina and Sokol'naya groups.

Of these ore bodies, the Main segregation is the largest and is composed of three ore shoots; it is located at the intersection of the Longitudinal fault and a synclinal fold. One small ore body is controlled by the Northern shear zone (the Tisha deposit).

The Tisha ores consist of segregated and uniform types. The segregated ores are thin veinlets and small lenses of sulfides, as much as 2 to 5 cm thick that are oriented parallel to the shearing of the hydrothermally altered rocks. The uniform ore form 25 per cent of the total ore volume.

The ores have been emplaced (Pokrovskaya, and others, 1977) in three stages: (1) pre-ore hydrothermal wall-rock alteration; (2) sulfide stage: (a) sulfo-pyritic, (b) chalcopyritic, and (c) polymetallic; (3) metamorphogenic-hydrothermal, post-ore stage, quartz-dolomite veinlets with some sulfides. The temperature of formation of the Tisha ores was between 355° and 120°C with the ore minerals being concentrated in the higher-temperature ranges. As is usual, zinc increases with depth and lead decreases.

The principal minerals of the Tisha ores are: pyrite, sphalerite, galena, chalcopyrite, quartz, sericite, chlorite, and dolomite, minor minerals are: tetrahedrite-tennantiem albite, K-feldspar, and calcite. Gold is less abundant than it is in the Ridder-Sokol'noe ores proper.

The ores are similar enough to those of Ridder-Skol'noe itself to be classified as Mesothermal; they are definitely of a slightly higher temperature, both in ore minerals and wall-rock alteration than those of Ridder, but there is no indication that "hypothermal" should be added to the classification. The age almost certainly is Late Paleozoic.

How anyone who has studied both Tisha and Ridder-Sokol'noe could think that one (R-S) was syngenetic-volcanogenic, at least in large part, and think that the other (T) was hydrothermal, I cannot see. They are too closely associated in time and space, are too similar in mineralogy and structure for anyone to consider them to have been formed one in one way and one in another.

ZHAIREM, KAZAKHSTAN SSR

Middle and Late Paleozoic	Barite, Zinc, Lead, Iron, Manganese	Sedimentary-A3 (Fe,Mn) Mesothermal to (Zn,Pb,Bar)

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Notes

The deposit at Zhairem is located at 48°20'N,70°30'E in west-central Kazakhstan, near the tiny village of Karazhal (48°00'N,70°55'E). Karazhal is connected to Karaganda (a major city at 49°53'N,73°07'E) by a secondary railroad. Karaganda is about 200 km NE of Karazhal.

The deposit was discovered in 1951 through the measurement of magnetic anomalies through the several meters to 10's of meters of Cenozoic sands, loams, and clays that covered the Paleozoic bed rocks. The anomaly was generated by a previously known iron-manganese mineralization; apparently as more or less of a speculation, this anomaly was drilled and a barite body and rich lead-zinc ore were found.

The ore mineralization at Zhairem is contained among the siliceous-carbonate rocks of late upper Devonian age. The rocks have been incorporated in the Zhailma synclinal structure. The rocks involved in this syncline are divided into two rock assemblages; these are: (1) sedimentary-volcanic rocks of middle and upper Devonian (Frasnian) age, composed of lavas and pyroclastics of silicic, mafic, and alkaline types; these alternate with sedimentary rocks; and (2) marine sediments that include carbonates, clay-carbonates, and sandy argillites of Famennian (uppermost Devonian), Tournaisian (lowest Carboniferous), Visean, and Namurian (lower and upper Carboniferous). Within the Zhairem ore body, only assemblage (2) has been exposed. Thus, the oldest rocks so far encountered in the mining process are lower Famennian that are made up of argillaceous limestones with intercalations of limestones and siliceous carbonate rocks; these beds are more than 150 m thick. In the upper Famennian, three substages are recognized; these are the productive, the intermediate, and the redbed. The principal rock types in this sequence are carbonaceous-clay-silica-carbonate sediments. The productive (the lowest member) of the upper Famennian includes all the ore bodies of the Zhairem deposit; both the thickness and the stratigraphic sequence of these beds are quite varied. The Western ore sector (the other two are the Far Western and Eastern). The Western ore sector appears to have been most and best studied, and this sector includes (from the bottom upward): (1) thin-bedded argillaceous-siliceous-carbonate rocks, as much as 90 m; (2) unevenly bedded argillaceous-siliceous-calcareous rocks, 70 m thick; (3) rhythmically arranged pyritic beds that are composed of argillaceous-siliceous-carbonate beds with rhythmic banding and including many seams of finely banded pyritic rocks, 30 to 50 m thick; (4) carbonaceous-argillaceous-carbonate-siliceous bedded rocks that contain seams of tuffs and tuffites, plus horizons of banded pyritic ores, 10 to 15 m thick; (5) lower flyschoid horizon that is made up of argillaceous-siliceous-carbonate rocks with carbonaceous material in pigment amounts, the upper parts of the rhythms contain seamlets of finely dispersed pyrite and sphalerite, 50 to 80 m thick; (6) horizon of carbonaceous-quartz-K-feldspar rocks containing carbonate nodules - an excellent horizon marker due to its unusual color and composition, 10-14 m thick; (7) middle flyschoid horizon, the same as the lower flyschoid, 10 to 15 m thick; (8) horizon of bedded iron ores (magnetite, hematite, ferruginous jaspers that alternate with seamlets of siliceous limestones), uncertain thickness. Above the productive member is the intermediate member that is composed of an alternation of rhythmically banded and nodular-banded limestones, 60 to 80 m thick (Smirnov and Gorzhevsky, 1977).

The uppermost member (the redbed) is mainly red limestone that contains an alternation of layers of red rocks - limestones, siliceous limestones, fine mudstones made up of hematite, hydromica, sericite, and chlorite-sericite; this member has a thickness of 80 to 100 m.

The rocks of the Carboniferous (all barren) include two horizons in the Tournaisian (lower and upper): (1) nodular bedded limestones with intercalations of argillaceous-sericite mudstones, 30 to 50 m thick and (2) carbonaceous limestones, that is, siliceous and argillaceous-siliceous limestones with much organic pigment, 50 to 80 m thick. In the upper Tournaisian, the three horizons are: (1) psammitic vitroclastic tuffs, tuff-mudstones, and sedimentary breccias, 5 to 30 m thick; (2) 'concretionary' limestones, that is, siliceous limestones and carbonate-silicia rocks that contain bedded compacted carbonate concretions (nodules), 40 to 50 m thick; (3) ash-gray limestones, made up of siliceous limestones and siliceous-carbonate rocks with massive and coarsely bedded structures and seams of argillaceous-sericite mudstones and calcareous sandstones, 350 m thick.

The Zhairem deposit is a representative of the Atasu type which means a bedded deposit which is a combination of layered iron-manganese and zinc ores in a distinct ore body on which hydrothermal barite-zinc-lead mineralization had been superimposed (Shcherba, 1980). This type of mineralization is not uncommon in the late Famennian of Soviet Asia and such are the rocks that are in the Zhail'ma* synclinal structure of the Zhairem deposit. (*variant spelling)

In discussing ore deposits of the Atasu type, Shcherba (1980) says that concentric zoning is typical of Atasu deposits. In general, where hydrothermal-metasomatic barite-polymetallic ores are confined to volcanic centers, the outward distribution through sedimentary, polymetallic, iron-manganese, and iron ores forms a concentric area with a radius of 7 to 14 km. At

Zhairmen in particular, such concentric zoning can be observed from central siliceous barite to lead-barite to barite-lead-zinc to bedded lead-zinc to bedded pyrite-sphalerite. Where iron-manganese is included in the depositional process, the zoning of these metals also is concentric. The exact form of the ore bodies, however, depends on the nature of the fractures (isometric or elongated) from which the metal-bearing solutions are discharged. In Zhairem, for example, where both iron-manganese horizons and bedded lead-zinc deposits are superimposed, the pattern of mineralization and the explanation for it become more complex. Shcherba points out that these patterns depend not only on the place of discharge of the ore-bearing solutions but also on the attitude of the strata they invade, their temperatures, their Eh and pH, and their concentration. If both Fe and Mn and Pb-Zn-Ba are carried in the same solution then three results may obtain: (1) if Mn is rich, then Pb and Zn are low; (2) if Fe plus Mn is rich, then the Pb-Zn-Cu content is moderate; (3) if Fe plus Mn is low, then Zn-Pb content is high.

In the Zhairem deposit (Smirnov and Gorzhevsky, 1977), the development of near-ore alteration is relatively restricted. In the periphery of the ore area, siliceous-carbonate alteration is present; in the intermediate zone siliceous replacement rocks with albite and dolomite occur; in the central portion of the field, pyrite-quartz and barite replacement rocks were produced; these rocks are the most widely distributed.

In contradiction, it has been argued that the siliceous rocks in the Zhairem ore field are not replacements but had a volcano-sedimentary manner of genesis. The rocks so produced were albitized and quartzified and, to a lesser degree, dolomitized, sericitized, and chloritized. Only the first three of these alteration types were developed during the period of the deposition of the ores.

The quantity and variety of magmatic rocks in the Zhairem field are limited. Such rocks include: hypabyssal bodies of trachyte porphyries (located in the footwall of the productive member), plus dikes of amydaloidal porphyries, present principally in the rocks of the Far-Western ore sector. Explosive breccias, on the contrary, were much widerspread; these rocks are made up of clastic materials with the proportion of fragments to matrix differing over a wide range. The fragments include not only all types of country rocks but also pieces of lead-zinc and barite ores. The matrix material is composed of quartz, dolomite, and albite. Such workers as Degtyareva, Mitryaeva (1967), and Rozhnov (1972) consider these rocks to be explosive breccias, the event that produced them being closely connected with the hydrothermal mineralization; they are, therefore, intra-ore.

Rozhnov (1972), describes the major structural feature of the Zhairem field as a brachyanticline (the Zhairem) that is a complicating feature on the Zhairem syncline. The upper part of the core of this anticline is composed of lower Famennian sedimentary rocks, and these beds have been penetrated by hypabyssal trachyte porphyries that make up most of the lower core of that structure. In contrast to the crest of the anticline, the limbs consist of late Famennian and early Tournaisian beds. The axis of the brachyanticline strikes more or less NE-SW (as does the syncline). On the northeast and southwest, the anticline is cut off by hidden faults with a more or less east-west-striking axis. In the Western and Far-Western ore sectors, other faults with more or less north-south and NE-SW strikes are known.

The varied structure of the Zhairem ore bodies resulted from their different manners of genesis and their relationships with the country rocks. Rozhnov (1977) believes that the mineralization was so emplaced as to result in the following six forms: (1) units of seamlets (from 0.1 to 15 mm wide) that contained finely distributed framboidal pyrite with which are interstratified seamlets of quartz-calcite with finely disseminated sphalerite; such rhythmic units are from 1 to 100 cm in thickness. In places, the sphalerite is present in the seamlets instead of pyrite to form sphaleritic instead of pyritic rhythmites; (2) calcite and quartz-calcite-barite veins containing segregations and nests of sphalerite and galena; these veins are conformable with the bedding and are from 1.5 to as much as 10 cm thick; (3) systems of veinlets of quartz-calcite that cut across the bedding; where these veins cut pyritic rhythmites, nests and segregations of coarsely crystalline galena and sphalerite are present; (4) primary lenses in the country rocks that have been replaced by fine-grained, gray barite that contains disseminations of galena and sphalerite, and, in places, chalcopyrite; such lenses may be 100 to 120 cm thick; (5) nests, veins, and veinlets of coarsely crystalline white barite contained in fine-grained gray barite; and (6) beds composed of various alternations of hematite, magnetite, ferruginous jaspers, and siliceous (in some places hematitized) limestone; such beds may be as much as 300 m thick or as thin as 0.5 m.

Essentially all ore geologists who have worked in the Zhairem field are convinced that beds of types (1) and (6) were formed at the same time as the country rocks in which they are contained; actually type (6) is the country rock and, in some places, so is type (1). The stratification of these beds is quite definite and no alteration selvages are present around the beds. Locally beds of types (1) and (6) contain the mineralized remains of radiolarians and ostracods and were involved in the folding and faulting of the enclosing rocks. The later types (2) through (5) are believed to have been produced by a later epigenetic stage and were deposited by replacement of the host rocks and by the filling of vein-fractures. Most of the ore of the epigenetic types are contained in the upper Famennian "productive" member, the lowest of the three members of the upper Famennian. Some barite ore bodies, however, have been found in the sediments both below and above the "productive" member.

The two types of syngenetic ores are in layers with a thickness that ranges between 5 and 25 m; these ore bodies are much longer than they are thick. On the contrary, the epigenetic ores are mainly in or near the crush zones, and they are especially well-developed in folded rocks where various beds were separated from their neighbors during folding. Despite the probable epigenetic genesis of these ore types, they are lens-like in shape and normally conform to the country-rock stratification.

As is clear from the description on the various ore types in the third paragraph above this one, the syngenetic ores are of two very different types. Form (6) contains iron and iron-manganese ores and form (1) pyrite-sphalerite ores. The epigenetic ores are classified as zinc-barite, lead-barite, lead-zinc-barite, and barite types with these ores being present in forms (2) through (5); apparently any one of the ore types can be present in any form appropriate to syngenetic or epigenetic ore, as the case may be.

The Zhairem ores are poor in minor and trace minerals and elements. The syngenetic ores contain traces of silver, cadmium, indium, bismuth, nickel, arsenic, gallium, thallium, and germanium, whereas, in the epigenetic ores, mercury, antimony, and cobalt have been noted but absent are arsenic, thallium, and germanium.

The textures of the syngenetic ores are normally bedded or have segregated fabrics, and pyrite is present in globular form. Colloform, zoned, radial, fibrous textures are less common; some cement-like textures were produced where sphalerite filled spaced between pyrite globules.

Textures of the epigenetic ores include a major development of the banded variety that were produced by replacement processes; appreciable breccia textures also were formed by crushing of the early minerals and their cementation by later minerals. The most abundant and most broadly produced minerals are: pyrite, sphalerite, galena, chalcopyrite, hematite, magnetite, marccasite, tetrahedrite-tennantite, bournonite, millerite, bravoite, jamesonite, boulangerite, quartz, barite, calcite, dolomite, albite, fluorite, K-feldspar, and sericite.

Mitryaeva, and others (1968) has recognized the following stages and phases.

(1) the sedimentary stage of two phases (a) deposition of sediments and (b) diagenetic phase in which the bedded pyrite-sphalerite ores were developed;

(2) the hydrothermal-replacement stage, the phases of which were: (a) albite-dolomite-quartz; (b) quartz-pyrite; (c) quartz-carbonate-sulfide-barite;

(3) late hydrothermal stage in which a quartz-calcite-pyrite assemblage with chalcopyrite and hematite was formed;

(4) the supergene stage.

An impressive horizontal zoning has been observed in the Zhairem ore sectors. In the centers of the sectors, the maximum development of hydrothermal-replacement of the host rocks took place; here only pyrite, chalcopyrite, and barite, plus quartz, were emplaced. This inner zone, designated as zone (1), is surrounded by several zones, successively farther from the center; these are: (2) galena-barite replacement, (3) galena-sphalerite-barite replacement, (4) rich lead-zinc ores with barite, (5) layered lead-zinc ores that locally are accompanied by barite, and (6) layered, almost completely zinc ore with some lean lead added.

Mitryaeva, and others (1967) suggest that the zones (5) and (6) contain the syngenetic ores in great preponderance. In the first four zones, the ores were formed mainly by replacement reactions generated by hydrothermal ore fluids.

It appears more probable that the layered sedimentary iron-manganese and iron ores and the layered sedimentary zinc ores were formed from volcanic exhalations that entered the sea and penetrated the sediments undergoing diagensis to produce the layered zinc ores. The ores in zones (2) through (5) were emplaced by the replacement reactions induced by hydrothermal fluids after the country rocks had been lithified, folded, and faulted. The exact time and space relations of the iron and iron-manganese ores to the sedimentary zinc ores are not clear but there appears to have been little overlap between the formation of these two ore types.

As for the replacement ores, it appears that they were formed under mesothermal conditions with some indication that minor amounts of the minerals in the hydrothermal deposits may have been formed under higher- or lower intensity conditions than mesothermal but not enough to justify adding either hypothermal or leptothermal to the classification.

The sedimentary zinc ores, formed during diagnesis, should be classified as Sedimentary-A4 (diagenesis).

The supergene ores appear to have been of minor importance, but they should be classified as - by ground water circulation, that is, IIB2. The age of the supergene ores is, of course, recent.

The age of the deposits almost certainly is late Paleozoic, if the uppermost Famennian can so be categorized.

KADAMDZHAI, KIRGIZ SSR

Late Paleozoic (early Mesozoic?)	Antimony	Epithermal

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Notes

Both the Kadamdzhai and Khaidarkan deposits are contained in the South Fergana antimony and mercury belt with the former being essentially an antimony deposit and the latter one of mercury. The Kadamdzhai deposit is located at 40°07'N,71°44'E and that of Khaidarkan at 39°57'N,71°20'E; the distance between them is about 45 km in a NE-SW direction. The similarity between the two deposits is remarkable in structure and stratigraphy, and the temperature and pressure obtaining during ore emplacement is quite close, even though the principal mineral in the one is different from that in the other.

The major difference between the two deposits is that Kadamdzhai is principally mineralized with stibnite and Khaidarkan with cinnabar. The Kadamdzhai deposit is included within the Kadamdzhai-Chauvay section of the South-Fergana mineralized belt and was discovered in 1914. It is contained within the westerly plunging portion of the Aktash anticline. The east-west length of the field that contains the Kadamdzhai deposit is about 20 km and the north-south width about 1.5 km. The various ore bodies, of which the Kadamdzhai is the most important, are scattered at irregular intervals along this length with mineralization differing appreciably from one portion of the belt to another.

The rocks of the ore-bearing areas consist of Silurian, Devonian, and Carboniferous rocks, and these locally are buried under unconsolidated Quaternary deposits, these latter having been emplaced long after the stibnite mineralization that probably is Permian in age. The middle Silurian to lower Devonian rocks are largely composed of mechanical sediments whereas the upper Devonian to uppermost Carboniferous rocks are mainly carbonates. The only igneous rocks are uncommon dikes of diabase porphyries.

The structures now exhibited by these rocks are highly complex with an allochtonous sheet of the Silurian and Devonian beds being overthrust onto the younger beds. The rocks of this Kadamdzhai overthrust sheet have been highly deformed into more or less east-west striking folds that form the western end of the Aktask-Boordin monocline. The main ore-controlling structure is the Aktask anticline; this structure has been cut by high-angle reverse faults; these strike between NE-SW and E-W, and their dips are nearly vertical to highly inclined to the north or south. Diagonal faults branch off from the high-angle reverse faults with the strike-slip of the east-west faults being about 1500 m and that on the diagonal faults about 200 m. The ores were developed in jasperoid-hornfels breccias that were formed during the overthrusting and into which the ore-forming fluids entered along nearly vertical reverse faults and spread out laterally into the breccia bodies. The ore is concentrated where the overthrust fault plane domes upward and cuts through Carboniferous rocks (either pure carbonates or shaly carbonates). Other ores were accumulated in highly porous portions of the layered Carboniferous limestones. The uppermost ore is found from 60 to 80 m above the lowermost.

