Symposium on Basins and Geosynclines of the Canadian Shield

Edited by
A. J. BAER

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BASINS AND GEOSYNCLINES OF THE CANADIAN SHIELD

A collection of papers presented at a symposium held in Camsell Hall, Ottawa, March 17-18, 1970.

Edited by A. J. Baer

DEPARTMENT OF ENERGY, MINES AND RESOURCES

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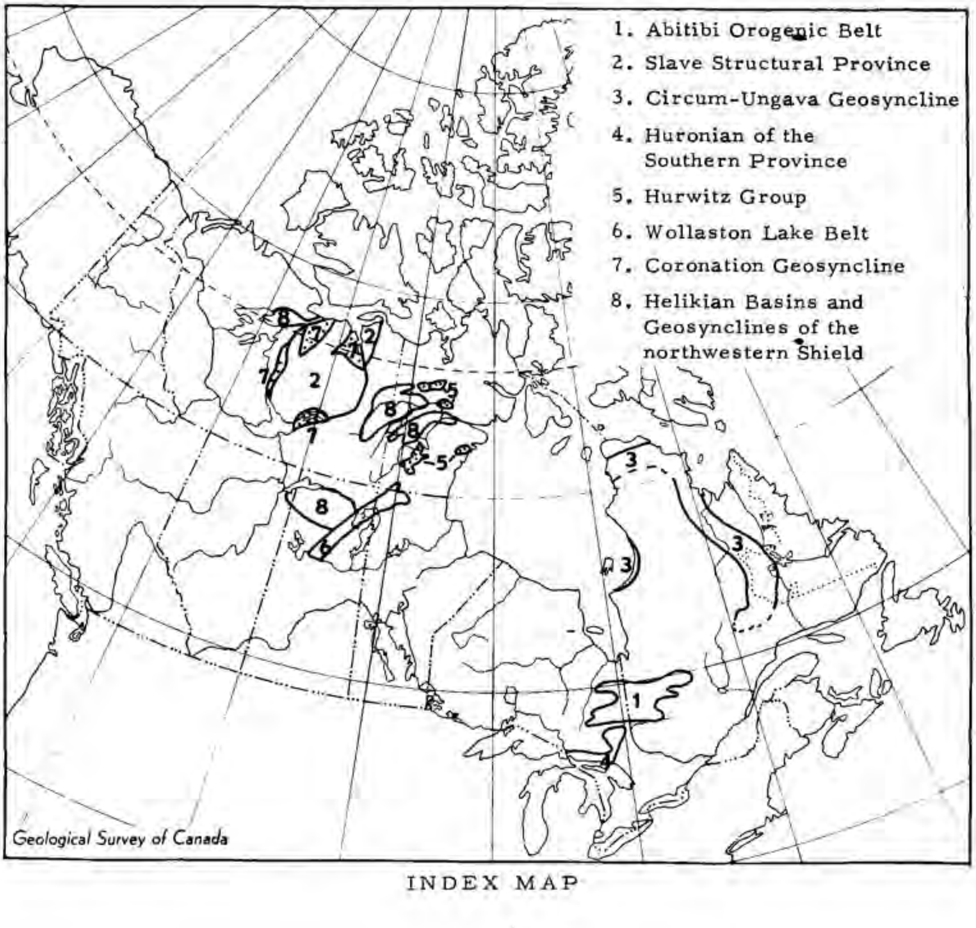
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INTRODUCTION

The Canadian Shield, that largest and best exposed Precambrian block and so rich in lithic and tectonic elements, contains a vital record of early crustal events - testimony to Earth's long and intricate evolution. By 1972 the Canadian Shield will have been completely mapped at least on a reconnaissance scale. This remarkable achievement is due mainly to the continuing efforts, extending over several generations, by members of the Geological Survey of Canada of which the Precambrian Subdivision represents one of the largest integrated working groups of Precambrian geologists in the world. From primitive Archean volcanic piles formed in small basins upon thin, supple crust, through younger iron-bearing sediments deposited in long, linear circum-cratonic troughs, to still younger fillings of broad, extensive basins on thick stable continental crust, the record is traceable throughout the length and breadth of this remarkable Precambrian Shield.

The effective completion of reconnaissance mapping provides a logical basis for analysis and synthesis of available data. Establishment of sound tectonic and sedimentary models will provide meaningful perspective, the basis of significant discussions, and a guide to future detailed studies.

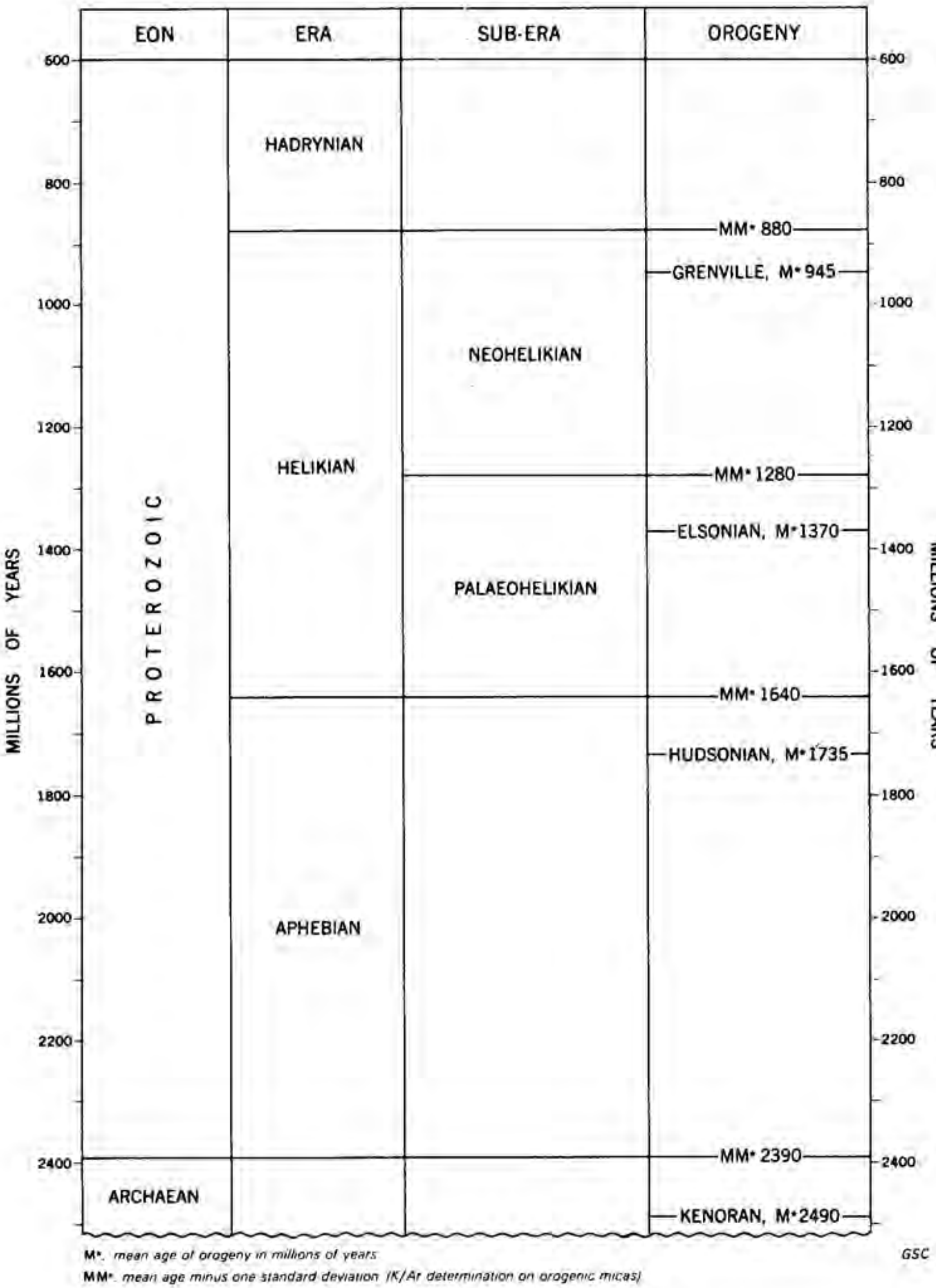
Within this context an organizing committee of the Precambrian Subdivision of the Geological Survey of Canada was formed in 1968 to select a suitable Precambrian topic, implement a two-day workshop and prepare the results for publication. The committee was composed of A.M. Goodwin (Chairman), A.J. Baer, M.J. Frarey, J.A. Fraser and J.C. McGlynn.

This symposium on Basins and Geosynclines of the Canadian Shield was held in Ottawa on March 17-18, 1970. Eight principal papers dealing with two Archean and six Proterozoic tectonic units formed the basis of the workshop. They culminated in the summation and appraisal prepared by a well-known expert of the Precambrian, F.J. Pettijohn. An invited audience of about one hundred earth scientists representing the governments of five Canadian provinces, fourteen Canadian universities, and several universities and governments in the northern United States, provided stimulating comments and discussion of the papers. The nine component papers were edited by A.J. Baer and submitted to the Geological Survey of Canada for publication, thereby providing a permanent record of this workshop.

Sincere thanks are due all those who contributed to the workshop and to this publication especially Dr. Y.O. Fortier, Director of the Geological Survey of Canada who officially opened the meeting and authorized publication of the proceedings by the Survey. J.A. Fraser supervised all local arrangements; A.J. Baer acted as editor; J.A. Donaldson arranged all the use of Carleton University Faculty Club; and Mrs. L.R. Mahoney prepared the manuscript for publication. Finally I personally thank fellow members of the organizing committee for their generous co-operation and efforts.

The Organizing Committee and all others who have contributed to this volume trust it will both provide insight into the complexities and fascinations of Canadian Precambrian geology and serve to stimulate detailed studies in this and other Precambrian shields of the world.

A.M. Goodwin,
Chairman of the Organizing Committee.



Precambrian time-scale for the Canadian Shield
from Geol. Surv. Can., Paper 67-2 Part A, 1968.

THE ABITIBI OROGENIC BELT

A.M. Goodwin, Department of Geology,
University of Toronto, Toronto, Ontario.

and

R.H. Ridler, Department of Geology,
University of Western Ontario, London, Ontario.
Present address: Geological Survey of Canada, Ottawa.

Abstract

This east-trending tectonic unit, some 500 by 150 miles in dimensions, is the largest single continuous Archean greenstone belt in the Canadian Shield. It is a characteristic Archean orogenic composite belt featuring mafic to felsic volcanics with coeval intrusions, volcanic sediments, both clastic and chemical including banded iron formation, and several large granitic batholiths. Metallogenic patterns conform to lithic distributions. Low- to medium-rank greenschist facies prevail. Supracrustal rocks have been isoclinally folded about east-trending, undulating axes resulting in substantial lithofacies compression. The belt is bounded north and south by granitic-metasedimentary crystalline terrains; it is truncated abruptly east and west by Grenville and Kapuskasing crystalline rocks, both products of younger Precambrian events. Thus the present belt represents only part of an original Archean tectonic entity.

A number of well-recognized mafic to felsic volcanic centres with intercalated volcaniclastics, iron formation and igneous intrusions are present. Despite common features each centre constitutes a semi-independent lithic assemblage of limited stratigraphic continuity. The centres may be confidently ascribed to processes of igneous differentiation and eruption largely by way of central vents. Three main stages of igneous eruption are recognized in the belt. The total time span was probably several hundred thousand years. Major geochronological problems remain.

Recognized mafic to felsic volcanic centres and accompanying clastics with iron formation are concentrated in two main east-trending bands, each 50 miles wide, which respectively cross the northern and southern parts of the belt. In contrast, the intervening or median part of the belt, also 50 miles wide, is underlain, as known, by uniform tholeiitic basalt, fine-grained clastics, and major granitic batholiths.

On this basis, the Abitibi belt is viewed as a remnant of a bilaterally symmetrical intra-tectonic orogen rather than a conventional asymmetrical continental-oceanic tectonic interface (e.g. island arc). Crustal thinning and other orogenic activities may be attributed either to conventional geosynclinal downsinking of thin, supple Archean crust or to spreading apart of Archean crustal blocks in the manner of present-day ocean floor spreading.

INTRODUCTION

Abitibi belt, 475 miles long by 125 miles wide, is the largest continuous Archean (older than 2,500 million years) greenstone belt in the Canadian Shield. Located in the southeastern part of the Superior tectonic province this east-trending belt is truncated east and west by northeast-trending crystalline rocks of Grenville province and Kapuskasing subprovince respectively. The present belt is part only of an originally longer Archean orogen which in Archean time included now-crystalline Grenville rocks immediately on the east and, likely, orogenic rocks of Wawa belt to the west.

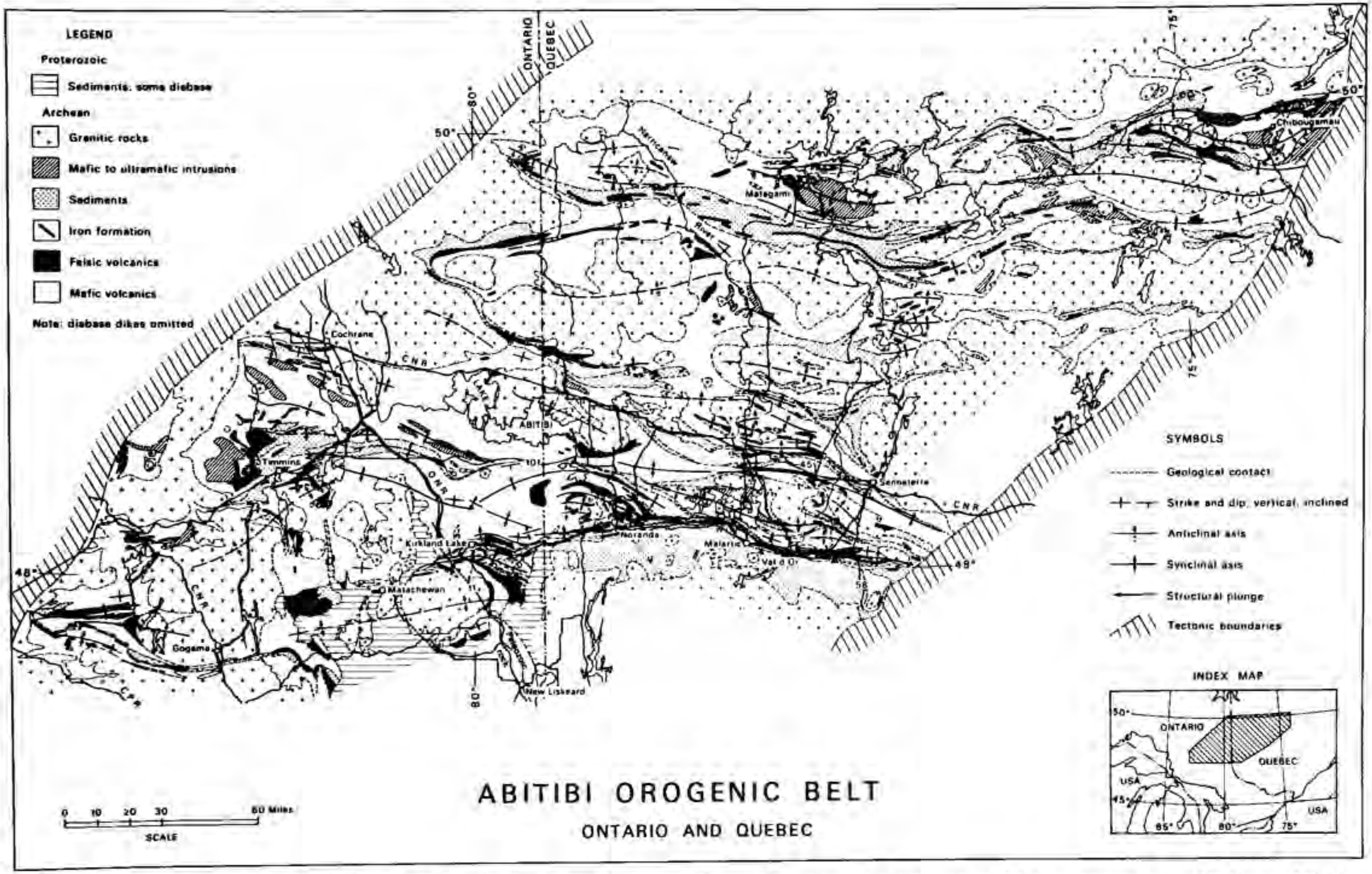


Figure 1. Geologic map of Abitibi orogenic belt. Omitted from the map for cartographic reasons are diabase dikes and some thin pyritiferous carbonaceous (± chert) zones in the Noranda-Val d'Or-Senneterre area. The latter are shown in Figure 2 as sulphide facies iron formation.

Abitibi belt has been geologically studied and explored for more than sixty years. Complete regional air-photo, aeromagnetic and gravity as well as local E.M. surveys are available. Many parts have been geologically mapped in detail. Sustained mineral exploration and mine development over the years have provided a wealth of data there being at least one hundred and fifty producers (past and present) including Au-Ag, Cu-Zn-Au-Ag, Ni-Cu, Fe, Mo-Bi, Li and asbestos deposits. However, extensive drift-covered tracts especially north of the main CN railway obscure many key geologic relationships. Despite the attention received, the major geologic problems including those of stratigraphic correlation and tectonic construction remain largely unresolved. Therefore additional studies, regional and detailed, direct and indirect, will be required for many decades to come.

In common with most greenstone belts of the Canadian Shield the geologic record of this belt doubtless includes a long and complex history of Archean events. So far, age-dating has not penetrated significantly the radiometric barrier imposed by the Kenoran event (at 2,500 million years) (Wanless *et al.*, 1968; see also Roscoe, 1965). This paper presents a brief integrated statement on relationships pertaining to tectonic evolution of this important primitive crustal unit. A principal purpose is to stimulate similar analysis of other Archean orogens in this and other Precambrian Shields of the world.

GENERAL GEOLOGY

The Abitibi lithic assemblage (Fig. 1) is characteristic of many Archean greenstone belts of the Canadian Shield. Older supracrustal and igneous rocks, now substantially deformed and of common greenschist facies, have been intruded by syntectonic to post-orogenic felsic to mafic intrusions ranging up to those of large batholithic dimensions. Numerous younger Precambrian diabase dikes, mostly of the Matchewan swarm, transect. Several tongues of flat-lying Proterozoic rocks and one of Paleozoic rocks protrude from the south. Pleistocene glaciation resulted in both clean-scouring of thick, steeply inclined Archean stratigraphic sections and accumulation of thick glacio-fluvial and lacustrine deposits which obscure bedrock relations particularly in northern parts of the region.

Volcanic Rocks

A variety of volcanic rocks is present. Mafic to felsic flows and pyroclastics of calc-alkaline chemical affinity, all representative of the orogenic suite (Baragar and Goodwin, 1969) greatly predominate. Some alkaline volcanic rocks are present at Kirkland Lake. Basalt flows and associated gabbroic intrusions are common in the lower parts of volcanic assemblages. Andesitic flows and pyroclastics are intercalated with basalt and characteristically increase in proportion upwards. Felsic rocks of dacite to rhyolite composition are generally present in upper stratigraphic parts. Most volcanic assemblages in the region display a generalized mafic to felsic compositional sequence of this type. In some areas the sequence is repeated in whole or in part.

Mafic volcanic rocks comprising basalt and andesite in that order of abundance, are the predominant supracrustal rocks of the region. They are broadly distributed particularly in the central and westen parts. Much remains to be learned about their petrochemistry. Thick mafic (i.e. basalt-andesite) accumulations exceeding 40,000 feet thick are present in Blake River group west of Noranda, and in Skead group south of Kirkland Lake. Similar thicknesses of mafic rocks may be present elsewhere in the region. In the Blake River group a lower zone of low-Al_2O_3 mafic volcanics, approximately 20,000 feet thick, is conformably overlain by equivalent thicknesses of high-Al_2O_3 mafic effusives (Baragar, 1968; Baragar and Goodwin, 1969). Similar low- and high-Al_2O_3 basalt and andesite are present in Swayze area to the west (Goodwin, 1968). In the Skead group south of Kirkland Lake a thick mafic assemblage approximately 28,000 feet thick which includes basanite lava flows exhibits significant upward increase in K_2O content (Ridler, 1970). Volcanic piles north and south of Timmins include tholeiitic basalt and andesite. Preliminary results of a reconnaissance geochemical cross-section of the volcanic rocks between Cadillac, Malartic area in the south and Matagami area in the north indicate a preponderance of basaltic rocks except for the felsic rocks shown in Figure 1 (J. Descarreaux, pers. comm.). Sparse chemical analyses of mafic volcanic rocks in the Matagami area suggest the presence of normal tholeiitic pillow basalt and andesite. In the Chibougamau area to the northeast, low-K_2O tholeiites including basaltic rocks with very low K_2O, high Na_2O and intermediate $K/_{Rb}$ ratios are present (Gunn, 1969).

Mafic lava flows are commonly 50 to 100 feet thick but flows up to 800 feet thick are known. Pillow forms, palagonite, variolites, amygdules and hyaloclastites are widespread and abundant thereby indicating prevailing subaqueous accumulation of the mafic lava. Primary textures are, on the whole, remarkably well preserved (Moorhouse, 1970).

Overlying the great thicknesses of mafic volcanic rocks are numerous concentrations of felsic volcanic rocks, each concentration characteristically representing a felsic eruptive centre (Fig. 1). Principal concentrations occur at Val d'Or, Noranda, Kirkland Lake, Timmins, Matagami and Chibougamau. Rocks of dacitic composition are most common in the felsic concentrations. Rhyolite is present locally as, for example, at Noranda, Timmins, Swayze, Joutel and Matagami but is absent in other centres, such as, for example, Kirkland Lake. Pyroclastic forms are very common in felsic concentrations of Swayze, Val d'Or, Kirkland Lake, Timmins, Matagami and Chibougamau areas. Massive rhyolite lava flows intercalated with andesite flows are common at Noranda. Subaqueous ashflows recently identified in Noranda area (R.S. Fiske, pers. comm.) may be widespread in other volcanic piles of the region. Trachytic flows and pyroclastics are present at Kirkland Lake.

In addition to specific felsic concentrations, the regional volcanic pile contains many thin discontinuous felsic tuff zones, some notably spherulitic, which are attributed to widespread wind and/or water distribution. Such zones are potentially valuable marker-horizons particularly where intercalated within thick uniform mafic flow accumulations.

Although mineralogically altered, felsic volcanic rocks exhibit a wide range of internal textures including spherules, phenocrysts, vesicles, perlite, pumice and shards, all commonly in excellent state of preservation (Moorhouse, 1970).

Sediments

Clastic zones range from thin discontinuous units to broad regional, east-trending belts of which the principal are situated south of Noranda-Val d'Or, east of Kirkland Lake, near Timmins, northeast of Lake Abitibi, south and southwest of Matagami and west of Chibougamau (Fig. 1). Sediments characteristically occupy higher stratigraphic positions. Most sediments are structurally conformable with associated layered rocks. However, local unconformities have been established at Kirkland Lake and Timmins.

Sediments of the "poured-in" turbidite association predominate thereby suggesting rapid accumulation in tectonically unstable environments. Two principal facies are identical: volcanogenic and flyschoid. The Timiskaming facies, a variation, is present locally. Rocks in the volcanogenic facies, ranging up to 10,000 feet thick include greywacke, shale, lithic sandstone, conglomerate, breccia and iron formation. Identifiable clastic components compare closely in composition with nearby volcanic rocks and have obviously been derived in large part from them by rapid mechanical erosion and subaqueous deposition in nearby troughs and basins. Chaotic textures and polymictic unsorted materials are distinctive features. This facies which is characteristic of strongly active tectonic zones, typically displays intricate soft-sediment deformational structures. Graded bedding and abrupt facies changes are common. Iron formation is present. Carbonaceous and pyritiferous zones are intercalated with finer grained clastic phases.

A prominent belt of more regularly bedded flyschoid facies (Pontiac group) lies south of Noranda-Malartic-Val d'Or. It comprises rhythmically bedded greywacke-argillite sequences of uniform construction and composition which lack marked lateral facies changes. According to J. Holubec (in preparation) the facies is 7,000 to 8,000 feet thick and was developed in a comparatively stable tectonic environment in proximity to a stable crustal domain to the south.

A variant facies, the Timiskaming, is present in narrow, east-trending zones near Kirkland Lake-Noranda and Timmins. This facies is characterized by coarse-grained detritus including conglomerate, abrupt lateral facies changes and erosional unconformities. It occupies narrow zones of pronounced tectonic instability which have been attributed to the local development of steep slopes (J. Holubec). At Kirkland Lake the facies is associated with a unique assemblage of trachytic flows and pyroclastics.

Iron formation is widely distributed (Figs. 1 and 2). Principal bands occur in Swayze, south of Timmins and Kirkland Lake-Val d'Or areas in the south and within a broad east-trending belt south of Matagami in the north. Local bands are present at Chibougamau, in the Barraute-Quevillon area of the central-eastern part and elsewhere. The iron formations are important stratigraphic elements and the closest approach to marker horizons in the regional assemblage. Their continuity across broad parts of the region demonstrates the presence of large extensive basins of deposition in Abitibi time.