The mineralization can be either essentially monomineralic stibnite in the carbonates or stibnite-fluorite mineralization in the terrigenous-carbonate rocks. The intensity of the mineralization (and of the hydrothermal alteration that accompanies it) diminishes appreciably from west to east and from south to north. Of all the deposits in the belt mineralized in stibnite that of Kadamdzhai is the richest, followed by that of the Levoberzheny; in both instances carbonate rocks occur both above and below the fault plane of the major overthrust. Another relatively rich deposit, North Aktask, has host rocks that are autochthonous-terrigenous-carbonate and the ores carries considerable fluorite.

The various ore deposits of the Kadamdzhai portion of the South-Fergana mineralized belt are arranged like beads on a string that strikes in an ENE direction; the band is, as was mentioned above, about 1.5 km in width. Because the plane of the ore-localizing overthrust plunges down with increasing steepness to the west, the depth of the ore becomes greater in that direction. Thus, to the east and northeast, the plane of the overthrust gradually rises and passes thereby from the carbonate to the terrigenous-carbonate part of the Carboniferous section. Further, in that direction, the fault plane is confined almost entirely to terrigenous-carbonate rocks.

The stibnite mineralization is most abundant in the carbonate rocks of the Carboniferous beds over which the Silurian-Devonian formations have been overthrust; this mineralization decreases as the ore enters the overlying terrestrial-carbonate formations of the Carboniferous and disappears in the entirely terrestrial beds that are the youngest rocks of the Carboniferous. Such cinnabar as is present in the Kadamdzhai ores is most abundant the higher the mineralization is in the Carboniferous sequence. It does not appear that the cinnabar is recovered in the processing of the Kadamdzhai ores, although this is not certain.

The bulk of the Kadamdzhai ores is mainly stibnite with minor amounts of such arsenic minerals as orpiment [As_3S_3], realgar [AsS], and getchellite [$AsSbS_3$]; the gangue minerals are: quartz, fluorite, and calcite.

The location of the stibnite and its associated minerals is controlled by both structures and stratigraphy, the ore-bearing breccias being best developed where carbonate rocks are in anticlinal folds and the rocks are cut by a complex group of NE- and E-W-trending faults, the southern sides of which have moved up. The principal development of ore is localized around the junction of the NE-trending North Aktash and the E-W-trending Northern faults and extending south to the Western-Field Fault (also N-E trending.)

The ores are surrounded by zones of hydrothermal alteration, so the ores can be said to be confined to such areas, although the mineralization and the alteration probably were both part of the same hydrothermal mineralizing process. The rocks of the alteration haloes consist of jasperoids, hornfelses, and calcite, all developed in brecciated rocks near the faults just mentioned. The jasperoids were produced by the replacement of calcite by quartz to give a light-gray to gray rock that breaks with a shinning fracture. Ninety-five per cent of the Kadamdzhai jasperoids are breccias composed of angular fragments of quartz and stibnite that retain the sharp outlines of the carbonate fragments that they replaced.

The shales and sandstones of the brecciated detrital-carbonate rocks appear not to have been appreciably metasomatized but the carbonate fragments they contain have been replaced by quartz to produce a hornfels that is a mixture of silicified shale and sandstone fragments that are dark gray to black in color. Hornfels breccias, in which the jasperoid fragments are dominant over those of the silicified detrital rocks are cemented mainly by quartz, stibnite, and fluorite.

The stibnite and stibnite-fluorite ores contained in the hydrothermally altered rocks take the form of sheets, veins, lenses, and impregnations. The sheet- and lens-like bodies are concentrated in the zone of the major Kadamdzhai overthrust, whereas the veins follow the fault planes of the steep reverse faults. The veins cut through the autochthonous limestones, and the sheet-like bodies follow the contacts between terrigenous and carbonate rocks. Where the steep reverse fault planes enter the detrital (terrigenous) rocks above the overthrust, they may be mineralized for as much as 80 m vertically. Mineralized veins also extend downward into the carbonate rocks beneath the Kadamdzhai overthrust along the reverse faults for usually as much as 80 to 100 m.

In addition to the 15 known stibnite deposits that are found along the 20 km length of the stibnite mineralized belt, the belt also contains some 20 occurrences of mercury; how many of these, if any, are mineable, is not known.

The combination of the steep Hercynian reverse faults and the uplifted limbs of the folds on the south sides of these faults produced a blocky arrangement of the sedimentary rocks. The major amount of brecciation has taken place in the northern part of the closure of the periclinal portion of the Aktash anticline and in the tectonic wedge, produced by the North Aktask reverse fault, the east-west normal Northern fault, and the Western Field fault, that parallels the North Aktask fault to the south. Zharikov (1977) considers that the east-west Northern fault was the channelway through which the major portion of the ore fluids (at least) reached an area in which the Kadamdzhai ores are mainly concentrated. Along this fault, stibnite mineralization can be followed for more than 2 km; the depth of mineralization along this fault is as much as 600 m. Some mineralization is contained in layers of hornfels breccia that occur in the shales that overlie the limestones; similar layers of mineralization can be found in the limestone near the plane of the overthrust Kadamdzhai fault.

The clasts of which the breccias are composed (principally silicified shales and limestones) range from 2 to 5 mm up to 2 cm are, as has been mentioned, cemented by quartz and stibnite. Where the base of such a hornfels breccia overlies broken limestone, the limestone fragments normally are cemented more by calcite than by quartz. The maximum thickness of these breccias is attained over the major anticlines with the thickness becoming less as the limbs are followed downward; the breccias are completely lacking in the synclinal troughs and on the southern limb of the anticline.

The production of breccia was particularly intense in the vicinity of the arch and upper part of the northern limb of the Sur'myanyi ridge syncline where the nearness of the Northern fault and the deformations along it caused the dip of the northern fold to be nearly vertical or even reversed and resulted in a major crushing along the sharply folded crest of the fold, producing breccias upwards of 25 to 30 m thick.

The distribution of the stibnite mineralization, however, is far from even with the strongest stibnite mineralization occurring along the crests of the anticlines where the tightly folded rocks are further opened up by the numerous faults. The resulting complex structure insures that the mineralizations in them are equally complex. Where the folding is less sharp and the crushing of the hornfelses and jasperoids rocks much less, the mineralization is appreciably less rich.

In addition to stibnite (and to the arsenic minerals and the minor cinnabar on the outer reaches of the mineralized rock volume) pyrite, marcasite, chalcopyrite, sphalerite, and galena have been noted but are exceedingly rare. Calcite, barite, dickite, and sericite are present in minor amounts. Gold and silver have been reported.

The ores are present in a variety of types - massive, brecciated, segregated (disseminated), banded, veinlets, comb-like, cockades, and drusy.

Where the ores have, by erosion, been brought near the surface, secondary minerals - kermesite [Sb_2S_2O], valentinite [Sb_2O_3], senarmontite [also Sb_2O_3], hydrocervantite [probably $Sb^3Sb^5O_4 \cdot nH_2O$], stibiconite [$Sb^3Sb^5O_6(OH)$], and hydromeite [?]. Where karst cavities have been developed, aragonite is present.

The ores were formed in at least three stages, which were separated by tectonic events. The first stage was the replacement of the rocks in the overthrust zone by fine-grained and cryptocrystalline quartz with hornfelses being developed from the shale fragements and jasperoids from the mostly thin-bedded limestones. During this stage of silicification, sericite and pyrite, chalcopyrite, galena, and sphalerite were introduced as fine disseminations.

The second stage was the further crushing of the breccias and their cementation by quartz and stibnite and lesser fluorite and calcite.

The third (and last hypogene) stage was the development of calcite-realgar-orpiment veinlets, the lowest-temperature phase of ore formation.

Zharikov (1977) believes that the ores were formed under near surface conditions as is evidenced by abundance of breccia bodies and of metacolloidal structures among the ores. On this basis, the ores are epithermal rather than telethermal.

V. Fedorchuk (1964?) has been reported by Zharikov (1977) to have obtained an absolute-age determination on Kadamdzhai sericite of about 230 m.y., which probably places the formation of the deposit in the latest Paleozoic. Zharikov, however, says that a later date for ore deposition is possible because of the widespread tectonic activity that took place in the area in Mesozoic to Cenozoic time.

The ores at Kadamdzhai probably were formed under somewhat less intense conditions than those at Khaidarkan because the central portion of the Khaidakan ores contains appreciable stibnite whereas the cinnabar at Kadamdzhai is peripheral to the stibnite-rich bodies. Both, however, probably were produced within the Epithermal range.

KHAIDARKAN, KIRGIZ SSR

Late Paleozoic (early Tertiery?) Mercury, Anitmony Epithermal

Arapetin, S. A., 1974, On the role of large gaps in the emplacement of mercury minerlization in the deposit of Kara-Aracha (Khaidarkan): Geology of Ore Deposits, no. 5, p.

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Suerkulov, E. A., 1973, Primary reaction rims from the diffusion of mercury in deposits according to type (part of the Zavod-Khaidarkan ore field): Geology of Ore Deposits, v. 15, no. 1, p. 118-121

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Zharikov, M. G., 1961, A new listwänite type of mercury mineralization in southern Fergana: Textbook on West-Central Asia, Inst. Geol. Geophys. Mineral., Syr'ya Press, v. 6, p. 103-113

Notes

As is true of the Kadamdzhai antimony deposit, the mercury ores of Khaidarkan are located in the South-Fergana antimony and mercury belt. The Khaidarkan deposit is located at 39°57'N,71°20'E and that of Kadamdzhai is at 40°07'N,71°44'E. The straight-line distance from Kadamdzhai to Khaidarkan is about 45 km and the direction is about NE-SW.

The Khaidarkan mercury deposit is only one of a number of deposits arranged in a more or less east-west direction from Metingbel Pass on the east to the Bogashin mine on the west. The Khaidarkan deposit was discovered in 1926 and was first put into production in 1941, probably this opening of the mine was directly connected with the need for mercury in World War II.

Much exploration has been carried out in the district and has resulted in the discovery of several ore deposits, each being worked as a separate mine. These deposits are of the relatively simple quartz-cinnabar, fluorite-realgar type of mineralization is jasperoid, the type so abundantly represented in the South-Fergana ore belt. Three rather different stratigraphic sections have been identified in the Khaidarkan district; these are: (1) terrigenous (shale-sandstone) - dolomite-limestone - the Alay type; (2) terrigenous (shale-sandstone) - the Chauvay type, and (3) limestone - the Aktask type.

The Alay section contains thick Tournaisian (lower Carboniferous - lower Mississippian) limestones and dolomite and massive Visean (lower Carboniferous = lowermost upper Mississippian) limestones. Above these carbonate rocks are fusulinid limestones (their upper part being ore-bearing) and sandy shales, both of the Moscovian (upper Carboniferous). To the north, the lower half of this section contains much limestone conglomerate and clastic limestones; these also are ore-bearing.

In the terrigenous section, the lower part is made up of Silurian carbonaceous shales and Devonian shales and siltstones. The lower Carboniferous rocks in this section are as much as 150 m of clastic and detrital limestones that contain intercalations of chert and range in age from Visean to Bashkirian (middle Carboniferous, below Moscovian). Above these lower Carboniferous rocks are lower Moscovian shales and conglomerates that are poorly preserved.

In the Aktask section, the rocks are thick limestones that range from the upper Silurian Ludlovian to the Baskirian.

Each of these sections is taken from a single second-order anticline; these folds strike east-west and are separated from each other by overthrust and reverse faults. Overthrust over the middle Carboniferous detrital-limestone-dolomite succession is a large-scale folded nappe of Silurian and Devonian shales. Within the ore field of Khaidarkan are two more or less east-west third-order anticlines, within the cores of which outcrop Carboniferous detrital, dolomitic, and limestone rocks. The Silurian and Devonian shales are overthrust on the flanks of these anticlines. The structure of these third-order anticlines is complicated even more by minor folds and by Hercynian faults of more or less east-west strike and others of diagnonal character (Kuznetsov, 1977).

The east-west faults are mainly reverse faults of which three (Ishmetau, Kara-archa, and Dolny) are most significant and extend to considerable depths. The diagonal faults are of the strike-slip variety with those of NE-SW strike being the most distinct.

As is the case at Kadamdzhai, igneous activity is weak and is represented by diabase porphyrite dikes and by small quartz monzonite stocks (these latter not being present at Kadamdzhai) [Igemberdiev, 1982].

On the contrary, Kuznetsov (1977) describes the Khaidarkan deposit as consisting of a series of closely adjacent ore sectors (including the Main ore field, the Mt. Plavikovaya, and the Kara-archa), but there are others not named. The ore field is composed of Silurian limestones and slates, Devonian limestones, lower and middle Carboniferous massive and bedded limestones and middle to upper Carboniferous shales. This description is not irreconcilable with that of Igemberdiev given above though less detailed. The complex structure of the ore field has been caused by several tectonic events. In the first of these, Silurian slates were overthrust onto the Carboniferous sequence of limestones and shales. The second was the folding of the sequence both above and below the thrust plane into box-like forms with gentle anticlinal arches with steep limbs. These folds then were broken up by both east-west and diagonal (mainly NE-SW) faults into a series of blocks that were uplifted to various degrees in relation to each other. A few of the dikes of diabase porphyry were introduced into this structural complex.

Along the traces of some of these steep reverse (E-W) and diagonal (strike-slip) faults and along the plane of the major overthrust (beneath the overlying and largely impervious shales) major bodies of breccia were developed, the fragments of which were largely silicified to produce jasperoid. These breccia bodies, therefore, either were lens-like in relation to either the vertical faults or to the plane of the major overthrust. The locations most favorable to the production of jasperoid were the crests and upper limbs of the anticlines and the traces of the steeply dipping faults (reverse or normal, east-west or strike-slip). The shapes of the bodies took several forms, such as sublayered disseminations, lenses, or mushroom-, column-, or nest-like bodies.

The introduction of the mineralization was a several-staged. As well as in the volumes of jasperoid confined beneath the overlying (and overthrust) shales, ores were located in the limestones beneath them and, to some lesser extent, in the more fractured portions of the shales above them.

As has been pointed out by Igemberdiev (1982), that the position of the Khaidarkan ore district in particular, and its associated ore bodies in general, was due to the fortunate coincidence of stratigraphic and structural patterns, the most important of which was the thick, highly porous and fractured lime-stones that were folded into various orders of anticlines and were overlain by largely impervious shales and folded into major anticlines that were broken by the major overthrusts and by the east-west and diagonal (NE-SW) faults. The various fault fractures acted as channelways for the ore-bearing fluids, concentrating them into rock volumes favorable for the deposition of their mineral loads.

Specifically, ores were deposited: (1) in the contact zones of the middle Carboniferous (Pyrkaf) limestones and the overlying Moscovian (upper

Carboniferous) (Toulbay) shales in the Southern anticline; (2) in the intricately disturbed limestone-detrital sequence of the lower portion of that same formation in the Northern anticline; (3) in the breccias that had been produced on the footwall carbonates (Carboniferous) of the major thrust fault that were overlain by Devonian shales, and in the broken rock of the steep faults on the limbs of the Northern anticline.

The ores at Khaidarkan are (Kuznetsov, 1977) quite complex with the major minerals being cinnabar, stibnite, quartz, fluorite, and calcite and the less important ones including pyrite, arsenopyrite, tetrahedrite-tennantite, getchellite [$AsSbS_3$], livingstonite [$HgSb_4S_8$ or $_9$] in which some of the antimony must be in the S^{+5} state, gold, galena, sphalerite, realgar, orpiment, sericite, dickite, and barite. It appears that, in some sectors of the Khaidarkan mine proper (and perhaps in others), livingstonite is abundant enough to be a major ore mineral and worth recovering.

The sericite is present principally in complex mercury-antimony-fluorite ores that also contain arsenic minerals and forms part of the matrix of the jasperoid breccias where it is associated with fluorite, stibnite, comb quartz, and cinnabar. In places, cinnabar is finely disseminated in the sericite to produce a material that has the color of red sealing wax. In crush zones in the shales above the ores, in addition to sericite, clay minerals and hydromicas are present, having been developed from alteration of these shales.

Selenium is present in the cinnabar in unusually large quantities (as much as 0.25 per cent). The pre-ore silicification of both limestone and shale fragments and the later and more complete production of jasperoid, plus sericitization and argillization of the shales (slates) constitutes the sum of the wall-rock alteration.

The appearances of the ore indicates that most of the ore was deposited by replacement, but considerable filling of open space also took place. The textures associated with replacementare: disseminated, streaky, nested, breccia-like, and massive; those associated with filling of open space are: drusy, comb, and cockade. The rounded textures of the major portion of these fillings of open space have been suggested to indicate deposition in the colloidal state, but more probably are the result of rapid precipitation from true solutions.

Kuznetsov reports that at least four depositional phases occurred; these have been identified as: (1) the pre-ore formation of jasperoid; (2) the precipitation of the first ore phase - quartz, fluorite, stibnite, and cinnabar; (3) the second ore phase - calcite and cinnabar; and (4) the third ore phase - realgar and orpiment. Other workers, however, think that the ore-forming process was a continuous one. In addition to quartz as the main mineral deposited (largely by replacement) in phase 1, minor amounts of sericite, fluorite, pyrite, and graphite were introduced. In phase 2, minor amounts of tennantite-tetrahedrite, chalcopyrite, and even smaller quantities of berthierite, famatinite, getchellite, metacinnabar, barite, hydromica, and calcite were deposited; stibnite was much more common than cinnabar in phase 2. In phase 3, quartz was minor in amount relative to calcite and stibnite relative to cinnabar. In phase 4, minor cinnabar was deposited along with the minor orpiment and realgar and quartz and calcite. The arsenic mineralization is most common in the Northern anticline, being concentrated in the shales overlying the mercury ore bodies.

Although Kadamdzhai was deposited near enough to the then existing surface and in sufficiently broken rocks that the ore fluids were subject to rapid loss of heat and pressure to be classified (here) as epithermal, it appears that Kuznetsov, following both Fedorchuk (1969) and Nikiforov (1969), believes that they should be classed as telethermal. The temperatures of deposition of the minerals at Khaidarkan range from 268°C to 68°C with the bulk of the cinnabar at least having been emplaced at less than 110°C, so this temperature range would fit either a telethermal or epithermal classification. After considerable thought, however, I think that, as far as I can tell at present, the structures of the area suggest that epithermal is the better classification for both Khaidarkan and Kadamdzhai.

The age of the deposit almost certainly is the same for Khaidarkan as it is for Kadamdzhai, so both are here categorized as Late Paleozoic, mainly on the age determinations made from sericite associated in time of deposition with the cinnabar.