Iron formations of the region are classified as oxide, carbonate or sulphide facies depending on the predominant iron mineral. The three facies are transitional across the region (Fig. 2). Oxide (magnetite) facies is common in Swayze, south Timmins, Kirkland Lake and Val d'Or areas in the south and in the principal belt south of Matagami in the north. Generally, oxide facies is transitional eastward and northeastward through narrow carbonate (siderite-ankerite-dolomite) zones to sulphide (pyrite-pyrrhotite) facies iron formation. These facies transitions are considered to define original shelf-to-basin bathymetry. For example in the Timmins area, magnetitic iron formation in the south is transitional northeastward through a narrow zone of sideritic iron formation to thin cherty pyritiferous iron formation. Magnetitic iron formation at Kirkland Lake is transitional eastward to siliceous carbonate iron formation as at Larder Lake and farther to thin pyritiferous zones at Noranda (Ridler, 1970). Similarly, magnetitic iron formation near Val d'Or is transitional northeastward to thin chert-pyrite-carbonaceous zones representing sulphide facies iron formation. Similar transitions are present elsewhere (Fig. 2).

Mafic Intrusions

Numerous mafic sheets, sills and dikes occur in the mafic lava accumulations. They are readily confused with diabasic phases of thick flows and commonly represent intrusive phases of the volcanism. They are also intrusive into felsic volcanic phases as at Noranda.

Differentiated sheets and sills up to several thousand feet thick typically comprise basal zones of peridotite, usually serpentinized, transitional upwards through pyroxenitic or mafic gabbro or norite, to gabbro and quartz gabbro and, locally, to granophyre (micropegmatite). In Chibougamau-Opemiska area, differentiated mafic sills include peridotite-pyroxenite, gabbro-pyroxenite and diorite-quartz diorite transitions. The Chibougamau anorthosite, 30 miles in diameter, comprises coarse-grained greenish white assemblages composed of saussuritized plagioclase and chloritized ferromagnesian minerals.

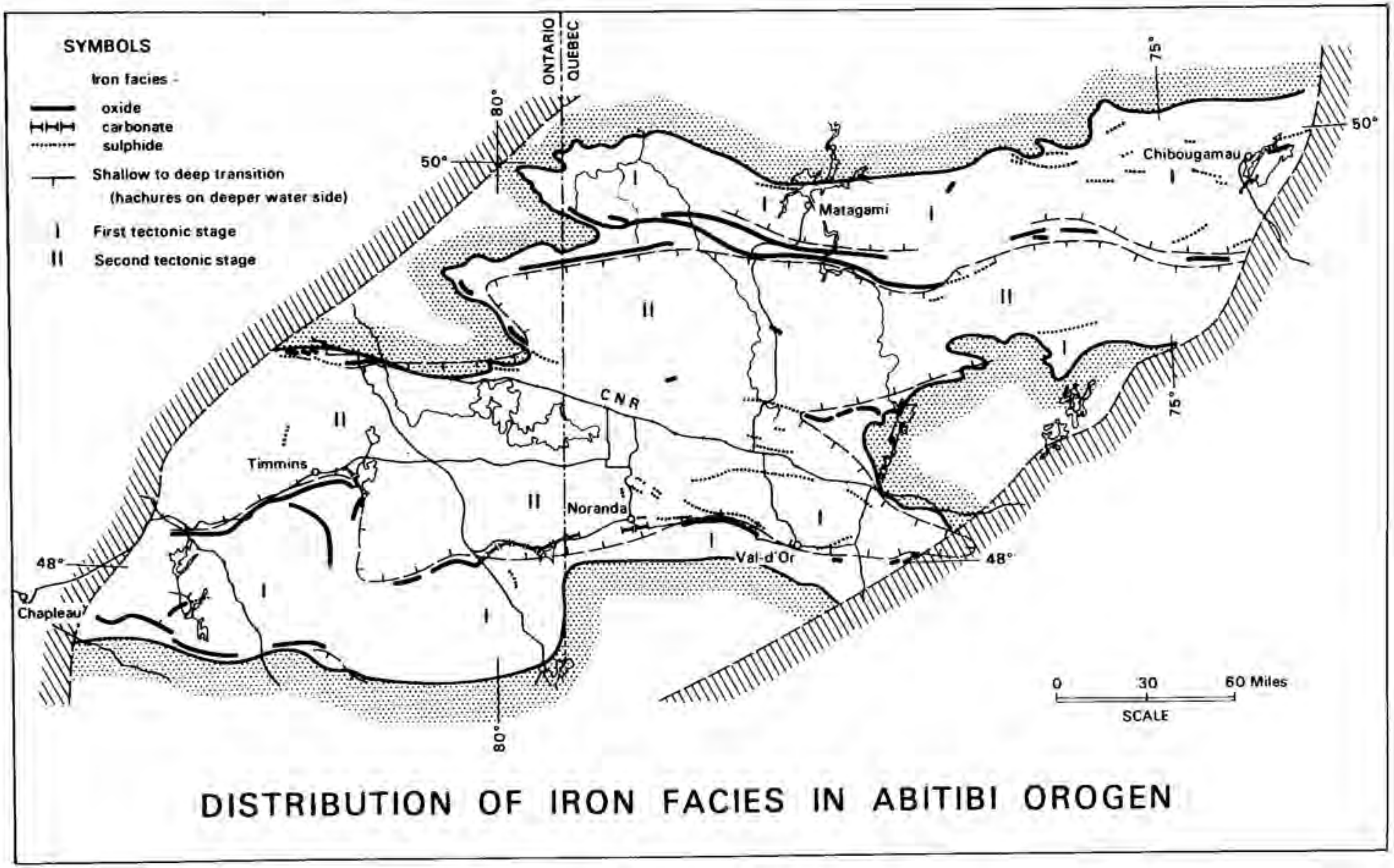


Figure 2. Distribution of facies in Abitibi orogen. Oxide-carbonate-sulphide facies of iron formation transitions delineate original shelf to basin slopes within the orogen. Tectonic stages refer to postulated episodic progressions in construction of the orogen. Presumed younger volcanic-sedimentary rocks are present towards the centre of the orogen.

The Bell River complex at Matagami, at least 15,000 feet thick, comprises strongly and rythmically banded norite, anorthosite, pyroxenite and associated gneissic cataclastic rocks. Another differentiated complex occurs at Kamiskotia Lake north of Timmins.

Granitic Rocks

A great variety of granitic rocks is present in the region but only a brief statement is made on the subject. Foliated to massive granitic rocks commonly underlie the margins of the region. They are associated with more or less metavolcanic and metasedimentary rocks. Where intrusive relations are apparent the granitic rocks have intruded the volcanic and sedimentary rocks and hence post-date them in terms of last emplacement. Large granitic complexes along the north and south boundaries may include primitive crystalline basement. By its nature the demonstration of the reality of such sialic basement is very difficult if indeed possible. The granitic rocks consist in large part of weakly to strongly foliated granodioritic gneiss of possible metasedimentary origin, together with varying proportions of schists, gneiss, migmatites, and undoubted igneous plutons. Based on geochemical reconnaissance of similar crystalline terrain in Ungava region of northern Quebec the average chemical composition of these crystalline rocks approaches that of common granodiorite (Eade, *et al.*, 1966).

In addition granitic plutons are widely distributed within the orogenic belt itself. At least nine major plutons each 40 to 100 miles in diameter are present, most in the northern part of the region. Additional smaller plutons enclosed within volcanic rocks are closely associated with felsic volcanic rocks e.g. Noranda, Malartic, Kirkland Lake and Chibougamau, of which they may represent coeval intrusive equivalents. The Round Lake batholith south of Kirkland Lake contains a southern part of leucrocratic quartz diorite and a northern, presumably younger part of hornblende granite. The Otto stock on the northeast is composed of syenite, nepheline syenite and locally, pegmatitic phases. To the east, in the Malartic area, the Bourlamaque batholith comprises a core of quartz diorite and a border of albite granite. The LaCorne-La Motte group of plutons in the north Malartic area, approximately 40 miles long by 16 miles wide, comprises respectively hornblende-, biotite-, and muscovite-granite phases, the last named associated with spodumene-bearing pegmatites and molybdenite mineralization. To the west the Taschereau, Palmarolle and Flavrian plutons comprise cores of pink granodiorite and borders of grey quartz diorite. Many other types of plutons are present in the belt of which the above provide some example.

Radiometric ages of various intrusions in the Kirkland Lake area, for example, indicate that plutonic activity has occurred there repeadedly from Archean to Mesozoic time. Particular examples are the Round Lake batholith (2,400 million years), the Otto stock (1,700 million years) and the Kimberlite dikes in the Upper Canada Mine which are ascribed to a Jurassic-Cretaceous age (Ridler, 1970). Clearly there are major problems concerning the times and mechanisms of igneous emplacement as well as the origin of sialic material contained in the plutons.

Lithic Proportions

A planimeter survey of the geologic map (Fig. 1) including only those rocks within the orogenic belt (Fig. 3) provides an assessment of lithic composition. Areas of cover rocks of Proterozoic and Paleozoic ages totalling 1,268 sq. mi. have been excluded. The belt has been divided into southern and northern parts with the main CN railway as the dividing line.

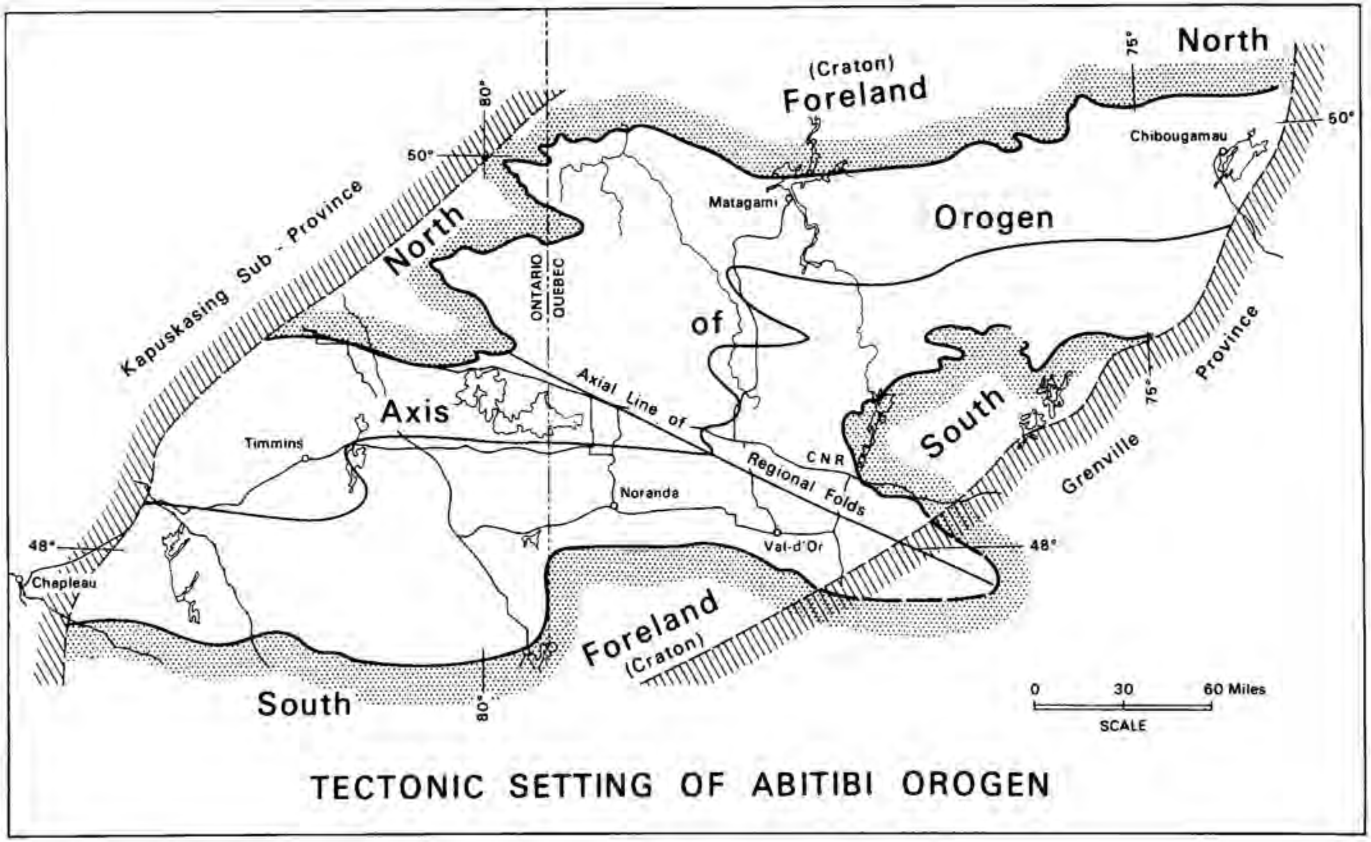


Figure 3. Tectonic setting of Abitibi orogen. Boundaries of the forelands are deformed on the basis of preponderant crystalline granitic rocks. S-shaped form of the orogen is thought to be due to structural deformation.