NORIL'SK, KRASNOYARSK KRAY

Early Mesozoic	Nickel, Copper, Cobalt, Platinum Metals, Gold, Silver	Magmatic-2a Magamtic-2b Metamorphic-C

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Notes

The Noril'sk deposits are located in the general vicinity of the city of that name at 69°20'N,80°06'E and actually are composed of two ore-containing areas. The more southerly deposits (in the Noril'sk area proper) are in the south-southwest part of the district and include (in 1980) only two active mines - Medvezhiy Ruchei (Bear Creek) open pit at the northeastern end of the Noril'sk I irregularly bounded intrusion and the Zalpoyarny underground mine about 4 to 5 km to the southwest of Bear Creek. The deposits of Noril'sk II (3 to 4 km east-southeast of Noril'sk I) and at Gora Chernaya (Chernaya Mountain), 5 to 6 km east-southeast of Zalpolyarny are reported to be subeconomic, (Slowikowski, 1978).

The more northerly of the two ore areas is Talnakh, which is about 20 km north-northeast of the Noril'sk area. In the Talnakh area, the operating mines are Mayak (1965), Komsomolski (1971) and Oktyabr'sk (1974). The ore body of Taimyr was under development in 1978. The ore bodies at Noril'sk are much lower in grade than those at Talnakh (Naldrett, 1981). Naldrett suggests that the amount of ore mined and to be mined in the Noril'sk area is at least 40 per cent of that in the Sudbury district and may be much larger; certainly the Noril'sk deposits are among the largest collection of ore bodies in any area of the world and constitute the largest reserves of any Ni-Cu ores in the Soviet Union. They are also of major importance for their content of platinum group metals (PGM), as well as for nickel, copper, cobalt, gold, and silver.

A traverse east from the Ural Mountains crosses the Yenesei trough (the western Siberian lowlands) and then enters into the Siberian platform. Both the Eastern-European-Urals block and the Siberian platform have been stable cratons since the end of the Paleozoic. To the north of the Siberian platform and separated from it by the Khatanga trough is the Taymyr platform. The combined Yeniesei and Khatanga troughs form a roughly "Y" shaped area with the Khatanga trough being the right-hand arm of the Y and a bend back toward the west in the border of the Urals block provides the left-hand arm of the Y.

The Noril'sk district lies on the northwest border of the Siberian platform, just south of the junction of the Khatanga trough with the Yenesei. In this district and for major distances around it, the lower structural stage development in the platform cover is composed of lower and middle Paleozoic carbonate and argillaceous sediments that contain some beds of evaporite sulfates. These are overlain (above a strong angular unconformity) by a continental sequence of sandy argillite and coal-bearing beds that are of middle Carboniferous to Permian age (the Tungusska series, Glazkovsky, and others, 1977). Naldrett,(1981) however, says that the lower Paleozoic marine argillaceous beds are overlain by extensive Devonian evaporite beds and lower Carboniferous shallow-water limestones. Unconformably above these sediments, Naldrett says, are middle Carboniferous lagoonal and continental sediments, including coal measures. In support of these statements, he cites M. F. Smirnov (1966) and Glazkovsky and others (1977); this support is not supplied by Glazkovsky, so it must come from Smirnov. The important difference between these two descriptions of the Paleozoic rocks is the relative emphasis on the Paleozoic evaporite deposits in the district given by the two authorities. The effect of these differences on the Noril'sk district ore deposits is discussed later.

The next rocks that were developed in the district are late Permian to early (?) Triassic flood basalts and tuffs that reached the area through the north-northeast to northeast-striking Noril'sk-Kharaelakh fault system and similar fractures parallel to it. Individual faults in this system may be 500 km or more in length and may have vertical displacements of as much as 1000 m. These volcanic rocks were introduced in huge volumes, which Naldrett estimates at about 1 million cubic km; the basalts are known as the Siberian traps. These flows were fed volcanically by tholeiitic magmas that ranged in composition from sub-alkaline to diabase (dolerite) to gabbro diabase (also dolerite). From these feeder faults, sill-like intrusions of the tholeiites were introduced contemporaneously into the sedimentary-volcanic pile.

Naldrett explains this faulting as the result of large-scale rifting within the Siberian platform that was caused by movement due to convection in the underlying mantle. Up through the fractures developed by this process, a basaltic magma (derived from the mantle directly beneath the fractures) ascended to flood out over the continental surface (the basalts in the Noril'sk area were a small portion of the total amount of basalt extruded). The major difference between the flood basalts and those of such flood basalts as those of the Columbia plateau and the Deccan plateau is the lack, in these two areas of mineralized, sill-like intrusives such as those in the basaltic piles of the Siberian traps.

In the area of the Yenesei and Khatanga troughs, major subsidence began in the early Triassic and continued intermittantly into the middle Tertiary. In these troughs, the sedimentary sequence there emplaced reached a thickness of as much as 12 km. In these troughs, the depth to the Mohorovicic discontinuity, as determined by geophysical measurements, is considerably less than it is beneath the neighboring platforms. Naldrett (1980) and Tamrazyan (1971) say that the coincidence of much subsidence, accompanied by crustal thinning, beneath the troughs indicates a rifting environment. The intrusions and extrusions that took place in the early stages of rifting lie on the Siberian platform and, also, as is indicated by aeromagnetic surveys, underlie major areas of the lowlands of the troughs themselves.

The mineralized early Triassic intrusions that radiate outward and upward from the intrusive centers and even penetrate the sedimentary sequence are, as has been mentioned, directly connected with the Noril'sk-Kharaelakh block fault. Although this fault is in the rocks of the Siberian platform, it is parallel to the fault system that forms the boundary between the platform and the Siberian lowland to the west. Although the Mt. Chernaya deposit in the Noril'sk area proper and the Talnakh-Oktyabr'sk deposits farther NNE, and located on this fault, other workable deposits are spatially connected to other faults of the system. To emphasize the length of the mineralization along the Noril'sk-Kharaelakh fault, a new area of Cu-Ni ore in gabbro-dolaerite 375 km north of Noril'sk, was discovered in the mid 1970's; the mineralized area is reported to cover 1500 km^2. Of all the ore deposits, economic or uneconomic now known in the Noril'sk area, only the Imangada in the Talnakh district is not in in the zone of the Noril'sk-Kharaelakh fault; it is located in the fault zone of the same name (Imangda) about 75 km east of the N-K fault zone.

At the same time that the lavas were extruded on the surface and the sills intercalated with the among the Triassic volcanics or Paleozoic sediments the region was subjected to intense tectonic activity that produced brachy-folds and faults. The form of the sills is layer-like, irregular, or trough-like, depending on their location in relation to the various portions of the synclinal folds and to the rock types into which they were injected. The Triassic faulting has cut not only the Paleozoic sediments but also the early Triassic eruptives and irruptives.

The ore-bearing sill-like bodies are differentiated (in contrast to various undifferentiated rock types such as labradorite basalts, titanaugite diabases, and titanaugite basalts) present in the general area.

The Noril'sk I deposit has been studied in detail and, although it is not the location of the most or highest-grade ore, it serves as an example of the processes obtaining in ore concentration. The deposit in question is confined to a layer-like, differentiated intrusion that also is called

Noril'sk I and has a length in a northeast direction of 10 km. In thickness it ranges between 30 and 350 m, to average about 130 m. In cross-section it is lensoid in form and its sides are steep. As the lens is followed to the northeast, it separates into two divisions, Coal Creek (west) and Bear Creek (east). The rocks beneath the various sills and between them in some places are Permian sediments, trachydolerites, trachybasalts, and andesite-basalts. The basalts that overlie the ore-containing sill have tholeiitic textures.

The structure beneath the intrusion consists of trough-like downfolds and hollows that may be as much as 150 m deep and are as much as 600 to 1000 m in width, separated by keel-like rises. On the sides of these depressions are gentle flexures. Over the sill is a much less deformed roof that shows gentle domes; the only complication is the occurrence of small apophyses of the sill material cutting up into overlying eruptives.

Within the Noril'sk I intrusion is a well-developed stratification in which the sequence is, from bottom to top: (1) volcanic rock of clastic appearance and contact diabases; (2) picritic gabbro (i.e. olivine-rich gabbro) and norite-diabase; (3) olivine-gabbro and norite-diabases [It is not clear what the difference is to Glazkovsky and his colleagues (1977) between picritic gabbro and olivine gabbro, but they must be much the same thing unless they consider the picritic rock to be poorer in feldspar than the olivine gabbro.]; (4) gabbro diorites, gabbro and olivine-bearing diabases; and (5) eruptive breccias, hybrid rocks, and taxitic and clastic-appearing volcanic rocks. The authors just mentioned say that up to one-quarter of the igneous sequence is ultramafic in composition. Certainly, the sequence is one produced by differentiation in place with the first rocks probably being a quickly solidified contact phase, the composition of which should be much of the average of the entire sill. Above this contact phase, it seems clear that the rocks of category (2) are more nearly ultramafic than any of those above them and it probably would not be inferring too much to look on category (3) as largely ultramafic as well.

The ore of this complex consists of disseminated sulfides in the near-basal contact phase (picritic gabbro), plus more massive sulfides accumulated immediately above the contact. In these rocks, the ores occur as quite coarse nests and tabular segregations and as small veinlets of sulfides. These ores are the usual (for this type of ore) pentlandite-chalcopyrite-pyrrhotite type with minor quantities of magnetite, cubanite, and PGM as metals and minerals. The ratios of Ni:Cu = 1:1.5; Co:Ni = 1:16; and (Pt+Pd):Ni = 1:500 are as given.

The geology of the ore bodies in the Talnakh area is appreciably more complex than in the area of Noril'sk I; the igneous rocks entered through the N-K fault and radiated outward from that fault system. The major development of post-intrusion faulting in the Talnakh area greatly complicated the relationships of igneous rocks to each other and to the ores.

The ores are divided into four principal classes: (1) uniform ores, that is, ore minerals disseminated regularly through the mass of igneous rock in which they were developed; (2) veinlet-segregated ores that were driven out from the igneous-rock side of the contact of these rocks with sedimentary or volcanic rocks by tectonic forces affecting the ore area after the ores had been emplaced in their primary ultramafic host; (3) massive (or segregated) ores accumulated at the base of ultramafics that contain uniform [type (1)] ores and crystallized there; and (4) massive ores developed outside the ultramafic hosts by post-ore tectonically mobilized uniform ores [type (1) above]. The veinlet ores take several forms, depending on the composition and texture of the exocontact rocks into which they were tectonically forced. For example, in the sand-clay sedimentary sequence of the Tungusska series, the so-called veinlet ores actually are dense, fine to coarse, but lean segregations of sulfides. These segregations may take the form of thin, short veins or lenses or nests of sulfides or large lensoid bodies that are composed mainly of closely associated veins, lenses, and nests of uniform sulfide ores. Where the host rocks of these exocontact ores are metamorphosed basalt flows, rounded (amygdaloidal) coarse segregations of sulfides are arranged in numerous discontinuous bands of rich ore. Where the sills are composed of titanaugite diabase, fine veinlet ores are dominant in areas of the concentration of such veinlets.

Not only are the massive ores [type (3) above] present at the lower contacts of the ore-bearing igneous rocks, but they also are formed at and near the ends of such igneous bodies. The segregated ore within such intrusions is normal disseminated pyrrhotite-pentlandite ore, but, in the massive ores near the contacts, a higher proportion of chalcopyrite is present. These ores, from disseminated to massive, change from pyrrhotite-pentlandite-chalcopyrite to chalcopyrite-millerite (NiS) to chalcopyrite types. Beyond these copper-rich varieties are zones of pyritization where copper and nickel minerals of any type are essentially lacking.

The usual opinion of Soviet ore geologists is that the massive ores at the contacts of ultramafics and other rock types were tectonically driven injections of mobilized sulfides [produced from normal disseminated or massive ores, types (1) and (2), above] present initially in the immediately adjacent igneous body itself. In 1959, however, Rogover suggested that this type of massive ores is older than the intrusions with which it is associated and actually were deposited from hydrothermal solutions (of unstated provenience). He goes on to propose that the gabbro diabase magma may have obtained all or part of its sulfide content by the assimilation of such hydrothermal sulfides. He apparently assumed that such massive ores as are contained entirely within an ultramafic body were the result of the accumulation of molten assimilated sulfides at or near the base of the ultramafic but that the exocontact ores were remnants of unassimilated early hydrothermal sulfides. This concept does appear to be rather far-fetched and that it is much more reasonable to suppose that those massive sulfides in exocontact rocks were derived by the mobilization during post-ore tectonic activity of volumes of ore of types (1) or (3); the mobilized ores then were driven out of the ultramafic rock in which they had been generated to crystallize as exocontact massive ores.

The difference between the two massive ore types [(3) and (4) above] is provided by the place and time of their accumulation. If they are inside the ultramafic, such massive ores probably are primary; if they are outside of the ultramafic, they are post-ore, having been mobilized and driven out of the host rock by tectonic metamorphism.

The vein segregated ores associated with the Noril'sk I intrusion (mainly in the northern half) have sharp but irregular boundaries with the rock in which they are contained and are of complex form. In Noril'sk I, these ores are confined to a tectonic zone that is of shallow dip and cuts both the near-contact portions of the intrusive and the country rocks adjacent to it.

The extension of these veinlet-segregated ores into the sand-clay sediments of the Tungusska series is the result of molten material (primary or remobilized) that was driven out of the ultramafic and into the sediments. These ores exhibit a variety of forms with the most common being fine veinlets, in some places accumulated in lensoid bodies and nests. In metamorphosed and fractured basalts, the most common form is rounded segregations of course grain that form discontinuous bands of rich ore - so-called 'contact ores'. In sills of titanaugite diabase, fine-veinlet ores are most common.

As the observer moves toward the intrusive contact, the composition of the uniform ores changes from chalcopyrite-pyrrhotite-pentlandite-cubanite in the center, surrounded by massive pentlandite-cubanite-pyrrhotite, and then by mesh pentlandite-chalcopyrite-pyrrhotite ores, and then by mineralized sandstones. This arrangement indicates that the conditions governing the minerals formed in the normal ore differ remarkably over quite short distances. In places, veins containing ores of different compositions intersect; for example, gently dipping veins that contain practically nothing but pyrrhotite cut nearly vertical veins of pentlandite-chalcopyrite ore demonstrating that there were certainly more than one period of the production of mobilized uniform ores and perhaps different times at which fractures were developed. In the pyrrhotitic veins the Ni:Cu ratio is 1:0.8, but, in the chalcopyrite veins that ratio ranges from 1:2 to 1:5. As is true at Pechanga, for example, the content of PGM increases in the chalcopyritic ores over those that are higher in iron sulfides.

Veins in the Noril'sk ores also contain veins or veinlets that contain such non-metallic minerals as carbonates and chlorite and sulfides such as galena, and sphalerite among others. The late time of the introduction of these ores and their non-magmatic character suggest to Genkin (1964) and other Soviet ore geologists that they were deposited from hydrothermal solutions and were not products of magmatic processes. These "hydrothermal" minerals are not of economic value, so they do not merit additional consideration here.

The Noril'sk II deposit is in the vicinity of Noril'sk I (II is less than 5 km southeast of I). The intrusive has been introduced outward into the surrounding rocks from a dike-like feeder into an irregularly shaped body. In section the intrusive body contains picritic dolerite layers alternatingwith taxitic diabase layers. The valuable ores occur as schlieren of uniform-type sulfides in the taxitic rocks.

The Gora Chernaya deposit is an elongated ore segregation that has a trough-like form; these ores are present only in the picritic, taxitic, and contact diabases. The intrusive includes a zone of silicic hydrid rocks in the intrusive roof, and the intrusive has an alteration aureole (40 m thick) surrounding it; this zone is made up of hornfels and Mg- and Ca-skarns.

The PGM in the Noril'sk deposit are particularly concentrated in the volumes of copper-rich ore. Even the so-called native platinum is present as alloys; even what is designated as "native platinum" contains no more than 84 per cent of that element but is a rare mineral in the Noril'sk deposits. The most common platinum alloy is ferroplatinum that contains 68.5 to 81 per cent platinum. The other two abundant alloys are $Pt_3Pd_2Sn_2$ and $Pd_9Pt_2Sn_2Pb$. Among the intermetallides, only isoferroplatinum (ideal formula, Pt_3Fe) is abundant. [The information given by Razin (1977) and Cabri (1981) differs considerably, but Cabri's ideal formula appears preferable to Razin's.] The common platinum-bearing mineral at Noril'sk is a palladium-platinum-copper stannide. Common non-platimum-bearing minerals are electrum (Razin's formula = $Ag_{1.9}Au_{1-4}$) and porpetzite (Razin's formula = $Au_{13.6}Pd_{1.7}Cu_1$). The Pt-arsenide (sperrylite = $PtAs_2$) also is common at Noril'sk.

In the deposits of Noril'sk, PGM mineralization has been concentrated in the segregations and inclined veins of the uniform ores [type (1), above]; it also occurs in layer-like bodies of such ores that are essentially flat. Razin (1977) considers that the PGM in the uniform-type ores were, along with their host Cu-Ni-Fe sulfides, differentiates of the parent gabbro-dolerite intrusion; they also are present in veins produced in the exocontact rocks. The PGM metals are contained in the bottom parts of the parent intrusions (and in the exocontact veinlets). The distribution of Pd and Pt is zonal with the different minerals (and compositions) arranged alternatingly. The maximum quantities of PGM metals are most likely to be in copper-rich ores, such as chalcopyritic, talnakhitic, and cubanitic with minor amounts of the PGM being in pyrrhotitic, milleritic, and Cu_2S and $FeCu_5S_4$ types. The concentrations of Pd and Pt average about 3:1; in the segregations of rich copper ores have ratios of between Pd:Pt of 4.0 to 5.1. Rhodium is much less in amount that Pd and Pt and Iridium, ruthenium, and osmium are even less common. Varied ratios are normal for all of the PGM with those between Pt and Rh.

The various combinations of PGM occur in disordered solid solutions and intermetallic compounds of Pd and Pt with Sn, Pb, Bi, Ab, Te, and As with very rare occurrences of sulfides, tellurides, and arsenides. The minerals with which the PGM are most commonly associated are valleriite, galena, and sphalerite and, to a lesser with gold and silver minerals. In the deeper parts of the ores of the Noril'sk district, such as Talnakh and Oktyabr'sk the prevailing Pd-Pt combinations are intermetallides of these elements with solid solution not being important; solid solutions, however, are much more common in the nearer-surface deposits such as Noril'sk proper.

Native gold and gold and silver tellurides occur in the PGM combinations. Noril'sk-Talnakh is now by far the largest producer of PGM in the USSR, being 30 per cent of total Soviet production in 1947 and about 90 per cent in 1977; much more of the Noril'sk-Talnakh PGM comes from Talnakh than from Noril'sk. 1965 was the first year of production from Talnakh, profits for the Noril'sk Kombinat rose by seven times; between 1965 and 1977 production was raised another six times.

In the Talnakh area of the district, the sedimentary-volcanic rock sequence is the same as that at Noril'sk. The Talnakh field is structurally contained in the southwestern flank of the Kharaelakh trough. This trough is intersected by Noril'sk-Kharaelakh (N-K) fault system that is oriented in an essentially NNE direction and is especially deep-seated. The Noril'sk-Kharaelakh fault system is cut by a system of genetically connected thrusts and transverse faults of low amplitude. The structure of the rocks in the trough is so tilted that they dip 5° to 15°NE.

The principal structure in the ore-igneous rock-containing area is the Central graben, bounded by two major fractures. This area also has been affected by a system of genetically connected thrusts and low-amplitude folds that are at least incipiently faulted along their crests. In the zone of the N-K fault, the rocks of the Tungusska series have been folded into a large flexure that has resulted in an impressive uplift of the deeply eroded Paleozoic beds; separate blocks in the Mesozoic basalts have been displaced in different orientations.

The eastern part of the Talnakh ore field is essentially a homocline that is disturbed by only a few, subparallel normal faults that displace the rocks they cut in a step-like pattern, the deepest step being nearest to the deep-seated fault. The western part of the ore field, however, has been broken into a complex mosaic pattern in which the rocks are much crumpled and disrupted. In most places, zones of subsidence have been formed over the roofs of the igneous bodies; these are tectonically connected to the flexural-thrust displacements along the margin of the subsided zone and with the synclinal downwarps below the floors of the ore-bearing intrusions.

The ore bodies of the Talnakh area and of the Oktyabr'sk deposits are genetically connected with the introduction of the intercalated and differentiated Talnakh mafic-ultramafic intrusion. This intrusion is made up of several individual massifs that branch out from the conduit that Soviet ore geologists assume to be located in the northeastern portion of the ore field. The massifs in many places are connected by apophyses or veins. The intrusives have the forms of layers along strike but, examined in cross-section, appear as lenses or troughs. The Talnakh intrusion reaches nearest to the surface at its south end but, at its north end, has sunk from the level of the Quaternary sediments to that of the lower Devonian beds.

The Talnakh deposit itself is in three branches of an intrusive body of mafic-ultramafic rock, introduced into the area from the northeast. This mass lies to the east of the Oktyabr-sk intrusion and above it. The individual branches (massifs) are the Central, Northeastern, and Southwestern with the Northeastern being essentially a continuation to the south of the Central branch. In many places, the branches are connected by apophyses and veins of ore. The three branches together make up the "upper (Talnakh) ore stage." The ore-bearing igneous body was a sill-like mass, several km long that followed the trace of a deep-seated fault, the sill later being broken into segments by further faulting. These sill-like bodies of Talnakh rocks are from 200 to 250 m thick and are trough-shaped in cross-section.

The lower (Oktyabr'sk) body is much wider but has only two ore-bearing branches, both on its west side. The Kharaelakh is more southerly and the Lesno-Ozerono the more northerly. These two branches were intruded into Devonian beds and show varied degrees of magmatic layering. Normally, these layers are (from bottom to top): (1) taxitic and contact diabases; (2) picritic diabases, olivinites, and troctolites; (3) olivine diabases; (4) non-olivine and olivine diabases; (5) gabbro-diabases (metadolerites); and (6) eruptive breccias, leucocratic gabbros, and various types of contaminating rock. From 10-20 per cent of this sequence is olivine-enriched picritic diabases. These sills are thinner (100-150 m) than those in the Talnakh area.

The metamorphic rocks around the branches are surrounded by alteration haloes that are as much as 150 to 200 m thick, although the thickness of these haloes is much greater over the roofs. The thickness of the haloes in the lower contact zone averages about 20 m; the rocks of the exocontact are quite varied in type. The margins of the faults that cut the igneous rocks are normally quite broken (Glazkovsky, and others, 1977).

Most of the nickel-copper ores are found in lower parts of the various intrusive branches, with the endo- and exocontact volumes containing the richest ore; only a few bodies of uniform and segregated ores have been found near the roofs of the intrusions. Only in olivine-rich rocks are mineable ores usually found, although a few segregations have been noted in gabbros. Vein-segregations are recorded in the exocontact sedimentary rocks, most commonly in the skarns and serpentinites.

The Glazkovsky and his colleagues (1977) recognize three types of ore in the Talnakh deposits. These are: (1) segregated ores in the picritic, taxitic, and contact diabases - instead of segregated, massive apparently would be an equally valid designation; (2) uniform sulfide (disseminated) ores, that is ore in which the original droplets of sulfides were trapped in the interstices of (mainly) olivine crystals; (3) segregated and veinlet-segregated ores in the altered exocontact rocks. In the discussion of Noril'sk proper in this volume, Talnakh type (3) was divided into two types, the veinlet segregation and massive segregated, a designation based entirely on morphology. In both instances, the sulfides in the exocontact ores were derived from the primary ores that were mobilized and forced out of the original host rocks by tectonic pressures. The only question that is not clearly answered in published works on Noril'sk (and Talnakh) is: were the sulfides driven out of the host rocks first crystallized in the ultramafics and then later mobilized or were they driven out while still in the molten state. I suspect that the former explanation is more commonly correct, but definite evidence on this point is not available (at least to me).

In the Talnakh and Oktyabr'sk deposits, Glazkovsky and his colleagues identify five segregations of ore sulfides, closely associated in space and made up of segregated, veinlet-segregated, and uniform ores. The plan and section, the segregated ores have the same general morphological situation as the intrusions themselves. In the northeastern portion of the intrusions (Talnakh proper), the ore bodies have an upper boundary determined by the contact between olivine diabases (below) and picritic diabases (above). The lower boundary is much less regular, being located in the exocontact rocks below the intrusion and resulting from the degree of expulsion of sulfides from the primary host rocks.

In the Oktyabr'sk deposit, the structure and morphology of the ores are much more complex than at Noril'sk proper. The ores are composed of many more or less horizontal layers and lenses of mineable ore with volumes of non-commercial ore or essentially barren rock between them. Bodies of uniform ore and horizons of segregated ore are present in both the intrusion and the exocontact rocks beneath it.

The segregated ores in the Talnakh host rocks provide about 70 per cent of the total mass of the mineable ore. The segregated lenses have much the same form as the intrusions that contain them. The thickness of the ore bodies depends on the total amount of olivine-rich rocks in the layered intrusive. Interstitial (normal or uniform or disseminated) ores compose from 10 to 20 per cent of the ore volume. Taxitic rocks (taxitic must have a different meaning to Soviet geologists than it does to the Glossary of Geology) normally are 40 to 50 per cent "coarse-nested" segregations.

In the northeastern portion of the Talnakh deposit, the intrusion dips in that direction at 5° to 10° with the greatest thickness of the ores being more than 50 m and resulting mainly from a trench-like downwarp in the sole of the intrusion. In the northwestern area, a considerable number of lenses of segregated ore are arranged in a more or less vertical series; between the lenses are sections of poorly mineralized diabases.

In the Talnakh segregations, the segregation ores mainly are of the pentlandite-chalcopyrite-pyrrhotite type; only in the vicinity of disseminated ores are chalcopyrite and cubanite both present. The Ni:Cu ratio is 1:2

In the Talnakh area in general, the uniform (disseminated) sulfide ores make up about 10 per cent of the total ore content; two-thirds of these disseminated ores are reported to be in the exocontact zone at the floor of the intrusion. If this statement of Glazkovsky and his colleagues is to be taken seriously, it would mean that two-thirds of these disseminated ores somehow were collected in small molten globules during the crystallization of the intrusive and transferred to the rocks outside the intrusive by some processes not specified. It seems more probable that this is a typographical error and "exocontact" should read "endocontact".

If these ores are "exocontact", they are no longer uniform ores but are segregated ores that were forced into the altered endocontact rocks by tectonic pressures, leaving their originally associated silicate minerals behind in the intrusive. Galzkovsky and his colleagues say (or their translater makes them say) that the Southern segregation is formed of uniform sulfide ores; if he had said "from uniform sulfide ores", such a statement would have made more sense. It also is possible, however, that what is meant is that the country rocks beneath the intrusion were so complexly faulted and folded that a melt of silicate and sulfide material moved downward from the intrusion into the folded and broken country rock beneath the igneous mass. In that position, the melt crystallized slowly enough that some segregation of the sulfides and some assimilation of the country rocks took place.

Internally, the segregated ores are, in many places, zoned with the major types of commercial ore being pyrrhotite, cubanite, chalcopyrite-talnakite [$Cu_9(Fe,Ni)_8S_{16}$] [Talnakhite and mooihoekite ($Cu_9Fe_9S_{16}$)] are not uncommon minerals in the Noril'sk area with the former having been discovered at Talnakh and the latter in the platinum-bearing pipes of the eastern Bushveld. The boundaries among these various ore types in many places are stratigraphic boundaries in the country rocks. It is reported (Galzkovsky and his colleagues) that the cubanite(-talnakhite) ores are most complex and probably resulted from replacement reactions affecting pyrrhotite ores.

Chalcopyrite (and chalcopyrite-type ores) are found at the extremities or bottom layers of pyrrhotite ores as is true at Pechenga as well as in Noril'sk and Talnakh. The richest copper-bearing ores have sulfide contents of between 30 and 50 per cent; near the main fault, these ores are broken into en echelon lenses arranged in series.

At the roofs of the intrusions, exocontact ores are in saddle-like lenses that are relatively low in sulfides. The exocontact ores are quite varied in their mineralogy. Although pyrrhotite normally is the main mineral, locally chalcopyrite may be the most important; in addition, substantial amounts of millerite, pyrite, bornite and magnetite occur in the ores. The chalcocite and the copper oxides present probably are due to supergene phenomena. The number of copper, gold, and silver minerals is quite large.

Locally, the ores show a vertical zonation. Both upward and downward from the intrusion, the exocontact rocks contain nearest the intrusion, pyrrhotite ore, then chalcopyrite ore types, then millerite-bornite-chalcopyrite ore, and finally, in both directions, pyritic mineralization.

As for the platinum-palladium contents of the Noril'sk (including Talnakh) the ratios of Pd:Pt are around 3:1; further, in the rich massive copper ores, segregations and veins, the ratios are from 4:1 to 5:1. In the deposits of Noril'sk proper, and of the Noril'sk type in general, rhodium is low (where platinum averages 5.79 ppm in chalcopyrite ore, the rhodium average is only 0.44 ppm). Osmium and ruthenium are not much over half that amount, and iridium is lowest of all. In cubanite ore, the rhodium content is highest of all (0.60 ppm), but in that ore osmium is even higher (0.65 ppm). The highest content of palladium is in cubanite ore (9.10 ppm); it is this ore that contains the highest platinum as well (3.33 ppm).

In the Noril'sk area, almost all the iridium, most of the rhodium, osmium, and ruthenium, and significant amounts of platinum are as trace elements in non-PGM minerals. Not less than half of the palladium and platinum are in distinct Pt or Pd minerals. The predominant PGM minerals are those of platinum and palladium, and paragenetically these are associated with valleriite, galena, and sphalerite and minerals of gold and silver. Rh- and Ir-bearing minerals are most unusual, and Ru and Os minerals have not as yet been found. In the deposits of the Talnakh area (deeper-developed than those at Nori-'sk) the dominant PGM minerals are intermetallic alloys of Pd and Pt; solid solutions are much less common. In the higher-level deposits of Noril'sk, disordered solid solutions are the main platinum minerals. Some Pt minerals also have major additions of gold and silver, whereas, gold and silver minerals have minor amounts of the PGM in them. Pt and Pd minerals also include amounts of copper in significant amounts, and they also contain notable amounts of nickel, tin, lead, and bismuth (0.n to -n per cent).

The PGM minerals are normally most irregularly distributed in the ore with their major concentrations in portions of the disseminated copper minerals, in the selvages of veins, at the contacts of hanging-wall segregations, and in the talnakhite veinlets branching out from these. In the massive (segregated) ores, the PGM, gold, and silver minerals are around the outer margins of these masses as well as in the hanging walls of schlieren of disseminated ores.

The common pyrrhotite ores in deposits of the Noril'sk type will be more likely to contain Pt and Pd minerals; these minerals are tiny (1 to 2 microns across) and are present, even in these ores, only where chalcopyrite, talnakhite, or cubanite are among the ore minerals. The least PGM concentrations are in ores containing millerite, Razin (1977) considering the millerite to be hypogene.

Segregations of Pt and Pd minerals are seldom more than 0.1 mm in diameter, and these are exceedingly rare, but in chalcopyrite, talnakhite, or cubanite ores, they may be 4 to 6 mm in size. Normally, monomineralic types may be 15 to 20 microns in diameter, and polymineralic types average between 30 and 100 microns. Needless to say these minerals were impossible to identify until electron microprobes were perfected.

How the various ore types of the Noril'sk and Talnakh deposits were developed is not definitely defined. From sections along and across an apparently typical Noril'sk type intrusion shows that the ultramafic portion of the igneous body is basal to the intrusion with most of that ultramafic mass resting on altered exocontact rocks, but with some of it lying directly on gabbro-diabases of the Talnakh intrusion. In a few places, the ultramafic core rests, without other rocks intervening, on the sandstones, shales, and argillites of the Tungusska series. Within the irregularly lens-shaped core of ultramafic rocks, the space is occupied mainly by various types of ore; that is, (1) pentalandite-chalcopyrite-pyrrhotite, (2) pentlandite-chalcopyrite-pyrrhotite-cubanite, and (3) pentlandite-cubanite-chalcopyrite. Obviously, not all the space in this lens is composed of sulfides; silicates, mainly olivine, take up the volumes not filled with sulfides, but sulfides are the dominant mineral. Just what percentage of the ore volume must be silicates before the ore classification changes from segregated (massive) to uniform (disseminated) is not clear but it must be appreciable, if not dominant.

How these lenses were accumulated probably was the result of the intrusive's first stages of converting from melt to solid-molten material beginning with the crystallization of olivines and the separation of a molten solutions of the sulfides available in the original magma. This mass of silicates and still molten sulfides, being heavier than the remaining molten material, sank through it. In most instances this silicate-sulfide mass reached to, or near, the base of the Talnakh or Noril'sk intrusion in question. There it either completed its crystallization (silicates and sulfides) or some of this silicate-sulfide mush or molten sulfides only entered tectonically created openings in the exocontact rocks to produce the various types of ore concentrations found in such positions. In general, the Soviet geologists working on deposits of this type believe that such exocontact sulfides were mobilized and driven out of their original sites of crystallization within the intrusive by tectonic forces that produced openings in the exocontact rocks into which the mobilized sulfides could enter. What probably happened was a combination of both processes. Some fractures existed in the exocontact rocks when the crystal mush-sulfide melt material sank into a basal position and into these fractured rocks the mush-like material entered to produce veinlet-segregated ores. Later, tectonic activity mobilized appreciable amounts of sulfide material and developed fractures in the exocontact rocks into which these sulfides could be forced to produce the volumes of massive (nearly silicate-free) sulfide ores.

If this is the correct interpretation of the geology of the Noril'sk-Talnakh ores, then they should be classified as Magmatic-2a,b, plus Metamorphic-C.

Gold is produced from various operations in the Noril'sk region, but the amounts are small. The Medvezhiy open pit is being exploited for gold (among other, much more important elements) even though the gold content is only 0.18 gm per ton. The underground workings at Talnakh and Oktyabr'sk probably have twice the grade of the Medvezhiy pit, but the operating combine at these mines is able to recover less than 1 ton per year total gold.

TETYUKHE (DAL'NAGORSK), PRIMOR'E KRAY

Late Mesozoic	Zinc, Lead	Hypothermal-2 to Mesothermal (?)

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Notes

The deposits of the Tetyukhe ore field are located around the small town of Tetyukhe (Dal'nagorsk) at 44°35'N,135°35'E, about 360 km ENE of the great port city and naval base of Vladivostock and nearly 40 km northwest of the port of Tetyukhe Pristan (Pristan = landing stage). The area has been known to have been mineralized since 1897 but the exploitation of the first deposit (Verkhnee) was not begun until 1910. The Verkhnee (Upper) Deposit is now essentially worked out, but several other ore bodies are being worked or will be worked when exploration is complete. These include: First Soviet, Second Soviet, Eastern, Middle, and Western Partisan, Svetloe (bright), and Tigrovo. Some of the newly discovered deposits (Nikolaevo and Sadovo) do not outcrop at the surface. Sadovo was scheduled to begin operation in 1976. All of these deposits appear to have been formed within a short time of each other and occupy similar geologic positions and have much the same ore body morphology, mineral composition (ore and gangue), and history of ore formation.

The Tetyukhe ore field belongs to the skarn type of lead-zinc deposits; the skarn minerals are mainly pyroxenes and garnets with sulfides and a wide variety of other high-temperature minerals. Like nearly all other deposits of this type, those in the Tetyukhe area are spatially associated in the exocontact zone of silicic igneous rocks of hypabyssal or near-surface loci of emplacement. From the large quantities of calcium ion in the carbonate rocks invaded by ore fluids rising along the contact of igneous and carbonate rocks, calcium silicates are normally numbered among the important gangue minerals of any skarn deposit. As the process progressed, additional and more complex silicates were formed and oxides and sulfides began to be introduced. Similar deposits to Tetyukhe are known throughout the world and include such deposits as the Hanover and Magdelena districts in New Mexico, Darwin in California, Concepcion del Oro-Proviencia in Mexico, Eureka, Nevada, and Trepca in Yuogslavia.

The ore field of Tetyukhe skarn deposits is located in the center of the drainage area of the Tetyukhe river that flows down for the eastern slope of the Sikhote-Alin (mountain-range) to the Sea of Japan.

Two structural events were prominent in the developemnt of the geological structure of this ore field. The lower stage was the production of the folded basement that include geosynclinally folded formations of Carboniferous, Permian, Triassic, Jurassic, and early Cretaceous age. These formations include siliceous breccias, siliceous shales, siltstones, sandstones, and limestone, the thickness of which equals about 8 km. These folded rocks have been appreciably faulted, producing a definite block structure in the basement. The upper (or second structural) stage began with the development of a cover over the basement of volcanic and volcano-sedimentary rocks of late Cretaceous and Paleogene age, these beds have a thickness of about 1 km and fill graben-like depressions on the basement floor. These late Mesozoic - early Tertiary rocks are crystallized lavas, ignimbrites, lava breccias, and terrestrial tuffs, mainly of intermediate composition in the lower part of the sequence and silicic in its upper parts. As related events, hypabyssal and subvolcanic diorites, granite porphyries, and syenite porphyries were intruded in the late Cretaceous and small bodies of diorites, granites, and dikes of diorite and diabase porphyrites in the Paleogene.

In the Verkhnee deposit, a sequence of Carnian (upper Triassic) limestone is a major structural factor; onto this limestone, thrust faulting [the Glavnyi (Main) fault] has brought quartz porphyries of upper Cretaceous age in over the Carnian (upper Triassic). Lower in the massive Carnian carbonates are intercalations of sandstones and clay and siliceous shales that are over 500 m thick. These overthrust quartz porphyries were originally part of a much larger lava sheet that covers a variety of materials, ranging from blocks of siliceous shales and limestones to the remnants of forests (indicating that these underlying formations and the lavas also were continental beds) (Smirnov and Gorzhevsky, 1977).

The Main (thrust) fault that brought the quartz porphyries into their present position moved in a northeast direction (045° to 070°) and has a southeastern dip of 30° to 40°: it has a vertical throw of hundreds of meters. The plane of the Main fault has been filled by tectonic clay gouge, the thickness of this gouge is as much as 0.5 m. In addition to the Main fault, minor faults are generally parallel to it but approach it at different angles; among these faults are ones designated as Soputsvuyushchii (= accompanying) and Amur.