Abitibi Orogen

	Southern Part	Northern Part	Total
Area	15,948 sq. mi.	20,811 sq. mi.	36,759 sq. mi.
Granitic rocks	26.2 per cent	37.0 per cent	32.3 per cent
Mafic volcanics	47.5 "	44.2 "	45.6 "
Felsic volcanics	5.5 "	2.2 "	3.6 "
Sediments	18.8 "	13.8 "	16.0 "
Mafic intrusions	2.0 "	2.8 "	2.5 "
	100.0 per cent	100.0 per cent	100.0 per cent

Recalculating to Archean supracrustal rocks only (i.e. excluding granitic rocks and mafic intrusions) the following lithic proportions are indicated to be present in the belt:

	Southern Part	Northern Part	Total
Mafic volcanics	66.2 per cent	73.5 per cent	70.0 per cent
Felsic volcanics	7.6 "	3.6 "	5.5 "
Sediments	26.2 "	22.9 "	24.5 "
	100.0 per cent	100.0 per cent	100.0 per cent

Thus the indicated ratio mafic volcanics: felsic volcanics is 13:1. Stated otherwise, felsic volcanic rocks (dacite, rhyodacite, rhyolite and minor trachyte) form an indicated 5.5 per cent of the regional volcanic assemblage. The figures must be viewed with caution considering the extensive drift cover in the northern parts particularly. Past experience suggests that further field studies in the region will reveal additional bands of felsic volcanic rocks, and result in reclassification of some sediments as felsic volcanic rocks. In the meantime the figure of 5.5 per cent felsic volcanics is offered as a working minimum, the true figure possibly approaching 7 to 10 per cent felsic volcanic rocks.

The northern part of the region includes substantially higher proportions of internal granitic rocks in the form of large batholiths and less felsic volcanic rocks. Despite the note of caution expressed in the previous paragraph the stated proportions are considered to reflect a genuine difference in lithic proportions between the northern and southern parts of the region. Major problems remain concerning the origin of the sialic material and the time and mechanism of emplacement of these large northern batholiths.

Relatively more sediments are present in the southern part of the orogen (i.e. 18.8 vs 13.8 per cent). This reflects inclusion within the orogen of Pontiac sediments - the main flyschoid facies - situated south of Noranda-Malartic-Val d'Or (Fig. 1). Present indications are and future studies may prove that this facies was derived from a craton to the south and should more properly be included in the southern foreland (Fig. 3). However in order to be consistent in defining the boundaries of the forlands i.e. preponderant granitic rocks, the Pontiac facies has been included in the orogenic belt for present purposes.

TECTONIC SETTING

The tectonic setting and principal parameters of the Abitibi orogen are illustrated in Figure 3. The orogen lies between two northeast-trending boundaries, the Grenville province on the east and the Kapuskasing subprovince

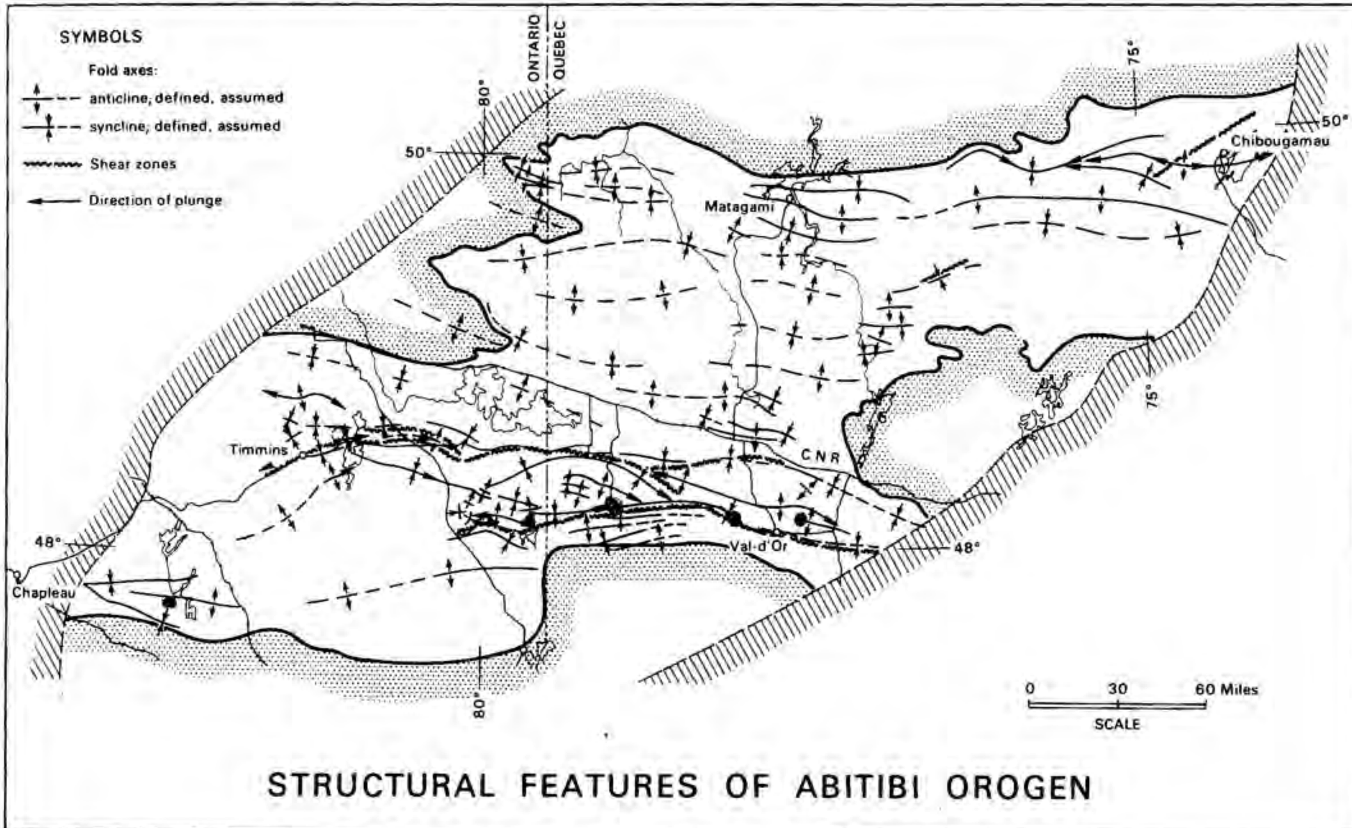


Figure 4. Structural features of Abitibi orogen. Additional folds and shear-zones may be present in drift-covered parts, especially in the north.

on the west. The orogen proper lies between two postulated cratonic forelands, the boundaries of which are defined on the basis of preponderant granitic rocks. Foreland boundaries are in fact gradational from predominantly supracrustal rocks within the orogen to predominantly crystalline complexes without. Also the foreland complexes locally include substantial masses of metavolcanic and metasedimentary rocks. This is particularly so in the north foreland as well as in the western part of the south foreland. The axis of the orogen has been placed midway between the forelands with due regard to internal lithic distribution.

Accordingly the Abitibi orogen constitutes a generally east-trending tectonic unit approximately 500 miles long and 60 to 100 miles wide. Lithologically it features orogenic supracrustal rocks, both volcanics and sediments, together with prominent granitic plutons. It is intracratonically contained between predominantly granitic forelands. The present S-shaped form reflects structural deformation.

STRUCTURE

The first-order structure of the region is an S-shaped regional fold which is delineated by the boundary surface of the orogen (Fig. 3). Thus the north boundary surface extends from Chibougamau vicinity on the east to Bradburn township north of Timmins on the west, and the south boundary surface from Marceau township situated south of Doda Lake on the east to the vicinity of Chapleau on the west. Two conspicuous second order S-folds which form part of the regional fold are present on both boundary-surfaces. Those on the north boundary surface protrude eastward in the area north of Lake Abitibi near the Ontario-Quebec interprovincial boundary. Those on the south boundary surface lie respectively northeast of Val d'Or and southwest of Noranda the latter indicated by the conspicuous southward trend of rock units south of Larder Lake. The axial line of the regional folds trends northwesterly (Fig. 3).

Folding of the compositional layering within the orogen is widespread (Fig. 4). The folds have east-trending axial lines, are characteristically isoclinal, doubly plunging, and commonly bifurcate along strike. Those to the north of the CNR line commonly plunge to the west e.g. Chibougamau and Matagami whereas those to the south commonly plunge to the east e.g. Noranda, Malartic, Kamiskotia. On a regional scale granitic plutons are preferentially distributed along anticlines and supracrustal rocks along synclines. On a local scale synclines commonly bifurcate to enclose an anticline which includes central plutons e.g. Noranda, Malartic, Matagami and Chibougamau. In the west Malartic area, volcanic (Malartic group) and sedimentary (Kewagama) rocks have been folded into a west-facing anticline; these rocks underlie younger volcanic rocks situated to the west and north.

Several prominent, steeply inclined, east-trending, discontinuous shear-zones of undetermined displacement have been identified in the southern part of the region e.g. Porcupine-Destor and Larder Lake "breaks". They follow lithofacies boundaries for the most part, including sedimentary-volcanic interfaces. Other regional shear zones of this type may be present particularly in drift-covered areas to the north. Recent field studies of the Larder Lake "break" cast doubt on its continuity and structural significance (Ridler, 1970). This suggests that a re-examination of the entire problem of the regional "breaks" is in order.

Comparison with modern volcanoes in island arc settings (e.g. Solo zone in East Java, *see* Van Bemmelen, 1949, Fig. 16, p. 26) suggests that the present elliptical outline of the Abitibi volcanic complexes (Fig. 5; *see* page 12) is due to foreshortening in a north-south direction of originally circular volcanic complexes. The present axial ratios of the volcanic complexes average 5:1. This implies that originally circular complexes have been foreshortened by at least 50 per cent. If the complexes were also foreshortened

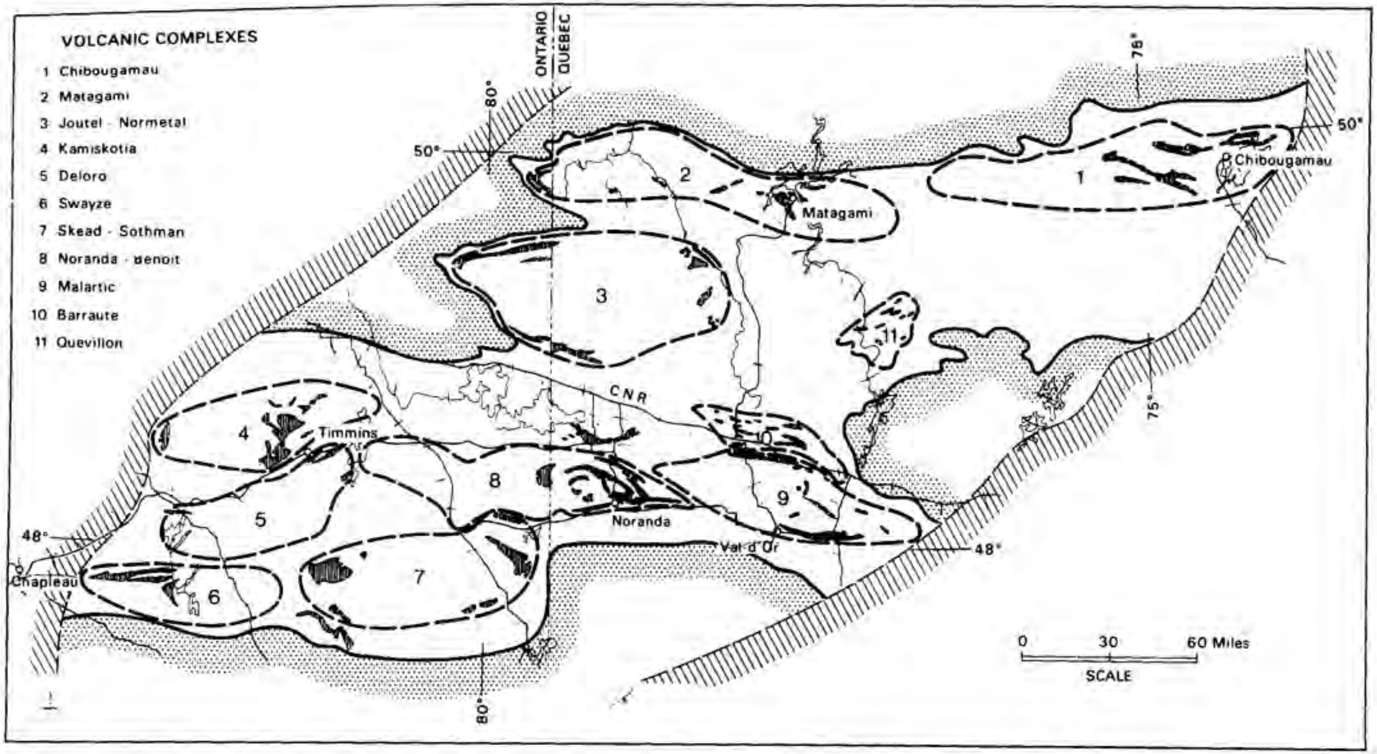


Figure 5. Distribution of volcanic complexes in Abitibi orogen. Felsic volcanic rocks shown by vertically lined pattern. Approximate boundaries of volcanic complexes including cogenetic intrusions and sediments is shown by heavy dashed lines. Present elliptical shape of presumably originally circular volcanic complexes is attributed to compression folding.

in an east-west direction as is suggested by the presence of doubly-plunging east-trending fold axes then the amount of north-south foreshortening may have substantially exceeded 50 per cent.