The main ore body at Verkhnee occurs in the limestones that are directly under the quartz porphyries of the overthrust. From this it follows that the contact between the limestone and the overlying igneous rock is tectonic and not intrusive. Nevertheless, this fault not only moved before the ores were deposited below it but also moved during the emplacement of the ore and after that event had been terminated.

The main ore body has the shape of a branching, pipe-like structure, somewhat flattened against, and extended along, the Main fault; at this stage, the fault dips about 45°SE. Near the present surface, the ore body was split into three branches; lower down it merged into two stems, and still further down these joined into one main trunk. From this trunk a major apophysis with which are associated several minor ones turned steeply down from the footwall of the trunk. At the surface, the ore body is some 400 m long; this length becomes less with depth, but, in the same volume, the width rises from 8 as far as to 60 m, giving much the same area as at the surface.

The main pipe structure can be followed downward for 500 m, measured vertically. On the northeast flank of the deposit, several minor trunks or branches take off from the main ore body. One of these branches, the Zaporfiritovo, is cut by a dike of porphyrite. From its field relations, it is apparent that this dike was injected after most of the ore had been deposited. It contains, however, veinlets of ore minerals that were emplaced in the late stages of ore formation. This dike, therefore, is intra-ore.

The Verkhnee deposit, ore and skarn, contains over 40 minerals of which the main ones are hedenbergite, sphalerite, galena, pyrrhotite, quartz, and calcite. In the upper reaches of the deposit, the ore body included large spheroidal or radiating aggregates of hedenbergite that were 1 m or more in diameter. This early mineral was considerably replaced by both sulfides and later gangue minerals, but, to a limited degree, these late silicates contributed to the total skarn mass. Large vugs are a common feature of this part of the deposit, and these are lined with impressive crystals aggregates of sulfides, quartz, and calcite. Some of these open spaces contained what is reported to have been silica gel with shreds of palygorskite $[(Mg,Al)_2Si_4O_{10}(OH)\cdot 4H_2O]$.

The upper portion of the deposit, down to a few tens of meters below the surface, was oxidized with the development of oxide minerals of iron, zinc, lead, and copper. The calamine ores were of much commercial importance but are now long exhausted; these ores were located in the footwall of the main ore body.

In addition to the principal minerals, just mentioned, some 18 minor minerals included chalcopyrite, arsenopyrite, pyrite, marcasite, axinite [general formula $A_3Al_2BSi_4O_{15}(OH)$; $A = Ca,Fe^2,Mn^2$], garnet, wollastonite, ilvaite, fluorite, datolite, siderite, epidote, zoisite, apophyllite, hisingerite, stilpnomelane, and dickite. In addition, there are 17 minerals listed as rare.

Radkevich and others (1960) recognized four stages of mineralization in the Verkhnee and other deposits of the Tetyukhe region. The first stage was the pre-ore skarn stage that these authors consider was formed above 600°C, and its major minerals were andradite garnet with an 8 per cent grossularite.

The second stage produced the bulk of the skarn and of the sulfides; the skarn minerals were mainly hedenbergite, garnet, and axinite, plus some ilvaite, datolite, and early pyrrhotite, arsenpyrite, sphalerite, galena, galenobismuthinite, quartz, and aclcite. Both the hedenbergite and ilvaite were Mn-rich varieties. These authors place the temperature range during this deposition between 600° and 400°C.

In the third stage, most of the sphalerite and galena were deposited with quartz and calcite and appreciable arsenopyrite and pyrrhotite; the ore contains several kinds of quartz and seven crystallographic varieites of calcite. The authors estimate the temperature range to have been between 350° and 120°C.

In the deposition of the fourth phase quartz, chalcedony, calcite, marcalcite, fluorite, hisingerite, zeolites, and palygorskite were formed in a temperature range from 100° to 20°C.

As the Verkhnee deposit is followed downward, the coarse-grained hedenbergite becomes finer in grain and axinite is present in large amounts. Also, with depth, sphalerite increases in relation to galena. Unfortunately, with depth, the tenor of the ore in metals decreases.

The other deposits of the Tetyukhe field also exhibit various forms, pipes, lenses, mantos, and veins, and in these are similar skarn-sulfide replacement bodies in the Carnian (upper Triassic) limestones.

Another of the Tetyukhe deposits is that of Nikolaevo; it is contained in a down-dropped block of Carnian limestone and is covered by upper Cretaceous rhyolitic tuff breccias. Both rock types were broken by NW-SE and more or less north-south faults. The ore body is located at depths of

700 to 1200 m below the surface and immediately below the overlying volcanic rocks; it does not outcrop at the surface. It principally is a gentle dipping lens from which pipe-like branches emerge. The ore is a typical skarn-type association of skarn silicates, sulfides, and quartz, quite similar to the other deposits of the Tetyukhe region. In the volcanic rocks above the ore, quartz-sulfide veins have been formed; these veins are surrounded by hydrothermal alteration haloes that contain some disseminated sulfides.

The East Partisan deposit is composed of a series of skarn-sulfide pipes, one of which is designated as the Main ore body. This ore body dips steeply and curves with depth and is confined to the body of broken rock produced where two major faults, one striking slightly west of north and the other NW-SE intersected. Into the volume of broken rock around the junction of these two faults, a neck-like body of andesite is reported to have been intruded before the ore was introduced. The first step in the actual process of ore emplacement was the development of a body of skarn, rich in hedenbergite in its upper levels and probably less rich in wollastonite further down. Neither of these two minerals would be expected as the hypothermal alteration products of an andesite. This indicates that: (1) the original volume of the andesite neck was much less than the total volume of broken rock at the fault junction, the rest being limestone or (2) the ore-forming fluid had essentially saturated itself as it worked its way up through the fractured country limestone and was able to convert the silicates of the andesite to wollastonite and hedenbergite or (3) a combination of the first two.

At any event, the broken rock at the fault junction was a calcium-silicate-rich skarn before ore deposition began. The hedenbergite skarn is rich in zinc and lead sulfides, but the wollastonite variety is much leaner. From the highly ore-mineralized rocks in the center of the rock volume broken by the two faults, apophyses of ore-bearing skarn follow the fault fractures out into the less broken country limestone around them.

The ores of the Tetyukhe district are typical of skarn-type lead-zinc deposits all over the world. Much, though not all of the skarn, was developed prior to the introduction of the sulfides, but the continued development of the skarn silicates during much of the precipitation of sulfides shows that a major proportion of the sulfides were emplaced in the hypothermal range. Nevertheless, it appears probable that an appreciable amount of the sulfides was produced at temperatures below those of the hypothermal range, that is below 350° to 300°C. Thus, the classification of the Tetyukhe ores probably should be Hypothermal-2 to Mesothermal. The skarn ores in the brecciated andesite in the East Partisan mine should be categorized as Hypothermal-1, but such ores probably constitute no more than a minor fraction of the total in the district.

Production from the Tetyuke ore field appears to be about 1.5 million tons per year from which, among other metals, about 1.13 tons of gold are recovered.

KAL'MAKYR, UZBEKISTAN

Late Paleozoic	Copper, Molybdenum, Gold Rhenium	Hypothermal-1 to Telethermal

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Notes

The Kal'makyr porphyry-copper deposit probably is the largest deposit of its type in the Soviet Union, producing some 23 million tons of ore a year as opposed to 14,000,000 from Kounrad. Kal'makyr is reported to have reserves of ore that contain 3 million tons of copper metal. The deposit is only about 70 km ESE of Tashkent, and is only about 35 km NE of the concentrating and smelting center of Almalyk (where ores from the Altyn-Topkan lead-zinc mines are concentrated and the zinc concentrates smelted); the lead concentrates from Altyn-Topkan are smelted in Chimkent, 160 km north of Almalyk; the Altyn-Topkan mines are in Tadzhikistan.

The Kal'makyr deposit was discovered in 1925 by S. Mashkovtsev, but open-pit mining did not begin until 1954, the original annual tonnage being 8 million; in 1960 work was begun to expand production to 20 million tons a year and, in 1970, to increase that to 23 million tons a year. The average grade of ore from Kal'makyr is 0.62 per cent Cu, 0.005 per cent Mo, and 1 gram of gold per ton. It appears that Kal'makyr yields over 17 tons of gold per year. Despite its low grade, molybdenum recovery at Kal'makyr is sufficient to make it an important by-product of copper mining.

The deposit is located on the northern part of what is known as the Kuraminsk structural-facies zone of the western Tyan-Shan mountains. Important left-lateral thrust faults with near east-west and northeast strikes were major influences in causing the structural pattern of the Kuraminsk facies. Two of these faults (Burgunda and Miskan) divide the Kuraminsk zone into three parts, in each of which the structure is different. The central block has been dropped down the farthest, and it contains Carboniferous sequences of volcanic rocks. The southern block is the most uplifted of the three and has been, thereby, most deeply eroded; it is composed of Silurian rocks that are intruded by large massifs of granodiorite. The northern block is intermediate in its elevation and contains an appreciable mass of syenite-diorites of the Almalyk intrusive complex, and about the same volume of carbonate and volcanic rocks of late Devonian to middle Carboniferous age.

The entire Almalyk ore region is contained in the northern block of Kuraminsk structural zone, a zone in which magmatic rocks of various ages and compositions take up more than 55 per cent of the total area. Volcanic rocks make up about 40 per cent, leaving somewhat less than 5 per cent for stratified beds (detrial sediments and carbonate rocks). The huge amount of igneous rock masses in the zone and the large number of varied fractures make it impossible to determine if major fold structures exist in the zone or not. Only where stratified rocks occur in the roofs of igneous massifs or in the most down-dropped blocks can unrelated segments of folds be worked out. Among the various rock types that have been identified are lower Devonian volcanic formations (andesites to rhyolites), Givetian (middle Devonian) conglomerates and limestones, and Frasnian (immediately above Givetian) and lower Carboniferous dolomites, limestones, and marls. This rock group is completed with silicic and intermediate volcanic beds with intercalated lenses of conglomerates and organic limestones (the Minbulak group of lower Carboniferous age). No further solid rocks were developed in the Almalyk region, although, in the Cenozoic, loess sheets covered the area.

The porphyry-copper deposits and the minor porphyry-copper indications in the region are arranged in two NW-SE-striking belts; these belts are the Balykta and the Nakpai deep-seated faults. These belts have been defined as: "concealed zones of deep-seated faults" or "zones of increased permeability". The rocks in these belts have been hydrothermally altered and highly jointed; these events definitely affect the gravity and magnetic fields; further, primary and secondary dissemination and various trace-element haloes and induced polarization (IP) anomalies have been produced. These belts also contain small massifs of granodiorite porphyry that were moved into these belts by thrust faults that strike more or less east-west. It is in these massifs that all the porphyry-copper deposits or non-ore mineralization have been produced.

The Kal'makyr deposit itself is contained mainly in igneous rocks of Carboniferous age, but it also includes remnants of the covering roof as well as xenoliths of quartz porphyries of early Devonian age and limestones of the late Devonian (Samonov and Pozharisky, 1977).

The intrusive rocks in the Kal'makyr deposits are part of the large Almalyk syenite-diorite massifs of middle Carboniferous age; this syenite-diorite is known far beyond the boundaries of the porphyry-copper deposit. Pokrovsky (1959) considered that the syenite-diorite massif was formed in four stages. Of the first stage, gabbro-diorites and gabbro were formed but only small xenoliths of them remain. The second stage rocks generally are syenite-diorites, and they make up a major part of the ore area; locally, these syenite-diorites grade into various other types such as syenite-monzonites and diorites. The syenite-diorites are greenish-gray rocks of medium to coarse grain; they contain plagioclase (Ab_{30} to $_{60}$) 45 to 50 per cent; orthoclase 25 to 30 per cent; biotite about 20 per cent (rarely hornblende substitutes for biotite); quartz 2 to 3 per cent; a wide variety of accessories constitute no more than 1 per cent.

The syenite-diorite rocks of the Kal'makyr ore field are intruded by granodiorite-porphyry stocks, and these are reported to have a major influence in the location of the ores in the syenite-diorite rocks. The granodiorites are grayish-pink to light pink and include two kinds of phenocrysts, oval ones of pink K-feldspar and abundant tabular ones of white plagioclase. These phenocrysts make up from 10 to 80 per cent of the granodiorite and are contained in a ground mass of K-feldspar, quartz, minor sericite flakes, and chlorite; the rock texture is microgranitic and micro-syenitic.

The porphyry-copper deposit contains two of these granodiorite-porphyry stocks: (1) the Kal'makyr stock which is centrally located and (2) the Akcheck stock that can be partly observed along the southern margin of the open pit. In the upper portion of the deposit, the stocks have steep contacts against the syenite-diorite; as they are followed downward, the dips become flatter and are characterized by numerous apophyses that extend for

10 to 200 m into the syenite-diorite and may be as much as 100 m across or as little as 1 m.

Still younger than the granodiorite stocks (and constituting the fourth phase of the Kal'makyr intrusion) are veins of granodiorite porphyries and of dacite porphyries; they are much darker in color than the earlier intrusive rocks. These dikes probably are older than the ores (pre-ore), but this problem apparently is not settled with certainty.

The entire suite of igneous rocks has been hydrothermally altered - chloritized, serpentinized, biotitized, and largely replaced by orthoclase so that such altered rocks seem, at first glance, to be orthoclasites. Viktorov (1964), however, points out that these various altered rocks are complexly zoned both vertically and horizontally. At appreciable depths, below the commercial mineralization at the base of the ore-bearing rock, orthoclase and magnetite are strongly developed. In the mineralized ore horizons above the orthoclase-magnetite, the changes in the granodiorite are quite varied. In the center of the ore field, the granodiorite has largely been converted to a hydrothermal quartzite; passing outward from this quartz-rich zone, the amount of this secondary quartz is drastically reduced, and that of secondary sericite is markedly but irregularly increased. In the outer margins of the ore body, the alteration consists of much secondary biotite, chlorite, and amphiboles; at lower horizons, this amphibole is replaced by epidote. Pyrite is important throughout the deposit and even extends beyond the ore volume. It should be noted, however, that this alteration, impressive as it sounds, is much less intense than that at Kounrad (see Kounrad description).

The Kal'makyr ore body is contained in a stockwork enclosed in the major volume of altered rock just described. The stockwork takes the form of an inverted cone with some elongation along a NW-SE direction; the actual line of demarkation between ore-bearing stockwork and the surrounding altered rocks can be determined only through sampling. The stockwork is hundreds of meters deep, and its lower portion breaks up into separate tongues; drill holes that reached as much as 800 m beneath the surface have not penetrated rock totally lacking in ore minerals. On each level of the deposit, the richest ore is found immediately adjacent to a core barren of ore. Toward the ore margins, the ore becomes of lower grade and is arranged in ring-shaped bodies. Beyond the ore boundary fixed by sampling, local concentrations of copper minerals are present for hundreds of meters; these apparently are not of sufficient size to justify mining.

The degree to which the rocks in the stockwork have been broken and crushed has produced a huge network of joints that formed a perfect pattern of channelways up which the ore-forming fluids could pass. With the passage of time (as well as of ore fluids) these veinlet-fractures gradually healed with their filling by quartz and, to a much lesser degree, by calcite or anhydrite. In addition, appreciable amounts of ore minerals, pyrite, chalcopyrite, pyrrhotite, tetrahedrite-tennantite, chalcocite, and molybdenite, plus much lesser amounts of other sulfide and gangue minerals were strongly disseminated. These veins range in thickness from fractions of a millimeter to 3 to 4 cm; their lengths were between a few centimeters up to tens of centimeters. To the present time, workers in the Kal'makyr ore region have not been able to decipher either definite strike orientations or clear patterns of their distribution. Locally, between the veins, small replacement bodies have developed and aggregated into connected accumulations of pyrite, chalcopyrite, magnetite, and, in places, minor amounts of other sulfides that may be big enough to be recovered commercially. The quantity of these aggregations (segregations) of sulfides are appreciably smaller in the ores at Kounrad.

The textures of the Kal'makyr ores are subhedral, porphyritic, and platy; some of the sulfides definitely are corroded by those later in the paragenetic sequence. The principal fabrics exhibited are veinlet (65 to 70 per cent of the sulfides) and segregated (30 to 35 per cent).

The three stages of mineralization (as recognized by Badalov, 1965) are: (1) early copper-molybdenum stage when veins as much as 1 m wide and veinlets as much as 1 cm were formed; these are present throughout the ore deposit but are most abundant in the groundmass of the ores in the ore-body margin; (2) the pyrite stage, the mineralization of which is widely and consistently distributed; the intensity of the 2nd stage mineralization increases toward the zones of copper ore; and (3) the third stage (chalcocite) is made up of mineralization disseminated in veinlets that occur in steeply dipping zones; these zones are from 3 to 4 to 8 to 10 m thick; the strike of these zones ranges, from one part of the deposit to another, from more or less east-west to NNE-SSW; these zones, in most places, have the same orientation as the faults in the ore body. The distances between the chalcocite zones are far from the same; further, the intensity of copper mineralization goes down, in many places between zones, essentially to zero.

The primary ores contain the overwhelming majority of the mineable reserves of the Kal'makyr deposit. In this ore type, the principal minerals are pyrite, chalcopyrite, magnetite, and molybdenite with most of the copper coming from chalcopyrite. Molybdenite occurs in monomineralic veins as large rosettes and as fine disseminations in the portions of the ore body that have been converted almost entirely to quartz; molybdenite also is in veins and veinlets with chalcopyrite and magnetite. The molybdenum contains traces of rhenium on a consistent basis; rhenium appears to be recovered. The rather uniform occurrence of gold (1 gm per ton) adds appreciably to the value of the ore of Kal'makyr.

It is difficult to be certain of the importance of the third stage chalcocite. In one place, page 131, Samonov and Pozharisky (1977) say that "of greatest importance is the somewhat later, strictly copper (chalcocite) phase...". On page 133, paragraph 2, they say: "The principal commercial value is in the chalcopyrite". This would suggest that the third-stage chalcocite is economically worth recovering, but that the mine would not be much if it had to depend entirely on the chalcocite contained in it.

The primary ore in the Kounrad deposit is here classified as "Hypothermal-1 to Mesothermal". The Kal'makyr ore appears to have been formed at a somewhat higher temperature than that of Kounrad, but that some of the Kal'makyr chalcopyrite was introduced in the Mesothermal range. Further, the primary veinlet chalcocite can only be classed as telethermal (consider the position of the Butte chalcocite). Thus, the Kal'makyr deposit was introduced over the range Hypothermal-1 to Telethermal with the bulk of the copper being contained in probably Hypothermal chalcopyrite.

As the silicic igneous suite with which the ores are spatially, temporally, and almost certainly genetically connected is of mid-Carboniferous age, the age of the Kal'makyr ore definitely is late Paleozoic.