The present elliptical shape of the volcanic complexes together with closely spaced isoclinal folds suggests that the mechanism of deformation was compression folding. Such deformation would also produce substantial vertical extension of the rocks. Thus Archean orogenic assemblages originally 35,000 feet thick (10 km) may have been increased in thickness to approach that of the existing continental crust (40 km). More definite statements on the mechanics and state of deformation would require additional structural studies in the field.

VOLCANIC COMPLEXES

The regional stratigraphy is dominated by the presence of semi-independent ellipsoidal volcanic-sedimentary domains each of which is termed a volcanic complex. Nine major and two minor volcanic complexes have been delineated in the region (Fig. 5) each with a mafic to felsic volcanic sequence, associated intrusions and sediments.

The best known domain is that of the Blake River group herein called the Noranda-Benoit complex (number 8 in Fig. 5). The lowermost identifiable strata of this group which are of prevailing mafic composition extend from Nighthawk Lake situated 15 miles east of Timmins, 1) eastward to north Noranda area, thence southeastward to Malartic, and 2) southeastward to Kirkland Lake, thence eastward to Malartic, the two boundary lines thus outlining the complex. An identifiable mafic to felsic volcanic sequence with associated intrusions and sediments is contained within this boundary. The single felsic volcanic centre lies at Noranda in the eastern part of the complex. Although lateral extensions of Blake River rocks may fall outside this boundary the bulk of the preserved assemblage appears to lie within it.

Similarly, other volcanic complexes have been delineated with the felsic volcanic concentrations of the region serving in their identification (Fig. 5). Some boundaries have been tentatively defined only e.g. terminations of Skead-Sothman (#7), Matagami (#2), Chibougamau (#1) and Kamiskotia (#4) complexes. Barraute complex (#10) may be a structural extension of Malartic complex (#9). Quevillon complex (#11) is of uncertain definition as are the mutual boundary areas of Deloro (#5), Swayze (#6) and Skead-Sothman (#7) complexes. Other as yet unidentified complexes may be present in the region. Modifications of existing boundaries are to be expected on the basis of future work.

Despite these problems of boundary definition it is suggested that the pattern of complexes illustrated in Figure 5 accurately reflects the style and format prevailing during construction of the Abitibi orogen - a constructional style dominated by development of numerous semi-independent mafic to felsic volcanic piles with coeval intrusions and sediments. Analogy with recent and modern volcanic piles including those of linear tectonic (especially island arc) association suggests that the Abitibi volcanic complexes originally had circular outlines, each apparently in the order of 60 to 100 miles diameter. Their present elliptical outlines are attributed to structural deformation.

The defined volcanic complexes of the region are concentrated along the northern and southern boundaries of the orogen. Thus comparison of Figures 3 and 5 shows that volcanic complexes 1 to 4 lie north of the axis of the orogen and close to the northern boundary, whereas complexes 6 to 11 lie south of the axis of the orogen and close to the southern boundary. This pattern may reflect original linear distribution of strato-volcanic complexes along tectonic structures such as geanticlinal uplifts essentially parallel to the margins of the developing orogen.

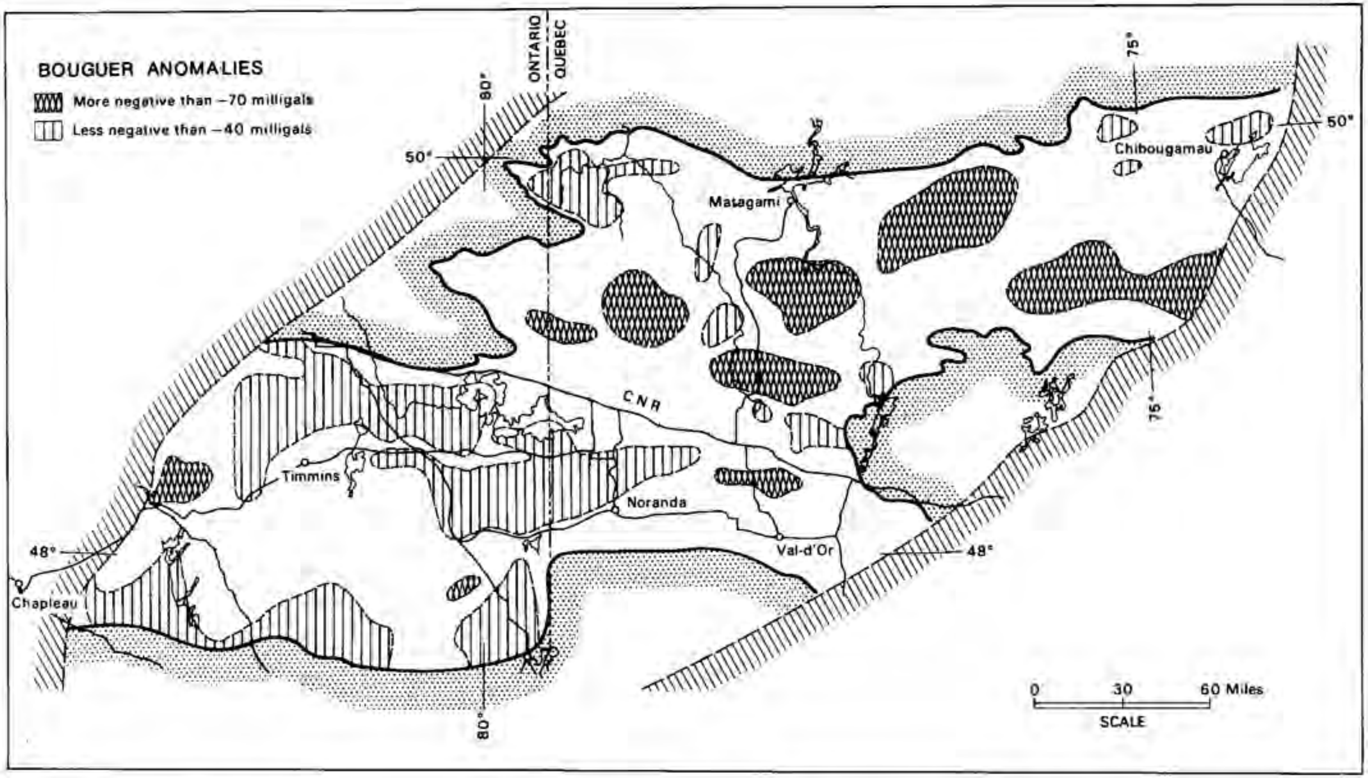


Figure 6. Gravity features in Abitibi orogen. Areas between the designated Bouguer anomalies have intermediate gravity expressions of -70 to -40 milligals. Bouguer anomalies more negative than -70 milligals correspond to granitic batholiths. Anomalies less negative than -40 milligals correspond to large areas of mafic igneous rocks.

GRAVITY

Gravity and geologic features are closely related in Abitibi orogen. This may be seen by comparing the Bouguer anomaly map (Fig. 6, which is based on Gravity Map of Canada GMC 67-1) and the geologic map (Fig. 1). The close correspondence of gravity and geologic features indicates that the regional gravity measurements summarized in the gravity map substantially reflect surface and near-surface crustal features rather than deep-crustal or mantle features.

Six principal high negative (less than - 70 milligals) Bouguer anomalies north of the CN railway coincide remarkably well with the main granite batholiths of the region. By demonstrating the presence in the crust of substantial bodies of low density material, they indicate that the batholiths have deep roots. Three high negative anomalies south of the CN railway coincide with smaller granitic batholiths (near Val d'Or, Kirkland Lake and Timmins). In contrast several granitic batholiths within the orogen do not have comparable high negative gravity expressions e.g. Matachewan, Timmins, Gogama areas. Such batholiths may terminate at shallow depths or alternatively, contain igneous material of intermediate or assorted densities.

The principal low negative (greater than - 40 milligals) Bouguer anomalies coincide with 1) major mafic volcanic-intrusive assemblages and 2) flat-lying Proterozoic diabase along the southwestern margin. For example, thick, extensive mafic assemblages are reflected by major low negative anomalies northwest of Noranda (Blake River group mainly) and north of Timmins. Four local mafic assemblages are reflected by correspondingly small, low negative gravity anomalies northwest of Senneterre, between granitic batholiths near Joutel-Poirier, west of Matagami, and near Chibougamau.

All other parts of the orogen, which yield an intermediate gravity expression (between -70 and -40 milligals) are underlain apparently by mixed lithic assemblages which collectively lack an anomalous gravity expression.

Gravity relationships clearly demonstrate that the part of the orogen north of the main CN railway contains less total mass (i.e. more low-density rock) than the southern part. This is attributed mainly to the presence of numerous granite batholiths with deep roots and proportionately less mafic igneous rock in the north. In addition some undisclosed deeper crustal influence may be reflected.

It is particularly significant that high negative Bouguer anomalies (less negative than -70 milligals) coincide with internal granite batholiths only and not with the regional granitic complexes of the northern and southern forelands as defined in Figure 3. This provides a ready method of differentiating true granitic batholiths from the mixed lithic predominantly granitic assemblages of the forelands. In this connection two high negative Bouguer anomalies, each presumably an expression of a granite batholith, lie outside the orogen; and within the bordering crystalline complexes. One lies in north foreland rocks, 50 miles north of Matagami and the other in Grenville province rocks 50 miles southeast of Chibougamau (*see* Gravity Map of Canada). With the exception of these two high negative areas the forelands have intermediate gravity expressions suggesting the presence of mixed crystalline complexes of assorted gneissic-migmatitic-plutonic composition.

METALLOGENESIS

Type and Distribution of Mineralization

The region contains several thousand known mineral occurrences including 150 producers (past and present) of which 57 lie in Ontario and 93 in

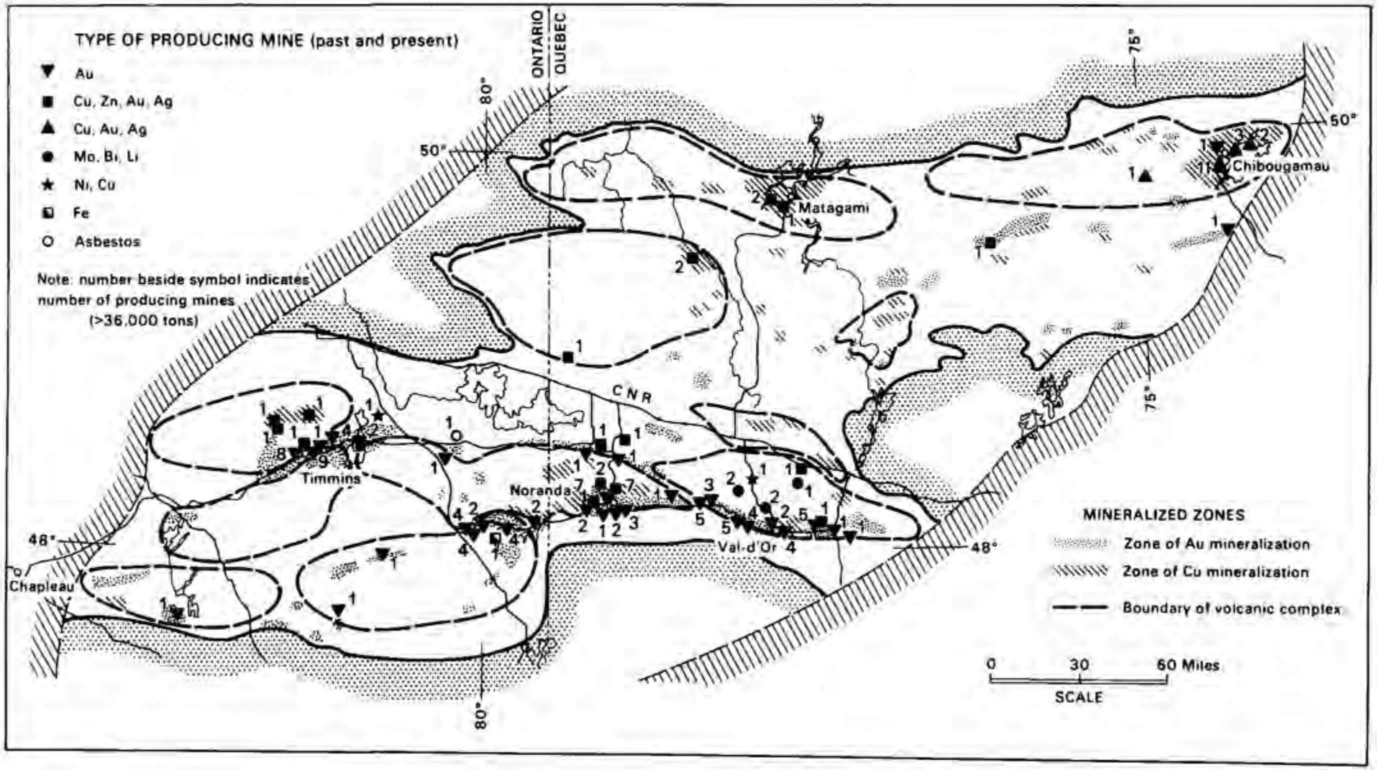


Figure 7. Metallogenic relations in Abitibi orogen. Distributions of main Au and Cu mineralized zones and of producing mines (greater than 36,000 tons; past or present) are shown relative to volcanic complexes.

Quebec. Most occurrences are either gold-bearing quartz veins or disseminated to massive sulphide (Cu-Zn-Au-Ag) concentrations. Others include magnetitic iron formation, asbestos, Ni-Cu, Zn-Ag-Pb, Mo-Bi and Li.

The main mineralized zones based on known occurrences are shown in Figure 7. A principal east-trending Au zone extends intermittently from Sothman-Matachewan area on the west through Kirkland Lake, Larder Lake, Noranda and Malartic to Val d'Or area on the east. A second well-established Au zone extends eastward from Timmins vicinity through Duparquet thence curving southeastward to join the first zone at Malartic. Smaller discontinuous Au zones are widely distributed in the region, for example, in the Swayze, south Timmins, Benoit and north Malartic areas; north of the CN railway in the Doda-Opawica area, and southwest of Chibougamau and Quevillon areas. Principal zones of Cu mineralization, on the basis of producers and known occurrences, lie in north Timmins, Noranda, Malartic-Val d'Or, Joutel, Matagami and Chibougamau areas. Smaller zones lie west and south of Matagami, west of Chibougamau and in the Barraute-Quevillon area. Several other small isolated mineralized patches without obvious pattern are present in the region as illustrated in Figure 7.

Lithic Relationships

Most Au deposits lie in intermediate to felsic volcanic rocks, their intrusive equivalents or associated clastic or exhalative sediments. Included are those at Swayze, Matachewan, Kirkland Lake, Timmins, Cadillac, Malartic and Val d'Or. Sulphide deposits of the Cu-Au and Cu-Zn-Ag types are associated with intermediate to felsic volcanic complexes at Noranda, Joutel, Timmins and Matagami. Mo-Bi and Li deposits are associated with pegmatitic rocks in north Malartic area. A single iron producer lies in banded iron formation enclosed in felsic tuff and mafic lavas south of Kirkland Lake. Asbestos deposits are associated with differentiated ultramafic complexes north of Matheson and west of Timmins. Ni-Cu deposits lie in mafic to ultramafic intrusions near Cochrane, south of Timmins and north of Malartic. Cu-Ag-Au deposits are associated with gabbroic, noritic and anorthositic intrusions of volcanic association at Chibougamau (Duquette, in press) and Cu-Zn-Ag deposits with anorthositic intrusions at Matagami (Sharpe, 1965).

Relation to Volcanic Complexes

As illustrated in Figure 7 most mineralized zones including all but three of the 150 producers lie within or marginal to the designated volcanic complexes. Principal Au zones lie at the margins of complexes and have traditionally been interpreted as being primarily associated with "shear-zones" e.g. Porcupine-Destor and Larder Lake "breaks". For example, one Au zone extends from Matachewan through Kirkland Lake at the northern margin of Skead-Sothman complex (#7 in Fig. 5) eastward along the southern margins of the Noranda-Benoit (#8) and Malartic (#9) complexes; a second Au zone extends from the mutual boundary of Kamiskotia (#4) and Deloro (#5) complexes eastward along the northern and northeastern margins of the Noranda-Benoit (#8) complex. Other, leaner, Au-bearing zones are present in the western parts of Swayze (#6) Skead-Sothman (#7) and Deloro (#5) complexes. In addition, several small Au zones without obvious relation to volcanic complexes occur in the area southwest of Chibougamau and north of the main CN line.

Most Cu-bearing sulphide zones lie in the volcanic complexes directly associated with felsic volcanic rocks. They are most common in the eastern parts of the following six volcanic complexes: Chibougamau (#1), Matagami (#2), Joutel-Normetal (#3), Kamiskotia (#4), Noranda-Benoit (#8) and Malartic (#9). Cu-Au and Cu-Zn-Ag producers are present in Matagami (#2), Joutel-

Normetal (#3), Kamiskotia (#4), Noranda-Benoit (#8) and Malartic (#9) complexes whereas Cu-Au-Ag producers are present in Chibougamau (#1) complex. Other small local zones of Cu mineralization occur in Barraute (#10) and Quevillon (#11) complexes, in the central and western parts of Matagami (#2) and Chibougamau (#1) complexes and elsewhere north of the main CN railway as illustrated in Figure 7.

Mo-Bi and Li occurrences are associated with pegmatitic phases of felsic intrusions in the centre of Malartic (#9) complex. Magnetitic iron formation is associated with volcanic tuff and mafic lavas in northeastern Skead-Sothman (#7) complex. Ni-Cu producers lie in ultramafic intrusions in or near Malartic (#9) and Noranda-Benoit (#8) complexes.

Lithic associations common to sulphide deposits are felsic volcanic concentrations reflecting central vent eruptions, sulphide facies iron-formation reflecting reducing environments, and absence or scarcity of coarse clastics reflecting offshore, deeper water sites of accumulation. Such deeper water environments may be attributed in part to tectonic collapse including cauldron subsidence following felsic volcanic discharge. Conversely some lithic associations are inimical to sulphide occurrences: e.g. oxide facies iron formation and clastic sediments, especially conglomerate. This association reflects topographically higher, shallow water, partially oxidizing shelf-environments. Accordingly, most mineralized sulphide zones in the region are directly associated with felsic volcanic concentrations having sulphide facies iron formation nearby e.g. Noranda, Malartic, Matagami and Chibougamau. On the other hand, many Au occurrences are directly associated with carbonate facies iron formation either at felsic volcanic-intrusive centres or at the margins of volcanic complexes.

Stratigraphic Relations

Cu-Zn-Au-Ag concentrations have the most direct stratigraphic relationship. Many lie at specific volcanic contacts, commonly felsic-mafic transition in upper parts of stratigraphic successions e.g. Kamiskotia, Kidd Creek Noranda, Matagami, Joutel and Malartic areas. Cu-Au-Ag deposits lie along fractures at the northern margin of a volcanic-enclosed, anorthositic complex at Chibougamau. Some Ni-Cu deposits lie in ultramafic intrusions either within stratigraphically lower mafic volcanic rocks or upper felsic volcanic rocks e.g. Marbridge and Alexo mines.

Au deposits have variable degrees of stratigraphic associations. One family of deposits has endogenous relations to felsic alkaline intrusions in the form of subvolcanic sills, discordant plugs or fairly large stocks. The gold occurs disseminated in irregular zones often associated with pyrite or concentrated into various types of vein structures. Although the plutons are believed to be cogenetic volcanic equivalents, and therefore Archean in age, it is possible that post-Archean igneous events have been accompanied by gold mineralization. Examples are the gold mines of Kirkland Lake, Granada, Yonge-Davidson, Eldrich, Powell-Rouyn and Sullivan mines. Flow and pyroclastic equivalents of the above class of plutons may be enriched in gold, either disseminated or in vein form. Examples are Dome, Hollinger and McIntyre mines of the Timmins area and certain parts of Upper Canada mine east of Kirkland Lake. Gold deposits intimately associated with carbonate facies iron formation include Omega and Kerr Addison (Ridler, 1970); those with interflow deposits include Wasamac mine west of Noranda and Dome deposit of Timmins area. Gold deposits of possible placer derivation in whole or in part include Pamour mine, lying in quartzite east of Timmins and McWatters mines in conglomerate south of Noranda, each situated in the upper part of the local stratigraphic succession. In addition, many of the gold deposits lie in dilatant shear and fracture zones into which the precious metal together with accompanying quartz appears to have been concentrated from the foregoing original stratigraphic sites during brittle deformation. This has partly obscured primary stratigraphic distribution and has produced a distribution pattern which has been traditionally related to a regional fracture system.

Finally, magnetitic iron formations are enclosed in volcanic rock in upper stratigraphic parts of local successions e.g. Adams mine at Kirkland Lake, whereas asbestos deposits lie in differentiated mafic sills and complexes towards the base of local stratigraphic successions, e.g. Munro mine north of Matheson.

TECTONIC INTERPRETATION

It has been established that Archean rocks of the Abitibi orogen have been structurally deformed to a considerable degree. Relations indicate that component lithofacies have been tightly folded about east-trending axes, resulting in substantial north-south shortening. Regional faults may be present. Deformations of this type have combined to obscure original tectonic relationships. Despite these reservations and uncertainties which are common to Archean greenstone belts in general, a tectonic reconstruction based on available data is offered for consideration as a working hypothesis.

The nature, distribution and composition of volcanic and sedimentary rocks points clearly to the presence in Archean time of a mobile belt or tectonically unstable, linear basin complex. Archean mafic to felsic volcanic assemblages have their modern analogues in orogenically active, thin-crustal environments such as those of island arcs which are characteristic of continent-ocean interfaces. Archean sediments, notably greywacke, conglomerate and breccia, are of the high-energy, "poured-in" turbidite association. Thus tectonic mobility is a clearly indicated characteristic of the orogen. Further, the present location of the east-trending orogenic belt between adjoining predominantly granitic terrains suggests that the orogen developed between two sialic forelands. The presence of flyschoid clastics (Pontiac group) in Noranda-Val d'Or area derived from a granitic provenance to the south (Holubec in preparation), rein forces this interpretation of the presence of a sialic foreland south of Noranda during Abitibi time. Therefore, we deduce that Abitibi orogen developed intracratonically upon thin mobile crust, between adjoining forelands of predominantly sialic composition. The apparent absence of flyschoid facies along the northern margin of the orogen and along the western part of the southern margin, together with the presence of some volcanic and sedimentary rocks within the adjoining predominantly granitic terrains suggest that comparatively subdued sialic forelands of mixed lithic composition existed in those parts during Abitibi time.

Two elements of regional stratigraphy are used to reconstruct tectonic development of the orogenic assemblage, namely facies transitions of iron and relative ages, as known, of specific volcanic complexes.

The main shelf-to-basin transitions in the orogen, based on iron formation facies-relationships, extend east-west across the northern and southern parts of the region. In addition other local transitions are present. The transitions, in effect, separate common shelf-type oxide facies iron formation, present along the northern and southern parts of the orogen, from deeper water or basinal sulphide facies iron formation towards the centre of the orogen. These relationships, in brief, point to the presence of tectonic basin(s). Successive basins may have been present during progressive development of the orogen. The local presence of sulphide facies iron formation near Chibougamau and Matagami may be due to cauldron subsidence or to some other form of local tectonic collapse.