The zones of oxidation and leaching are in much the same topographic positions; the granodiorites and syeno-diorites are converted to a greenish color and locally to a brownish hue. Within the zones of oxidation and leaching, the degree of oxidation of copper minerals differs widely over short distances, so that oxidized, mixed, and enriched ores are contained in adjacent wall rocks in minor volumes. The zone of oxidation, as would be expected, is located at contacts between granodiorites and syenite-diorites and along faults. Where the solid rock-loess contact forms the base of small depressions, the oxidized zones are of minimum thickness.

The sections of the oxidized zone that contain commercially valuable copper have most peculiar shapes determined mainly by the original character of the ore and the downward paths available to groundwater. Malachite is the major copper mineral in the zone of oxidation with, locally, enough chrysocolla, turquoise, and other oxide minerals of copper to be of economic worth.

There are only three broken, lens-like, steeply dipping bodies of secondary chalcocite ore in the deposit, and these are confined to fault zones. These confining fault zones are the Kal'makyr and two parallel faults, all of which strike to the northeast. Along strike, these secondary bodies extend for 200 to 2500 m with a maximum thickness of 60 m.

These three independently economic secondary enrichment bodies are not located in response to relief, or the position of the contact of granodiorite-porphyry stocks. In addition to these three, there are many small enriched masses in which chalcocite is the main ore mineral. These smaller deposits seem to follow no definite pattern of location; the most important minerals are chalcocite and covellite, plus lesser pyrite, chalcopyrite, and molybdenite (presumably all relics of primary mineralization) and some hematite and rare galena, sphalerite, and gold, also probably all relics.

The oxidation and secondary enrichment bodies are so small that they formed over a geologically short period of time, probably in the Mezozoic and before the Paleogene transgression when the sea covered the entire Kuraminsk area. Some further oxidation may have taken place at the beginning of the Paleocene and during a major period of mountain building which brought the Kal'makyr deposit to the surface again. The tectonic activity ended abruptly and was followed by the development of the loess cover that prevented the penetration of the deposit by surface waters.

The oxidation and secondary enrichment deposits should be classified as Ground Water-B2, but their abundance is so small as not to make it worthwhile to mention them officially.

MURANTAU, WESTERN UZBEKISTAN (SOUTHERN TYAN' SHAN')

Late Paleozoic	Gold, minor Silver	Hypothermal-1 to Mesothermal

Borodaevskaya, M. B. and Rozhkov, I. S., 1977, The Muruntau deposit, in Smirnov, V. I., Editor, Ore Deposits of the USSR: Vol. III, Pitman Pub. Co., London, p. 35-40

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Petrovskaya, N. V., 1967, Mineral associations in the gold deposits of the Soviet Union: Central Scientific Investigations, Geol. Devel. Inst. Pr., no. 76, p. 78-112

Sher, S. D. and Yudin, I. M., 1971, Polymetallic transformations of the ore-enclosing sedimentary rocks in the Muruntau gold-ore deposit: (central Kyzyl-Kumy): Central Scientific Investigations, Geol. Devel. Inst. Pr., no. 96, pt. 2, p. 30-48

Zakharevich, K. V., 1980, The genesis of the gold and silver deposits in central Kyzylkum: Problems on Magmatism and Metamorphism, v. 6, p. 105-114

Notes

The gold deposit of Murantau is located approximately at 39°00'N, 70°00'E and is reported to be (Glynn, 1978) the largest, and among the lowest cost, of the gold mines of the Soviet Union. The Murantau gold field lies in the center of the Kyzyl Kumi (Red Sands) desert at the foot of the Murantau (Nose mountain) that forms the southeast extremity of the Tamdytau ridge of mountains. As the Tamdytau mountains appear to center about

41°23'N,64°23'E, the Murantau deposit almost certainly cannot be where the translator says it is. This is particularly true when it is noted that 39°00'N,70°00'E is in the Tadzhikistan SSR and not in Uzbekistan at all. If Murantau actually is at 41°23'N,64°23'E (it probably is about 10 km Se of that point), it is about 190 km almost due north of Bukhara. The statement by Borodaevskaya and Rozhkov (1977) that Muruntau is in the southern Tyan'-Shan' also must be challenged as it can be seen from Sher and Yudin (1971) that Muruntau definitely is in the Kyzl Kumi desert which is at least in part in western Uzbekistan. It hardly seems possible that a mistake in location of this magnitude could have passed the editors unnoticed.

It is probably safe to accept Glynn's statement (1978) that gold was discovered at Muruntau in 1958 and that thereafter widespread exploration was conducted, mainly by trenching through the 1 to 3 m of red sand overburden and by ore drilling. Why the original exploration was carried out here in the Kyzl Kumi desert has not been explained in any literature I have seen. At any event, what is known of the deposit suggests that it has a tree-like shape with the trunk (which extends down to considerable depth) containing a higher grade of ore than that found in the branches.

The gold (Glynn, 1978) occurs in both large quartz veins (as much as 20 m thick and 300 m long and in stockwork structure, the veins of which are smaller quartz-filled fractures in which the sulfides are pyrite and arsenopyrite. Actually, the whole rock volume at Muruntau is reported to make up one huge stockwork, the area of which is about 4 km^2. The deposit can be separated into five ore zones divided from each other by largely barren zones. The boundaries between gold-enriched ore zones are indistinct and indefinite and can be established only be detailed sampling. It is Glynn's opinion that the Soviet geologists who have worked on the deposit are not agreed completely on the details of how it was formed, but the generally accepted concept is that it was formed by a multistage process that repeatedly alternated metamorphic and hydrothermal activity. The Soviet view of the details of the ore-forming process will be discussed below.

Almost all the gold in Muruntau is free gold that occurs in two varieties: (1) gold averaging 900 fine and (2) about 750 fine. The 900-fine gold occurs in the upper part of the deposit and may represent a supergene concentration of the primary 750-fine gold. The gold particles range in size from less than 0.001 mm (very fine) to rather coarse (up to 1 mm). The gold is present as unevenly disseminated fine particles between individual quartz grains and also in association with sulfides. These sulfides make up on 0.5 to 1.5 per cent of the ore, but gold also is located among the sulfide grains (as it is among those of quartz); the gold particularly is found with arsenopyrite, a common case in gold deposits all over the world.

The area is so dry that, in the early stages of work in the area, water had to be transported by truck from an artesian source at Tamdybulak, 50 km to the north of the mine. Before milling could begin, it was necessary to construct a 1.2-m pipeline from the Amudar'ya river, 220 km to the southwest. Construction of both open pit and mill were begun in the spring of 1967, and the first gold bar was poured on 25 July 1969. The first stage of construction was finished in the early 1970's, and immediately thereafter a two- to three-fold expansion of mine (including the opening of an underground mine) and mill was begun. For a detailed statement as to mining and milling practices, the reader is referred to Glynn's article.

Borodaevskaya and Rozhkov (1978) begin by saying that the ore deposit is located within the Hercynian geosynclinal belt of the southern Tyan'-Shan'. Considering the considerable distance (perhaps as much as 1000 km) between the Tyan'-Shan' mountains and the Kzyl Kumi desert, this statement must be taken with a grain of salt. I am accepting, perhaps unjustifiably, the detailed report on the geology on the Muruntau mine because the faking of such a study would be more work than it possibly could be worth.

In the event that my assumption is correct, the ore body at Muruntau is located in the oldest sedimentary-metamorphic formations in a Hercynian geosynclinal belt whatever its true name may be. The age of these formations is in dispute with some workers considering the beds to be Silurian and others that they are Precambrian. At any event, these beds have been divided into a number of groups. These are (from bottom to top): (1) Auminza; (2) Taskazgan; and (3) Bespan. Carbonate-detrital-volcanic sediments are the dominant type in the lower part of the rock sequence; at present, the rocks are composed of green amphibole and quartz-mica schists, quartzites, and limestones; metamorphic minerals present include: muscovite, biotite, hornblende, and garnet. In the upper part of the series, the ratio of greenschists to other rocks goes down and the detrital rocks are most abundant. The metamorphism is of lower grade than in the lower portion, being in the chlorite-sericite facies.

The Murantau ores are contained in the rocks of the Bespan group; these rocks comprise a flysch-like series of thin, interstratified beds. Both the physical and mechanical properites of the beds and their composition did much to determine the morphology and distribution of the gold ores.

What igneous rocks there are in the area are largely located to the margin of the ore region and are made up of leucocratic dikes that are confined to several zones with the dike orientation being different in each zone. One dike zone is about 7 km long, contains 34 separate dikes, and the belt and the dikes strike about east-west; the zone is in the northern part of the ore field. Another dike zone, to the east of the deposit, includes 44 dikes that strike NE-SW. Still another dike zone, containing 50 dikes, in the southeastern part of the ore region also strikes about east-west. In this southeasterly zone, there also are two granodiorite stocks that are 120 by 300 m in area. Yet another dike swarm is south of the ore area, can be traced for 1 km, and contains 20 dikes, also with east-west strikes. In all of the zones, the dikes have a nearly vertical dip. The dikes are composed of plagiogranite porphyries, syenite porphyries, and spherulitic syenite porphyries.

The structure has been produced by a combination of disjunctive folds and faults; the two major folds are the Muruntau anticline and the Southern syncline. The folds are subordinate and complicating folds on the northern flank of the larger Taskazgan anticline; the axial plane of this fold lies to the south of the ore field. These fold structures (with more or less east-west axes) are further complicated by additional crumpling that produced gentle bends in the beds. It is within one of the crumpled beds that the Muruntau deposit is contained. The beds in the area are essentially flat lying but this rather simple picture has been confused by steep folds that are near to being faulted along their crests close to large fractures. The three principal faults are the Southern, the Structural, and the Northwestern.

The Southern fault, not unreasonably, borders the ore field on the south, is a strike-slip fault, has a more or less east-west strike, and dips vertically or steeply to the south; the width of the crush zone associated with this fault is 5 to 7 m. In the eastern part of the ore field, however, the strike of this fault shifts to NE-SW and again reverts to an east-west direction; on the flank of the major anticline in the eastern part of the ore field, this fault again changes strike to NE-SW. At this end of the fault, it is accompanied by parallel feathering tectonic joints.

The Structural fault, of east-northeast strike and of the strike-slip type, actually is a series of closely spaced joints and crush zones that cut the central part of the ore deposit and extend beyond its margins. In its western segment, this fault dips to the north, but in the central and eastern portions, the dip is to the south. The crush zones of this fault are 2 to 10 m in width, and the fault has been filled with the tectonic clay gouge and breccias that have been cemented by carbonates.

Rather short tectonic joints have been formed in the ore area on a large scale; these joints have been confined mainly to those members composed of thin interstratified sandstones, siltstones, and shales, and most of them contain quartz veinlets. En echelon, "break-away" joints have been developed strongly in the center of the field; they are grouped into zones, all of which strike between N50°E to N55°E and dip between 70° and 80°SE; they contain, in many places, zones of disoriented crushing and have been filled with quartz of various grain sizes and with quartz-tourmaline breccias. Slip joints also are prominent and have more or less north-south and east-west strikes and dip steeply to the south, southeast, and northeast. The east-west (sublatitudinal) joints also have been filled with tectonic clay gouge and breccias and are cemented by carbonates; they normally cut conformable veinlets of the same composition. In the small sublatitudinal joints pyrite-arsenopyrite and quartz-pyrite-arsenopyrite veins and veinlets are localized. The more or less north-south (sub-meridional) slip joints were developed after coarse- and medium-grained quartz and contain essentially all the columnar quartz and calcite.

Borodaevskaya and Rozhkov consider that the deposit is a large stockwork, the plan-area of which is quite large, but mineralization is unevenly placed in this fractured body both in plan and in section. The ore grades so gradually from ore to waste that the boundaries between the two are determined only by sampling. Within the stockwork, zones of mineralization are complex in shape and include both stockwork veinlets and larger quartz veins. These zones of mineralization are located near to NE-SW-striking faults, and they are localized by east-west zone of subparallel "breakaway" and slip joints. The bulk of all the known gold is in segregations in three of these ore zones.

The shape and structure of the ore bodies in these ore zones is quite complex. They consist of a combination of steeply dipping and gently dipping quartz vein and veinlet zones. The stem or trunk portions of these zones have steep dips; outward from these trunks branches extend, and these branches are conformable with the stratification of those beds that are lithilogically favorable to the deposition of gold. The discordant portions of the ore zones, however, are the richest, the ore being in a series of parallel en echelon quartz veins and bands of thin quartz, quartz-sulfide, quartz-tourmaline and carbonate veinlets. In some parts of these zones, the carbonate and quartz-tourmaline bodies are brecciated and are larger than normal. The various varieties of veins and veinlets usually coalesce into zones of appreciable thickness that strike more or less east-west for several hundreds of meters; their extent along the dip also is considerable. The en echelon patterns of the veins also is shown down the dip, in which direction a vein or set of veins may die out to be replaced immediately down dip by another set in an en echelon arrangement to the first. This results in the ore zone being steadily displaced to the south as it is followed downward the acquiring a gentler dip, although the dip of the veins in the discordant zones averages between 60° and 70°S. The largest veins may be as much as 15 to 20 m wide where they assume lensoid bulges. The contacts of the veins in phyllites are quite sharp; in the sandstones and siltstones, the quartz veins are bordered by belts of veinlets, the contacts of which make gradual transitions into the stockworks. Where the rocks have been most sheared, the veins are most abundant, particularly where the veins are confined to stratigraphic layers; some veins and veinlets are cross-cutting, but these are in the minority. Some of the zones of discordant veins have been repeatedly fractured and the veins themselves crushed and cut by thin veinlets containing carbonate, sulfides, and networks of fine quartz-tourmaline veins. In such sectors, the veins dip steeply and usually contain the largest amounts of gold.

Where conformable vein zones or portions of such are close to the stem or trunk portions of the deposit, they contain appreciably less gold, but its distribution is much more complex. These veins are gold-bearing only where both discordant veinlets of quartz and sulfides are associated with bedded veinlets. The thickness of layer- and lens-like-shapes increases where they join with discordant vein zones.

The morphology of the vein zones has been controlled by the combination of the concordant and discordant elements of the vein pattern; the result is a most complicated structure. Such sulfide mineralization as there is is contained in the markedly more regular, usually thin, discordant veinlets; that is the veinlets are discordant to the stratigraphy but are regularly arranged in relation to each other.

As has been mentioned, the sulfides make up only 0.5 to 1.5 per cent of the total ore volume with the main vein mineral being quartz; potassium feldspar, biotite, calcite, tourmaline, and albite are known only in small quantities. The ore sulfides are mainly pyrite and arsenopyrite; locally and rarely, there are small segregations of scheelite, sphalerite, galena, bismuthinite, native bismuth, and microscopic amounts of other minerals of similar temperature and pressure of formation characteristics. In quartz veins, where gold is most common, it is unevenly disseminated as fine inclusions and veinlets or even locally small nests in association with sulides, rock fragments, or boundaries of quartz grains.

The fineness of the gold is reported by Borodaevskaya and Rozhkov to be between 880 and 910, an appreciable difference from the two fineness types stated by Glynn to be 900 and 750 (see above). There do not appear to be enough traces of any elements in the gold to make these worth recovering; that is the less than 1000 fine gold is not due so dominantly to silver that that element can be commercially recovered.

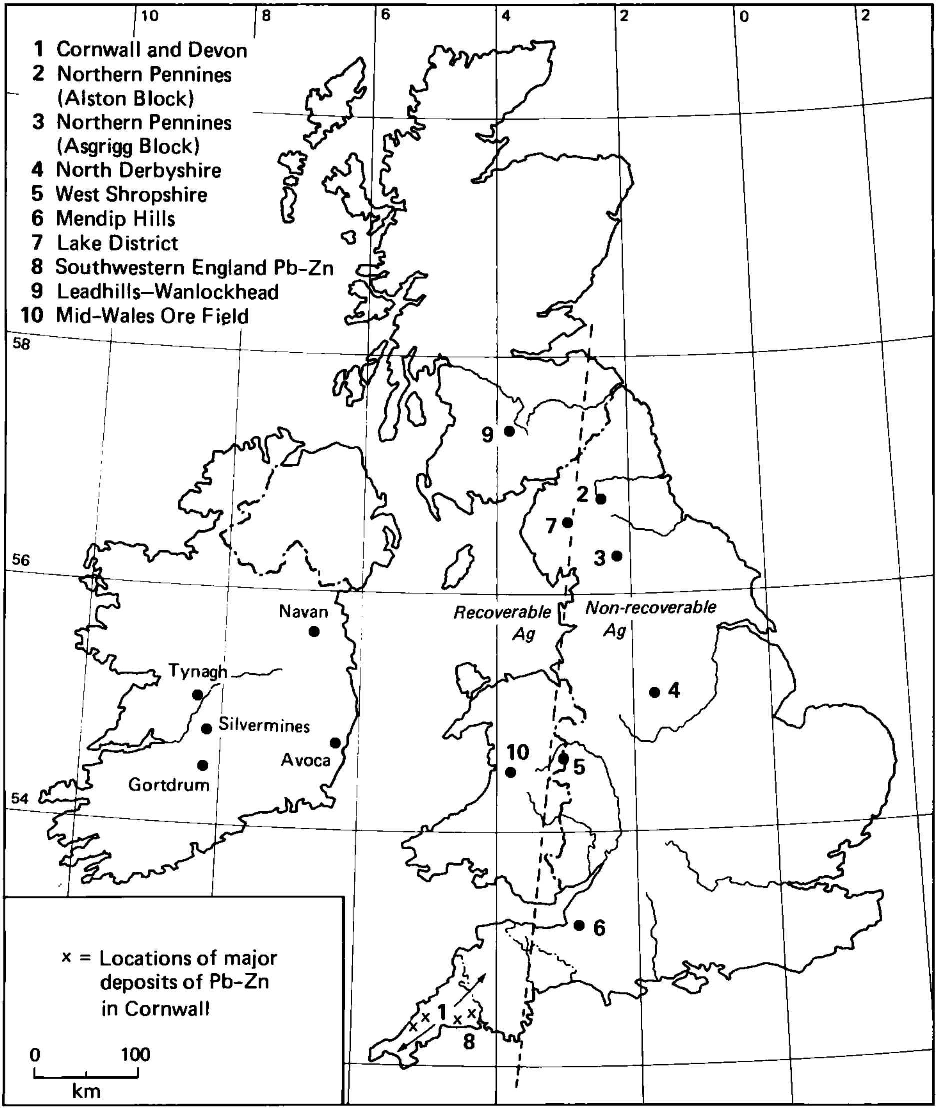
Within the ore zones, the regionally metamorphosed rocks into which the quartz-gold-sulfide ores were introduced have suffered contact metamorphism and hydrothermal replacement reactions that have considerably changed the character of the host rocks of the ores. The changes due to contact metamorphism include the formation of hornfelses through the development of newly formed biotite and actinolite, plus pyroxene (particularly diopside) in minor amounts. The production of hornfels coincides with the contours of the ore stockworks, but, in places at least, is more widely developed than the ore. The degree of hornfelsing increases with depth, suggesting to Smirnov's authors that this process may be related to solutions given off from a granitoid magma at depth. In addition to the hornfelsing, a complex series of hydrothermal replacement reactions have affected that rock in the ore zones; they mainly are the production of quartz-microcline and quartz-rich replacement bodies. The quartz-microcline replacements form selvages around gold-bearing quartz veins and veinlets for distances of between 2 and 3 cm. Later phases of mineralization are accompanied by quartz-albite and quartz alone in similar positions to those occupied by the quartz-microcline.

The mineralization has been divided, by a number of workers (N. Nesterova, Yu. Zarembo, and others) into a series of stages and phases. The first stage - quartzose - contains most of the vein quartz. This stage appears to be separated from the later ones by the injection of various dike rocks. The second stage - quartz-sulfide - is divided into three depositional phases: (1) columnar-comb quartz; (2) pyrite-arsenopyrite, and (3) pyrite-tourmaline. In the third stage was the gold-polymetallic, the one with rosette-stellate quartz and gold-polymetallic sulfides; subphases include gold-quartz, polymetallic sulfides, and carbonate. The authors believe that gold was deposited only at the end of ore formation, although these authors appear to believe that gold was repeatedly precipitated and re-precipitated during the lengthy hydrothermal development.

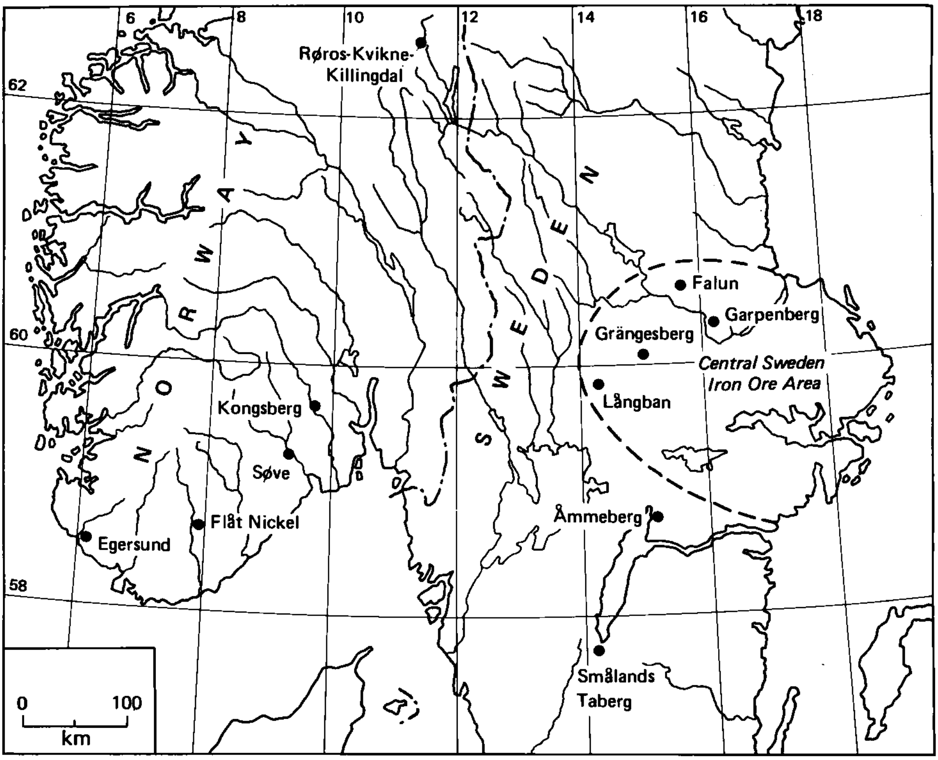
From the close association of gold with sulfides, particularly arsenopyrite, it would seem reasonable to consider the gold to have been emplaced at high temperatures, high enough at least for the gold to have been considered as hypothermal in temperature and pressure at the time of its deposition. The accessory and rare minerals are such as would be expected to have been deposited with gold under these conditions, and the ores are here classified as hypothermal-1.

If the statement at the beginning of Borodaevskaya and Rozhkov's paper in Smirnov's volumes II can be accepted as factual, then the development of the Muruntau deposit in a Hercynian geosyncline, apparently in the same general period as the Hercynian orogeny, suggests that the ores should be categorized as Late Paleozoic.

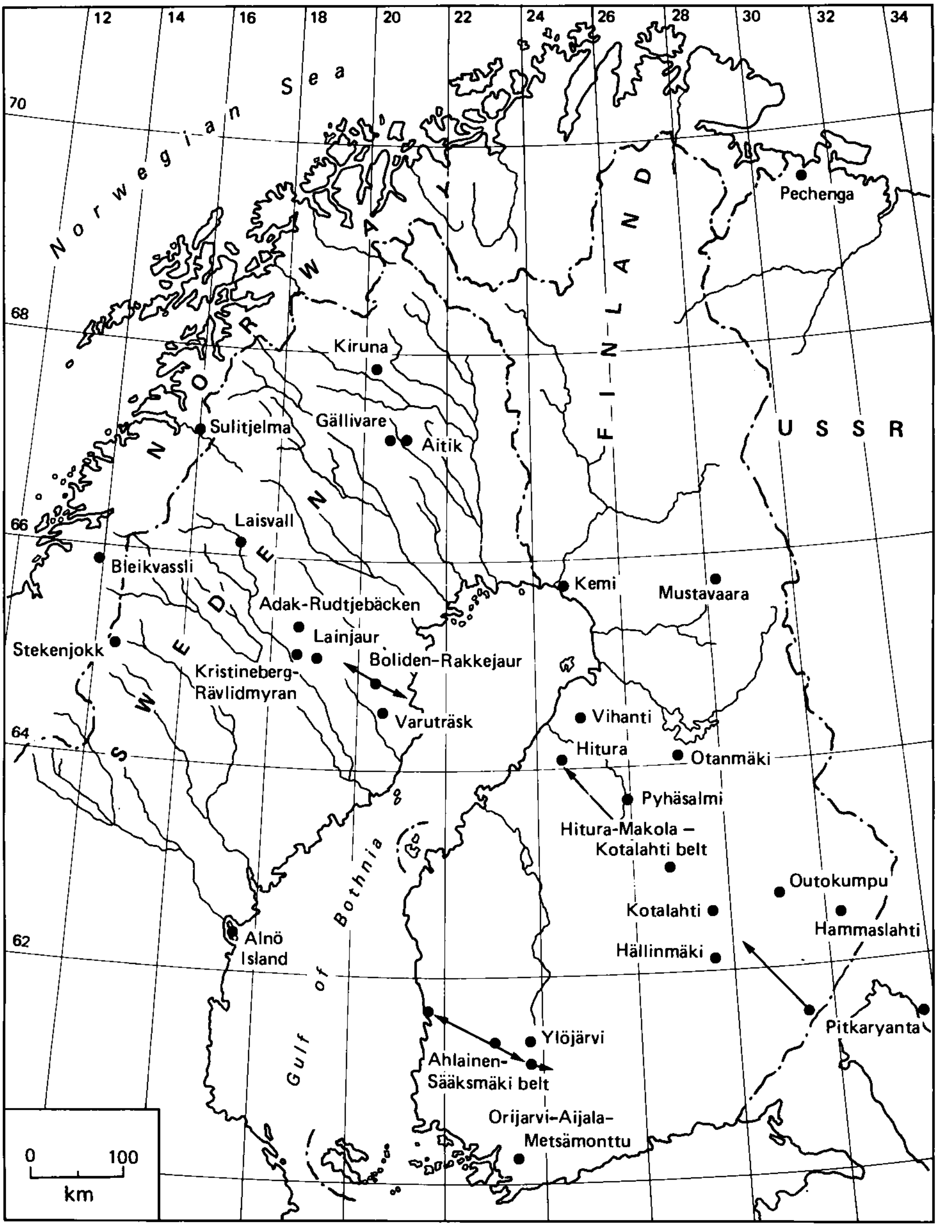
MAPS



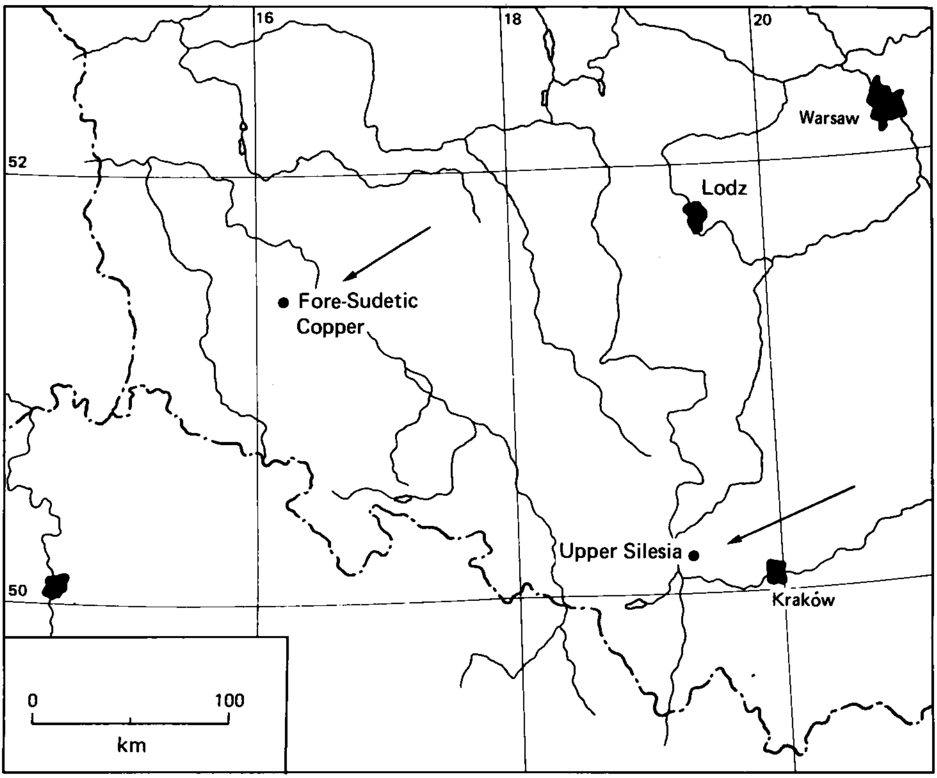
Great Britain and Ireland.



Southern Norway and Southern Sweden.



Northern Norway, Northern Sweden, Finland and Northwestern USSR.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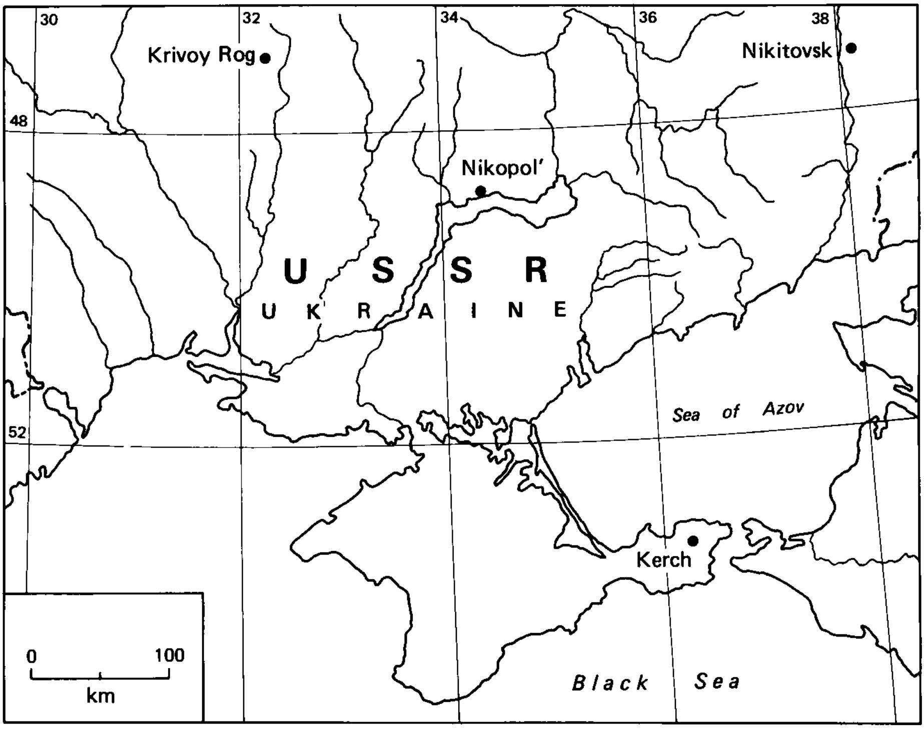
Poland.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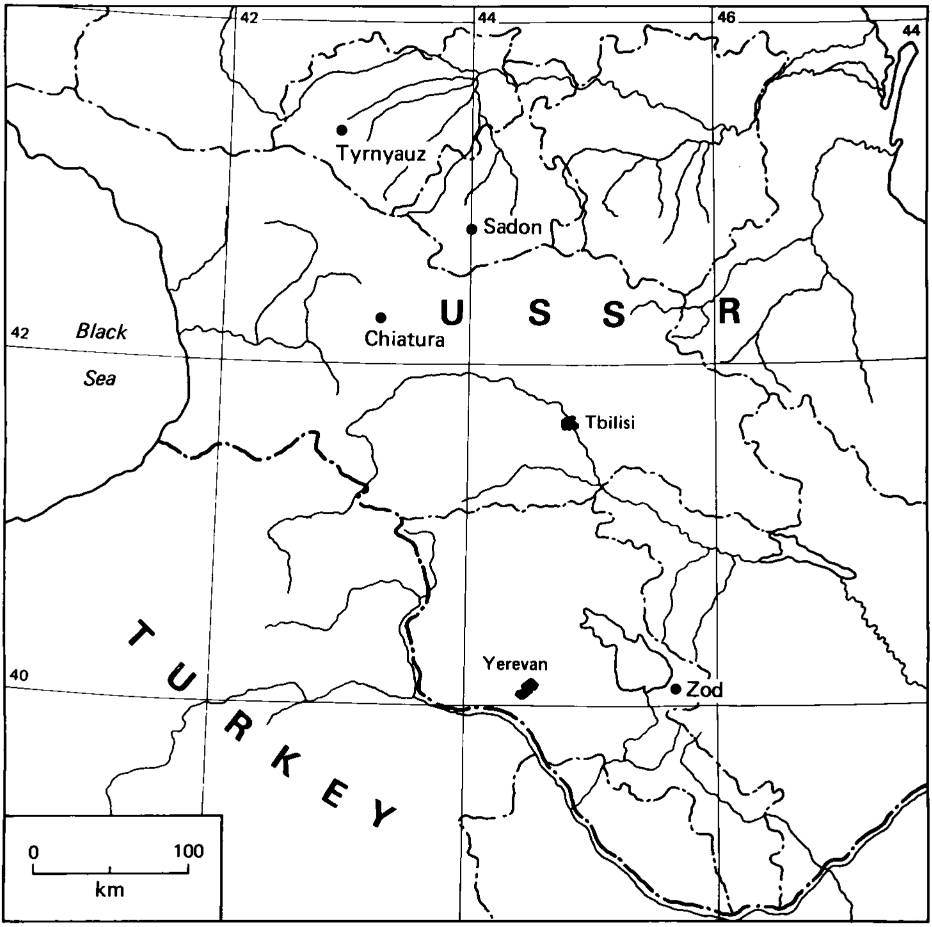
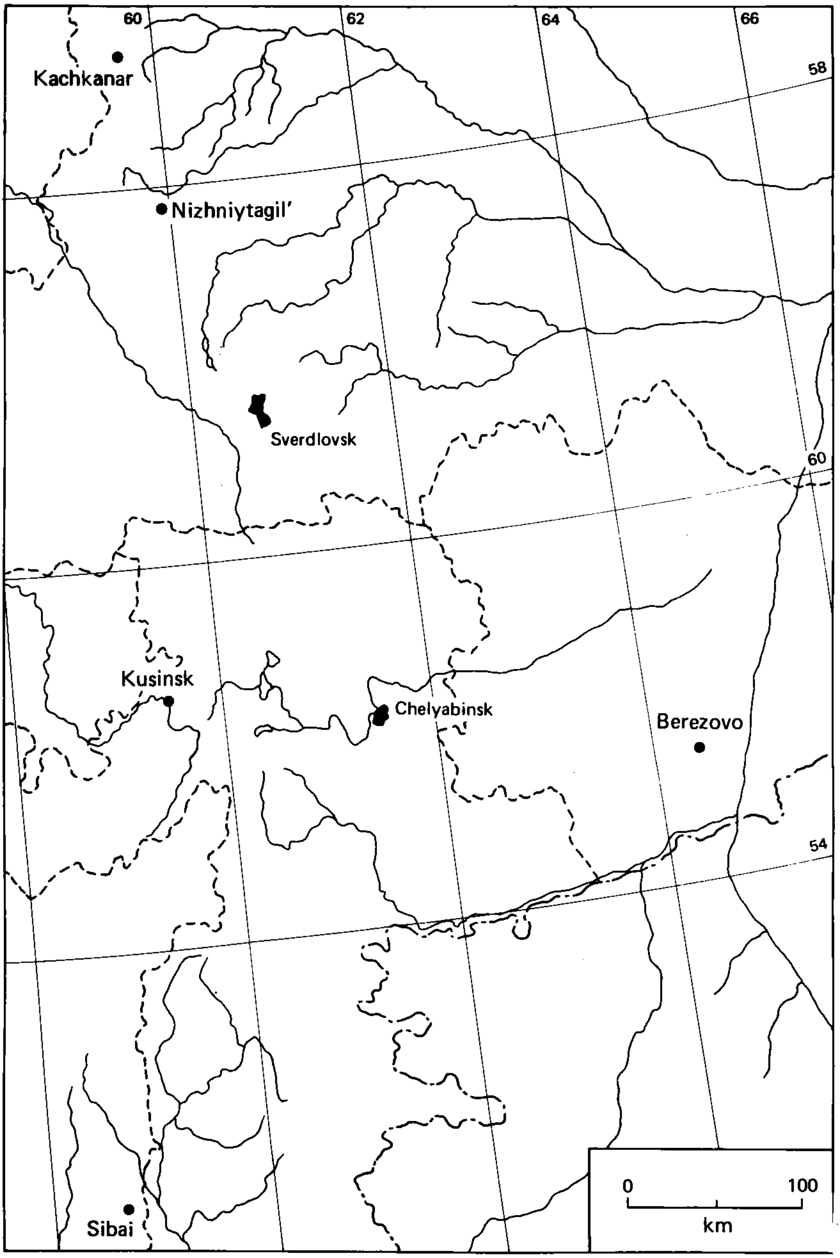
48
52
Krivoy Rog
Nikitovsk
Nikopol'
U S S R
U K R A I N E
Sea of Azov
Kerch
Black Sea
0
100
km

Ukraine.



Georgia, Armenia and Environs.