Detailed stratigraphic studies in Kirkland Lake area (Ridler, 1970) indicate that an older volcanic assemblage (Skead group) mainly to the south of Kirkland Lake proper underlies younger Blake River volcanic rocks to the north (*see* Figs. 1 and 5); that older Skead volcanics constituted a stable shelf

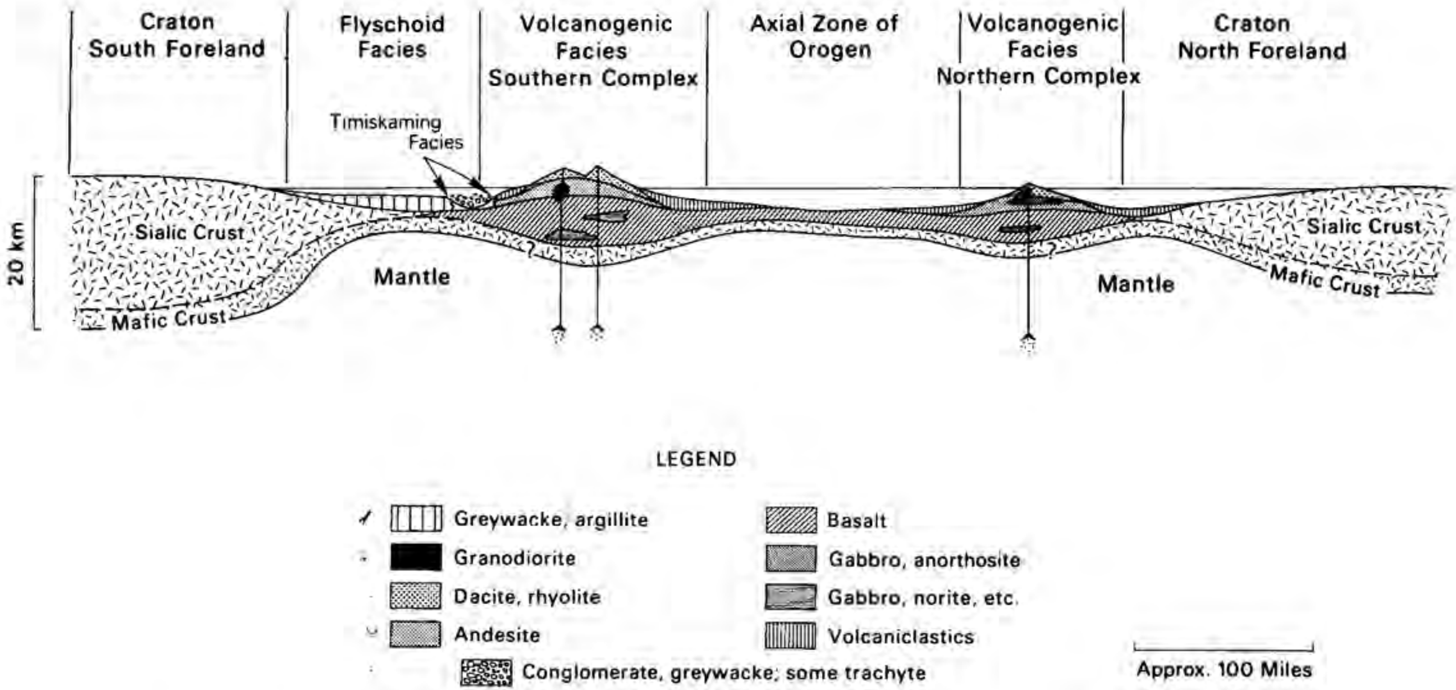


Figure 8. Hypothetical tectonic reconstruction of Abitibi orogen. Vertical cross-section. Length of the orogen is approximately 500 miles. Distance between southern and northern volcanogenic facies may have been in the order of 200 miles. Vertical scale is approximate only. The line of the section is intended to cross Noranda in the south and Matagami in the north. The smaller size and relief of the northern sialic foreland is indicated by the absence of flyschoid facies along its southern margin. The nature and thickness of mafic crust underlying the orogen is conjectural. In the absence of direct evidence of sialic contribution to the volcanic and sedimentary assemblages of the orogen (except for the flyschoid facies, i.e. Pontiac, at the south margin), no subjacent sialic crust is shown as being present, at least during the early development stages of the orogen.

relative to the Blake River basin of mafic volcanic accumulation to the north; and futhermore that the shelf-to-basin transition is marked by oxide- and carbonate-facies iron formation. In the Timmins area the east-trending iron formation lies between underlying Deloro volcanics to the south and what is generally interpreted as younger Tisdale volcanics to the north. Malartic volcanic and sedimentary (Kewagama) rocks stratigraphically underlie volcanic rocks to the west (Blake River) and to the north.

Therefore it is suggested that the main shelf-to-basin transitions of the region, marked by iron formation, delineate volcanic assemblages of slightly different ages. The younger assemblage are interpreted as occupying the deeper water side, towards the basin and hence mainly towards the axis of the orogen. Accordingly, the following volcanic complexes and rock groups appear to possess some degree of contemporaneity: in the south, Deloro (#5 in Fig. 5), Swayze (#6), Skead-Sothman (#7) and Malartic (#9) complexes and Pontiac sediments; in the north, Matagami (#2) and Chibougamau (#1) complexes. Somewhat younger and bearing a similar degree of contemporaneity, on the same basis, are Kamiskotia (#4), Noranda-Benoit (#8) and Joutel-Normetal (#3) complexes together with associated belts of sediments. Thus the postulated sequence of events is one of episodic progression in tectonic development from orogenic margins to the axis of the orogen. This working hypothesis requires detailed testing. In this regard radiometric dating, isotopic and paleomagnetic studies, and detailed stratigraphic studies may be fruitful.

Most of the volcanic complexes with their felsic concentrations lie close to the boundaries of the orogen. All felsic concentrations lie within 40 miles of a foreland boundary and all but two (Kamiskotia and Joutel) lie within 25 miles. Thus the main bulk of defined felsic volcanic rock is situated near the margins of the orogen. Because most mineral deposits (excluding asbestos and Ni) are closely associated with felsic rocks, it follows that metallogenic patterns bear a similarly close spatial relationship to foreland boundaries. The two exceptions referred to above are in 1) the Timmins area where major Au and Cu-Zn deposits lie 40 and 30 miles respectively from the northern foreland boundary and 2) the Joutel Cu-Zn deposit approximately 40 miles east of a folded spur of the same boundary. However, rock assemblages in both of these areas are highly deformed and extensively drift-covered so that their true stratigraphic positions relative to original foreland boundaries are difficult to interpret.

Within the limitations imposed by the existing state of deformation, drift-cover and knowledge of stratigraphic relations the main tectonic features of the orogen have been reconstructed and presented diagrammatically in Figure 8. Thus the orogen is considered to have developed intracratonically between two sialic forelands. Accumulation of mafic to felsic volcanic complexes proceeded, presumably under deep-fracture control, in proximity to sialic forelands i.e. at thick-thin crustal interfaces. Igneous differentiation produced effusive products ranging from basalt through andesite to rhyolite, in that general order of abundance. The role of sialic contamination in development of felsic differentiates has not been adequately tested but may have been slight. Within the orogen and remote from sialic forelands, tholeiitic basalt was apparently the predominant volcanic effusive, to the exclusion of felsic differentiates. This apparent restriction of felsic differentiates to the vicinity of sialic forelands may reflect, in addition to deep-fracture control at thick-thin crustal interfaces, a more advanced stage of igneous differentiation in high-level magma chambers at topographically higher, near-shore sites. Conversely, the central part of the orogen with predominant tholeiitic basalt and fine-grained sediments may reflect deeper water basinal sites at lower elevation resting upon thinner mafic crust.

Construction of volcanic complexes apparently proceeded in epidsodic progression, the axis of successive volcanic accumulation shifting away from forelands and towards the axis of the orogen. Accordingly younger assemblages appear to lie towards the axial zone. This distribution pattern may reflect progressive tectonic spreading of sialic forelands possibly a manifestation of crustal floor spreading in Archean time. Accordingly the apparent preference of Abitibi volcanic complexes for the margins of the forelands may correspond to the world-wide distribution of currently active volcanoes which are mainly in areas where the moving sea-floor turns down under the continents (Heirtzler, p. 9).

As previously stated, metallogenic patterns are intricately related to volcanic patterns in the orogen. Many deposits, e.g. asbestos, Cu-Ni, Mo-Bi, Li, Cu-Zn-Au-Ag and iron are directly enclosed in original lithic hosts either intrusive or extrusive. Other deposits, e.g. Au-Ag, have been involved in structural deformation with more or less migration of components to favourable sites. On a regional scale, there is little doubt that mineralization represents an integral part of the tectonic evolution of the orogen.

√CONCLUSIONS

The east-trending Abitibi orogen, 500 miles long by 60 to 100 miles broad, is bounded east and west by northeast-trending crystalline rocks of the Grenville province and Kapuskasing subprovince respectively. The orogen is intracratonically contained between northern and southern predominantly granitic forelands. Its present S-shaped form reflects severe structural deformation, dominated internally by tight isoclinal east-trending folds. The original orogen constituted a much longer and in all likelihood, broader mobile belt.

Based on planimeter survey of the geologic map, the present orogen is 37,000 square miles in area. It is underlain by 32.3 per cent granitic rocks, 45.7 per cent mafic volcanics, 3.6 per cent felsic volcanics, 16.0 per cent sediments and 2.5 per cent mafic intrusions. The proportion of granitic rocks, in the form of large batholiths, is higher in the northern parts of the orogen, a fact supported by regional gravity anomalies.

Archean supracrustal rocks, of predominantly orogenic type, are products of tectonically unstable, thin-crustal, mobile environments. Mafic to felsic volcanic sequences of predominant calc-alkaline affinity correspond in significant degree to those of modern island arcs situated at continent-ocean interfaces. Accompanying clastic sediments including volcanogenic and flyschoid facies have the essential characteristics of turbidite associations including greywacke-conglomerate assemblages, graded bedding, soft rock slump structures and other evidence of rapid, high energy, "poured-in" accumulation histories.

Iron formations, widely distributed in the regional assemblage, comprise oxide, carbonate and sulphide facies. The facies are arranged across the orogen in shelf to basin transitions corresponding to original basin bathymetry.

Regional stratigraphy is dominated by the presence of numerous semi-independent, elliptical volcanic masses, each with a mafic to felsic extrusive sequence, cogenetic intrusions and sediments. Of eleven such volcanic complexes within the orogen four lie close to the northern foreland and six close to the southern foreland.

Mineral occurrences and ore deposits, mainly Au-Ag, Cu-Zn-Ag, Cu-Au-Ag, Ni-Cu, Mo-Bi-Li, asbestos and iron, preferentially lie within volcanic complexes, where they are commonly associated with felsic rocks. Cu-Zn-Au-Ag deposits favour deeper water lithic associations marked by felsic volcanic concentrations, sulphide facies iron formation and scarcity or absence of coarse clastics. Many Au-Ag occurrences are associated with shear zones in shelf facies with felsic concentrations, clastic sediments and carbonate facies iron formation. Asbestos and Ni deposits, are directly associated with mafic intrusions.

Abitibi orogen is considered to have been developed intracratonically between spreading sialic forelands. Igneous differentiation leading to development of mafic to felsic volcanic sequences occurred within the orogen under presumed deep-fracture control in proximity to sialic forelands which represented the thick-thin crustal interfaces of Abitibi time. The role of sialic contamination in development of felsic differentiates is uncertain but may have been minimal. Accumulation of supracrustal assemblages apparently proceeded in episodic progression, with the younger rocks being distributed towards the axis of the orogen.

Tectonic development may be attributed to either, 1) intracratonic ocean-floor spreading in response to mantle convection in the manner of modern plate tectonics, or, 2) withdrawal of deep-crustal support with consequent downsinking and progressive filling of the resulting mobile belt or basin complex in the manner of conventional geosynclinal development. Analogy with modern crustal architecture lends some support to the concept of Archean ocean-floor spreading. However a great deal of additional work would be required to test the hypothesis.

ACKNOWLEDGMENTS

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Comments from: A.J. Naldrett,
University of Toronto, Toronto

Re: <u>Ultramafic and related mafic rocks of the Abitibi orogen</u>

Ultramafic rocks and their related mafic rocks in the Abitibi orogen fall into three broad classes:

1) large (20-50 miles long, over 20,000 ft. thick) igneous complexes
2) smaller (1-5 miles long, up to 4,000 ft. thick) layered intrusions
3) lenses of peridotite and pyroxenite, 10 x 200 ft. up to ½ mile x 4 miles in area.

The large complexes of class 1 include, the Dore Lake complex of the Chibougamau area (Allard, 1970), the Bell River complex of the Matagami area and possibly the large body of gabbroic rocks located 15 miles west of Timmins. The first two of these consist largely of alternating stratiform layers of anorthosite, pyroxenite and gabbro, successive layers becoming gradually more quartz-rich upwards in the sequence. Anorthositic layers are particularly common towards the base of each complex and both complexes are notably rich in titaniferous magnetite and ilmenite. Sharp (1965) comments that multiple intrusion has played a role in the development of the Bell River Complex. However, the main feature of these complexes is their stratiform nature and the conformity of this stratification with that in the enclosing volcanic rocks. This indicates that cooling and fractionation of the complexes were essentially complete when folding first affected the orogen.