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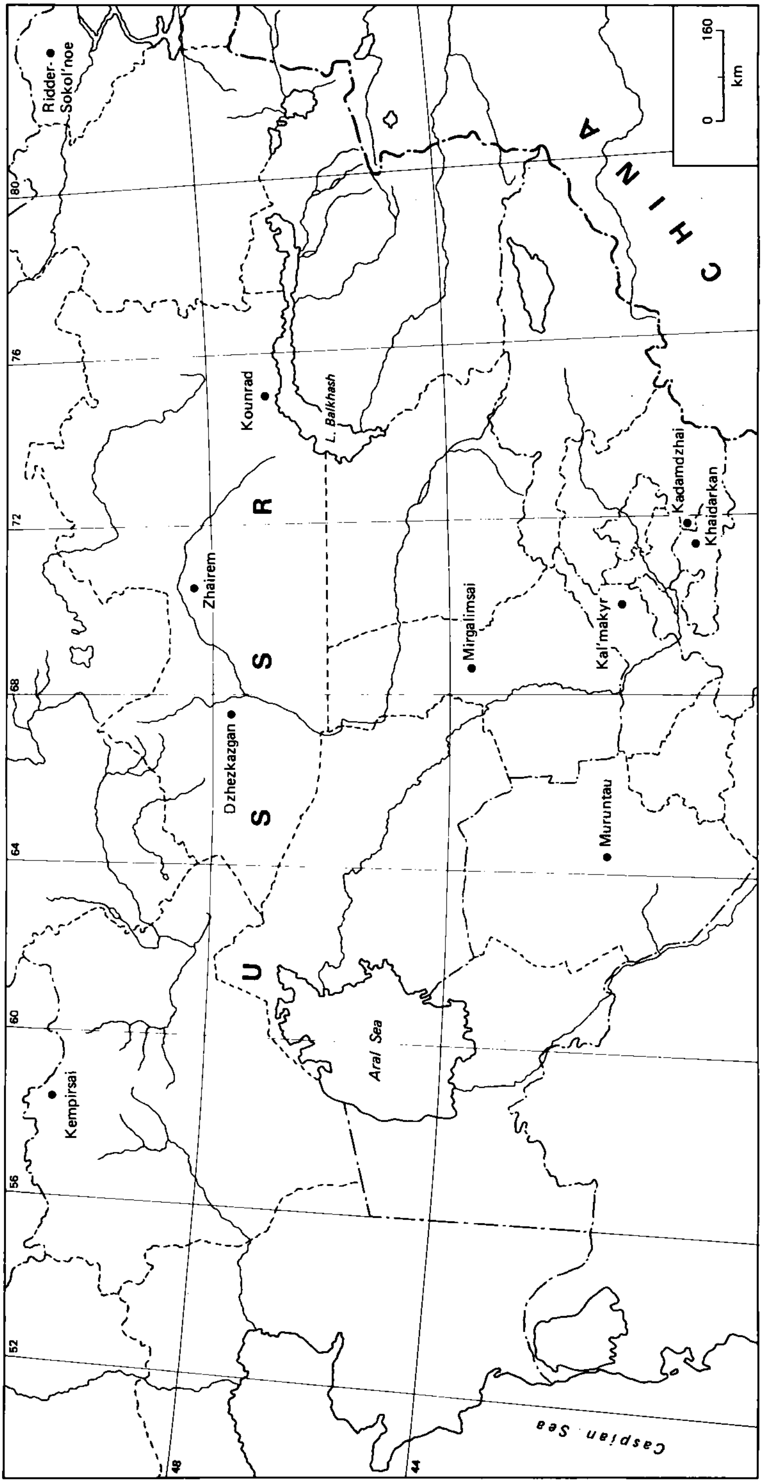
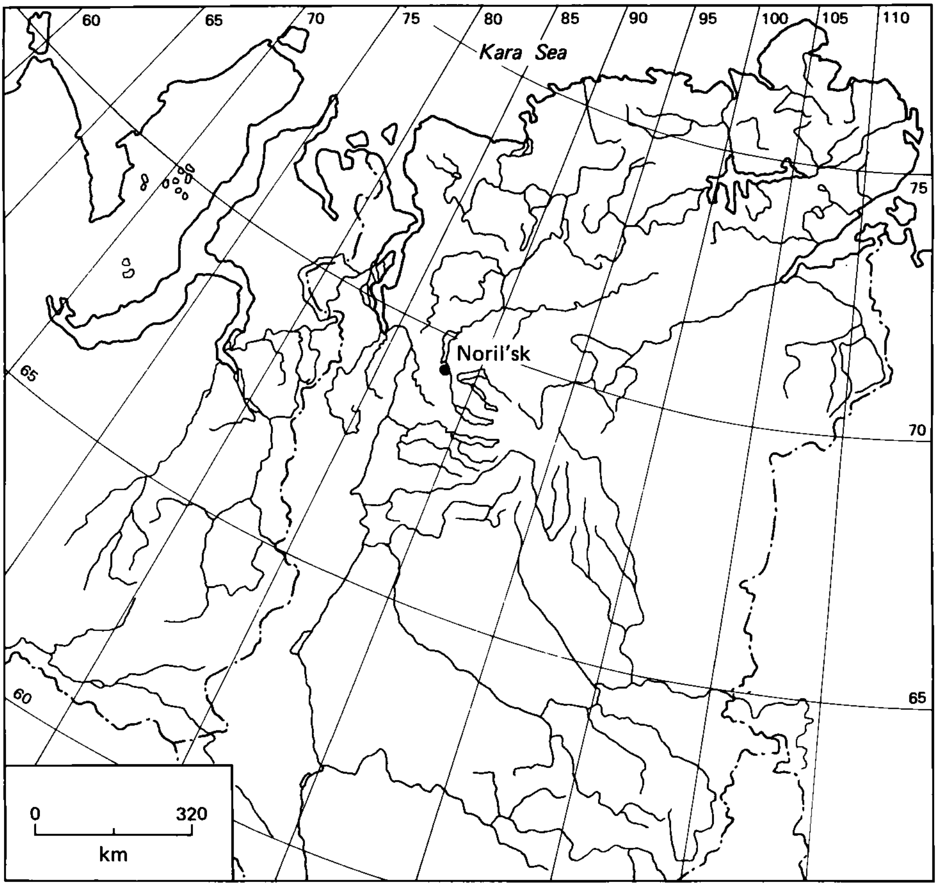
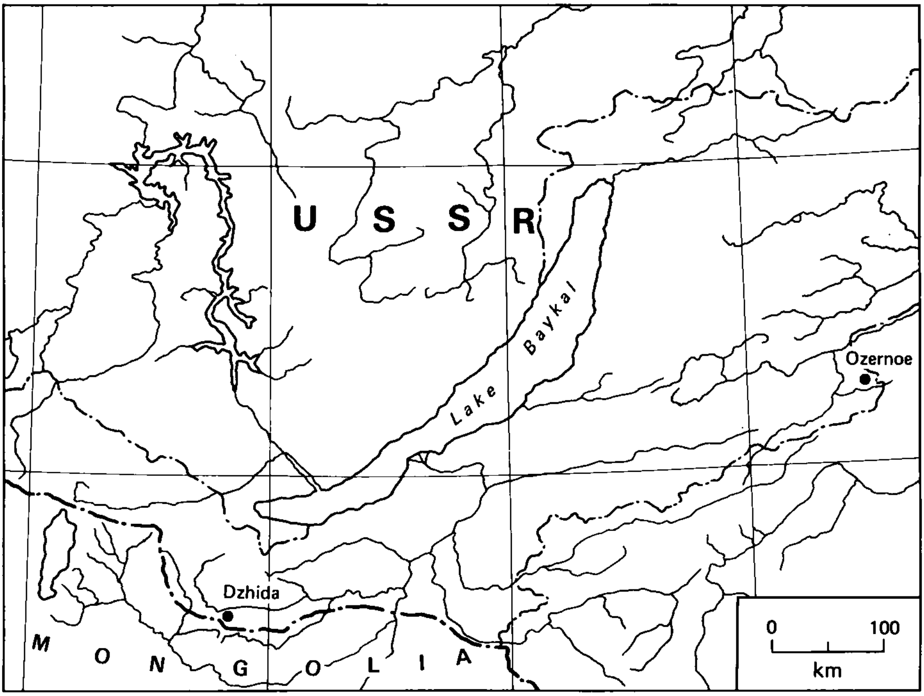


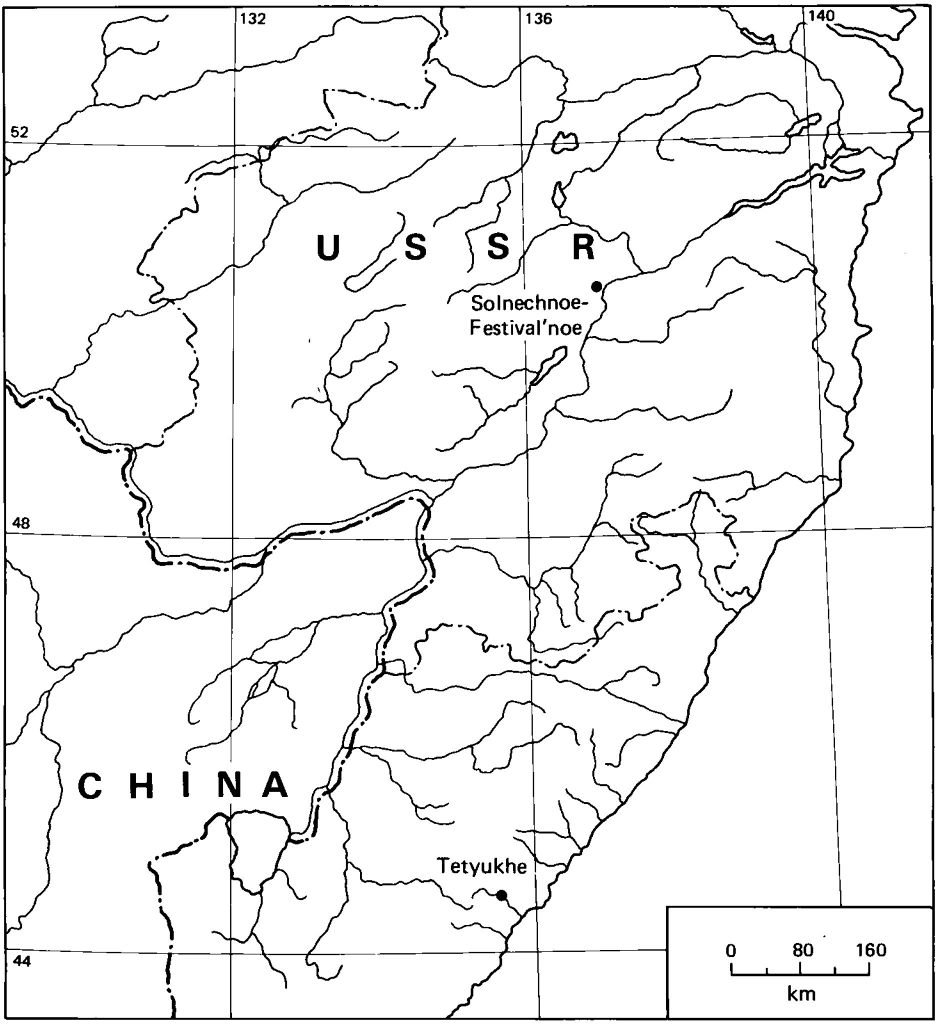
Kazakhstan, Uzbekistan, and Environs.



North-Central Siberia.



Buryat ASSR and Environs.



Far-Eastern Siberia.

INDICES

INDEX OF AUTHORS

Note This index contains the names of 902 authors or groups of authors. The actual numbers of authors included in the 84 bibliographies in this Volume is appreciably greater than 902 because all papers with more than two authors are listed under the name of the senior author, followed by the phase "and others." Also more than one author is included in the designations: "Anonymous," "Academy of Sciences Group," "Geological Staff, Outkumpu and Otanmäki," and "Working Group."

ALPHABETICAL INDEX OF DEPOSITS CITED OR DISCUSSED IN THIS VOLUME

This index records the location in this Volume of all deposits mentioned in the references or notes, whether they be mines, ore bodies, ore areas, ore segregations, districts, or even countries. In this Volume, 84 deposits, districts, or countries are considered in detail, and many other types of ore concentration are mentioned in little or some detail. If any one of the 84 ore areas given a thorough discussion appears under single designation, for example "Tynagh," only that name is placed in the index, followed by the number of the page on which its bibliography begins. If a deposit or district has a designation that hyphenates two names, for example, "Ridder-Sokol'noe," this designation is followed by the number of the page on which the bibliography of this ore district commences. Further, the name, "Sokol'noe," also is given a place in the index, followed by the number of the page on which the discussion of Sokol'noe begins. If the deposit name should include three specific ore locations, all three being separated by hyphens, an example is, "Orijärvi-Aijala-Metsämonttu." Each of these three names will have an individual entry in this index.

Any actual mine, active or abandonned, that is not important enough or for which sufficient literature is lacking for full treatment, nevertheless is listed by name, followed by the designation the one of the 84 descriptions to which it is connected. For example, the Stora Malmberget mine, located in the Gällivare portion of the Kiruna-Gällivare district, is listed, of course, under "Stora Malmberget," followed by the name Gällivare and the pages in the Kiruna-Gällivare bibliography and notes where it is mentioned or discussed.

If a mineralized rock volume in a given district is described by some other designation than "mine," for example, "ore body," or "ore area," or ore segregation," or whatever, the name of the ore concentration is followed by the type of ore concentration in parentheses, for example, "Almalyk (District), Kal'makyr."

In some instances, a mine may have been named for a person. For example, one of the Krivoy Rog mines has been named for V. I. Lenin. In this index, the mine is listed as "Lenin, V. I. (Mine), Krivoy Rog."

All names given here have been arranged according to the English alphabet with letters in the Scandinavian alphabets such as Å, Ä, Ö, or Ø, being alphabetized as if no diacritical marks were involved. This will be confusing to a Scandinavian for a few minutes, but he or she should readily adjust.

The differences between the English arrangement and that of the Russian alphabet are much more extreme, for the Russian alphabet involves a much different order for letters similar to English and also contains nearly 20 characters that are not present in any alphabet using Roman letters exclusively. Thus, several of these Russian letters are transliterated by two or more Roman letters; for example, the combinations zh, kh, ts, ch, sh, and shch are used, not to mention mute hard and soft signs, the former being transliterated as '; other Russian letters may appear as iu (yu) and ia (ya) and are the last two letters in the Russian alphabet; they will appear under the letter "y" in this index. Thus, any Russian who tries to use this index will need a little time to become accustomed to the place to look for some names for ore deposits in his country.

INDEX OF DEPOSITS ACCORDING TO METALS AND MINERALS PRODUCED

The index that follows contains only the economically and/or geologically important metals and minerals that are (or have been) recovered from the deposits included in this Volume.

Only the name under which a given deposit appears in the Table of Contents is used in this index. Other names associated with any particular deposit can be found in the Alphabetical Index of Deposits.

Each deposit listed in this index, no matter how many times it may occur here, is followed by the number of the first page on which the deposit in question is to be found.

The numbers of deposits containing the various metals and minerals included in this Volume are as follows:

Andalusite 1
Antimony 2
Apatite 2
Arsenic 3
Barite 11
Cesium 1
Chromite 2
Cobalt 4
Columbium 2
Copper 37
Feldspar 1
Fluorspar 3
Gold 15
Iron 15
Lead 27
Lithium 1
Manganese 4
Mercury 2
Molybdenum 4
Nickel 6
Platinum Metals 3
Pyrite 13
Rhenium 2
Rubidium 1
Silver 20
Tellurium 1
Tin 2
Titanium 4
Tungsten 6
Vanadium 6
Witherite 1
Zinc 32

This table cannot, of course, be totaled because many of the deposits are mined for two or more of the commodities listed.

INDEX OF DEPOSITS ACCORDING TO AGE OF MINERALIZATION

INDEX OF DEPOSITS ACCORDING TO THE MODIFIED LINDGREN CLASSIFICATION

In this index is given a list of the deposits discussed in detail in this Volume; these are arranged under the appropriate heading in the Lindgren Classification. Of the 65 deposits included completely or partially in the hydrothermal category, 32 fall into one of the Lindgren subdivisions, whereas 33 of them were formed through two or more of the Lindgren intensity ranges. For example, the Cornwall and Devon Sn-W-Cu deposits were developed over the gamut of ranges from Hypothermal-1 to Leptothermal and the Kal'makyr deposits all the way from Hypothermal-1 to Telethermal. These two deposits are the only ones in this Volume with such wide varieties of conditions of formation. Fourteen deposits, from Adak-Rudtjebäcken to Sulitijelma, were produced through the Hypothermal-1 to Mesothermal ranges, and 6 were formed through the Hypothermal-2 to Mesothermal ranges, whereas only 11 were formed under hypothermal conditions alone and only 8 solely under mesothermal ones. It seems, therefore, necessary to list each hydrothermal deposit under the particular intensity range or ranges through which it was produced.

This arrangement, given in the Table that follows, shows most clearly the variations in intensity limits through which ores can be formed in hydrothermal deposits.

In this Table, no attempt is made to provide a similar breakdown to that used for hydrothermal deposits for those formed under magmatic conditions. This is done because deposits placed in the four classes of magmatic deposits (see the Modified Lindgren Classification in Volume I of this series) normally are not part of a continuum as often is true of hydrothermal deposits. The same reasoning is used in the presentation in this Table of Deposits formed as primary sediments that may have derived their contents of economically valuable materials from surface or volcanic sources or both.

*Lindgren Classification Category**	*Number of Examples#*	
Magmatic-1a	2	(6)
Magmatic-1b	2	(4)
Magmatic-2a	5	(3)
Magmatic-2b	3	(7)
Magmatic-3a	3	(9)
Magmatic-3b	8	(9)
Magmatic-4	2	(6)
Sub Total	25	(44)

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* For a detailed discussion of the Modified Lindgren Classification, see Volume I of this series. A few changes in the classification were added in Volume II.

The numbers in parentheses that follow the numbers of examples of deposits in Northern Europe and the USSR are those given for the Western Hemisphere in Volume I of this series.

Hypothermal-1	11		(28)
Hypothermal-2	8		(13)
Hypothermal-1 to Mesothermal	14		(21)
Hypothermal-2 to Mesothermal	6		(14)
Hypothermal-1 to Leptothermal	1		(6)
Hypothermal-2 to Leptothermal	0		(2)
Hypothermal-1 to Telethermal	1		(2)
Hypothermal-2 to Telethermal	0		(1)
Mesothermal	6		(29)
Mesothermal to Leptothermal	7		(22)
Mesothermal to Telethermal	1		(5)
Leptothermal	1		(9)
Leptothermal to Telethermal	3		(2)
Telethermal	6		(13)
Sub Total	65		(251)
Xenothermal	0		(1)
Xenothermal to Kryptothermal	1		(3)
Xenothermal to Epithermal	0		(3)
Kryptothermal	0		(1)
Kryptothermal to Epithermal	1		(1)
Epithermal	3		(17)
	5		(26)
Sedimentary-A1a	4	(2,?)	(7)
Sedimentary-A2	2	(?)	(0)
Sedimentary-A3	6	(2,?)	(3)
Sedimentary-A4	2	(?)	(0)
Placers 1B and 1C	2		(1)
Ground Water-B2	2		(29)
Sub Total	18	(8?)	(41)
Grand Total	131		(362)

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