Phinney (1970) has pointed out that the differentiation trend of the early Keewenawan lavas along the north shore of Lake Superior can be explained if they are regarded as successive extrusions of magma, decanted from the main body of the Duluth complex which was undergoing fractionation at this time. There are some close similarities between the anorthositic gabbros of the Duluth complex and the lower portion of the Dore Lake complex as described by Allard. We should seriously entertain the possibility that the large igneous complexes of the Abitibi orogen are actually examples of magma chambers, similar to the Duluth complex, in which differentiation was occurring to give rise to the felsic portions of the volcanic pile. The close correlation between the igneous complexes and centres of felsic volcanism at Timmins, Matagami and Chibougamau as shown on Goodwin and Ridler's (1970) geologic compilation is certainly remarkable. If these complexes are in fact the source of the felsic volcanism, it is possible that they are also the source of the economically important fumerolic activity associated with the felsic volcanics.

The small layered intrusions of class 2) have been described by MacRae (1969) and Naldrett and Mason (1968). Many of them are restricted to a narrow belt, approximately 20 miles wide, extending 90 miles, from just south of Cochrane, east-southeast across the orogen to the Ontario-Quebec border. Although the proportion of mafic to ultramafic rocks varies from intrusion to intrusion, most of them are classic examples of a differentiated sill, with a basal zone of cumulus olivine passing upwards through an augite cumulate and then a plagioclase-augite cumulate to a capping of granophyric gabbro. The igneous layering in these bodies coincides very closely with stratification in the enclosing volcanic rocks, indicating that these intrusions also crystallized before folding had affected this part of the orogen. The only economic mineralization known to be associated with this class of intrusion is the Munro asbestos deposit near Matheson, Ontario, mined by the Canadian Johns-Manville Co. Ltd., from 1950 to 1964.

The lenses of class 3 are best developed along and to the south of the axis of the orogen, particularly south and east of Timmins, Ontario and north of Malartic, Quebec. The lenses are important because all of the important-looking nickel sulfide deposits of the orogen are associated with them.

Typically, the lenses are conformable with the surrounding volcanic rocks and consist of a core of closely packed equant olivine crystals with interstitial hypersthene and augite and a marginal zone, 10 to 100 feet wide, over which the proportion of olivine to pyroxene gradually decreases. It is thought that many of the lenses were intruded as a suspension of olivine crystals within a pyroxenitic liquid. As is usual in cases of the flow of suspended solids (Gibb, 1969), some of the interstitial liquid appears to have migrated to the margins and served to lubricate the intrusions.

Naldrett and Mason (1968) have described some lenses of class 3 type from Dundonald township, east of Timmins, in which the marginal portions contain large skeletal crystals of olivine, pyroxene and chrome spinel (Fig. A). Pyke (1970) has reported similar rocks from south of Timmins and MacRae (personal communication) has observed them in the Matheson area. Except that they are much larger, these skeletal crystals are identical with those produced on quenching experimental charges of appropriate compositions and it seems

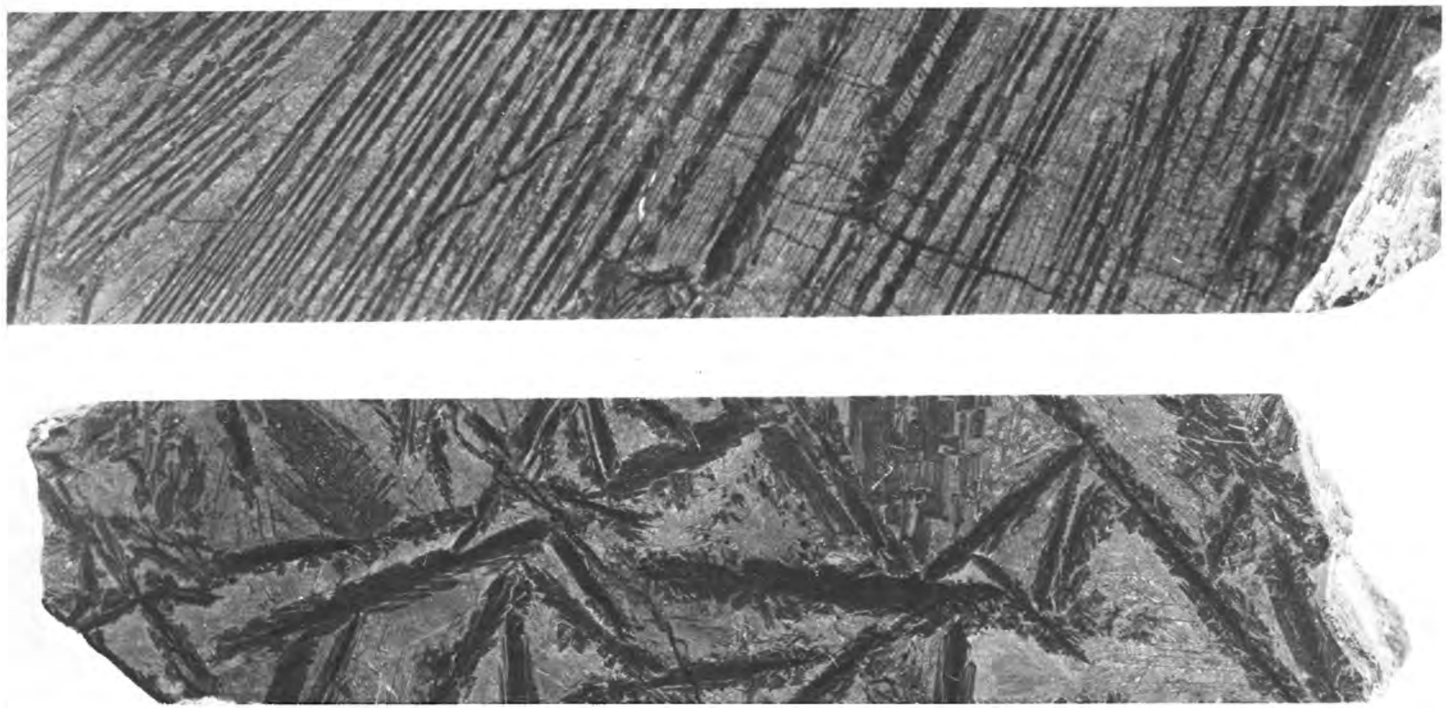

Figure A. Samples of drill core (6" long) containing skeletal and plate-like crystals of olivine (dark) in a matrix skeletal pyroxene crystals. Sample from Dundonald township, Ontario.

certain that they are also the result of very rapid chilling. The rocks in which the crystals occur are pyroxenites containing up to 50 modal per cent olivine, which indicates that the liquid from which they have formed was ultramafic rather than mafic in composition.

Figure B. Thin section of rock containing skeletal olivine (light grey) from the Komati formation of the Onverwacht group, Barberton Mountainland, Republic of South Africa. X10.

The rapid cooling indicated by the skeletal crystals raises the question of whether these lenses are extrusive or not. So far, poor outcrop has prevented our solving this problem with observations on the Abitibi rocks (most of our information has come from drill core) but Viljoen and Viljoen (1969) have described rocks with very similar skeletal olivine (Fig. 2) from the Barberton Mountainland of South Africa. The rocks of the Mountainland are part of a well preserved Archean geosyncline that has been dated as 3.4 billion years old. The exposure at Barberton is much better than in Canada. The skeletal rocks are closely associated with peridotite lenses, up to two miles long, that are interstratified with pillowed basalts in the lower portion of the volcanic pile. Pillow structures are also developed in the ultramafic rocks and the Viljoen brothers regard all of this sequence as extrusive. It seems very likely that some of the Canadian examples are also extrusive. Whether this is true or not, the Canadian rocks clearly crystallized in a very cool environment, close to the top of the volcanic pile, showing that they are part of the eugeosynclinal volcanism.

In conclusion, I stress that all of the ultramafic and associated mafic rocks of the Abitibi orogen crystallized at an early stage in the history of the orogen, before much folding had occurred. This indicates that the magmas involved may have a common origin with those forming the volcanic rocks. The large anorthositic complexes may represent the magma chambers in which differentiation occurred to produce some of the more felsic volcanic rocks. Petrographic and geochemical work on the complexes and surrounding volcanic rocks should be able to prove or disprove this suggestion.

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Comments from: E.C. Appleyard,
University of Waterloo, Waterloo, Ontario

The model that has been presented for the Abitibi Orogenic Belt has been questioned both on the grounds of the mechanism for orogenesis over a spreading oceanic ridge and the source for the granitic plutons which are emplaced within it. In addition it has been suggested that the belt has suffered considerable north-south shortening during the development of the major internal fold system. In terms of the model it is important to know the original size and orientation of the belt. The evidence that has been presented for crustal shortening may be true but could be largely circumstantial. Can you therefore strengthen your arguments in favour of the original setting of the belt, possibly by referring to quantitative structural studies such as might indicate the actual strain suffered regionally throughout the belt?

Response by: A.M. Goodwin

We know very little of the mechanisms which operated during development of Archean greenstone belts. We do know that they represent substantial concentrations of predominantly mafic igneous rocks which accumulated in active orogenic sites. Modern counterparts which satisfy this twin lithic-tectonic association are found mainly at continental-oceanic interfaces (e.g. island arcs) and in oceanic basins. So far we do not have the necessary detailed stratigraphic, geochemical and geochronological data on Archean greenstone belts to satisfy a more sophisticated genetic interpretation. Hopefully such will be provided in the years ahead as a result of diligent sophisticated studies. In the meantime a main requirement is to maintain a thoroughly flexible viewpoint within the bounds of scientific discipline. In this context the general concept of ocean floor spreading and plate tectonics is a potent candidate for serious consideration, not so much in terms of present crustal content but in that of primitive Archean times. The "spreading oceanic ridge" concept is suggested as a possible mechanism in this light.

The known supracrustal associations of the Abitibi orogen do not include any significant evidence of the existence of a major immediately subjacent sialic presence during development of the orogen. Such evidence might be expected to take the form of numerous sialic sedimentary intercalations, products of active erosion of geanticlinal sialic welts developed intermittently within the developing belt or of large masses of sialic extrusives, the products of wholesale crustal melting. On the contrary the Abitibi supracrustal association points to a predominantly mafic igneous source. For this reason, in constructing the model, we omitted sialic crust immediately beneath the orogen. On the other hand there is strong evidence, in the form of widespread quartzo feldspathic sediments (i.e. Pontiac group), of a major sialic presence marginal to the orogen, but this does not necessarily mean that sialic crust extended under the crust. The problem of the presence of large sialic plutons within the belt remains unsolved. If such plutons were, in fact, derived from immediately subjacent crust then our model must be changed to include thick, subjacent, Archean sialic crust. But we do not yet know the relative ages of the internal plutons, which are Archean and which are younger. Do they represent new, mantle-derived, syn- or post-orogenic sialic material intruded into the supracrustal assemblage? Or do they represent older Archean sialic crust moved in plate tectonic fashion from original marginal sites to subjacent orogenic sites before upward intrusion into the supracrustal piles? Again we require more data and flexible viewpoints.

The proof of crustal shortening in Abitibi orogen is likewise uncertain. It rests primarily on the assumption of an originally more circular outline of the volcanic piles compared with their present elliptical shapes. This assumption is based on modern analogue as stressed in the text. Taken together with the prevailing east-trending undulating fold patterns this has led to the structural interpretation offered in the text. Supporting quantitative studies have been conducted in the structural laboratories of the University of Toronto by F. Schwerdtner (personal communication). Again, final proof will require further detailed studies in the field and laboratory.

ARCHEAN VOLCANISM AND SEDIMENTATION IN THE SLAVE STRUCTURAL PROVINCE

J.C. McGlynn and J.B. Henderson
Geological Survey of Canada, Ottawa

ABSTRACT

Archean volcanic and sedimentary rocks of the Yellowknife Supergroup within the Slave Province occur in a number of discontinuous northerly trending belts separated by large areas underlain by granitic rocks. Volcanic sequences that in part border these belts are composed of mafic lavas with both calc-alkali and tholeiitic differentiation trends and are commonly capped by more silicic volcanic and pyroclastic rocks. The central parts of the belts are filled with volumetrically much more important greywackes and mudstones that have been interpreted as turbidites. The composition of the greywackes suggest derivation from a sialic source. Such paleocurrent evidence that is available indicates transport into the basins more or less perpendicular to the basin margins from areas now occupied by large granitic batholiths. With a few exceptions the contact between the sediments and the volcanics is conformable and in places is marked by a conglomerate composed mainly of angular volcanic clasts but commonly with well rounded granitic boulders. A few instances of apparent granitic basement to the Yellowknife rocks have been documented. The supracrustal rocks have been complexly folded, metamorphosed at a relatively low grade and intruded by granitic rocks of various ages.

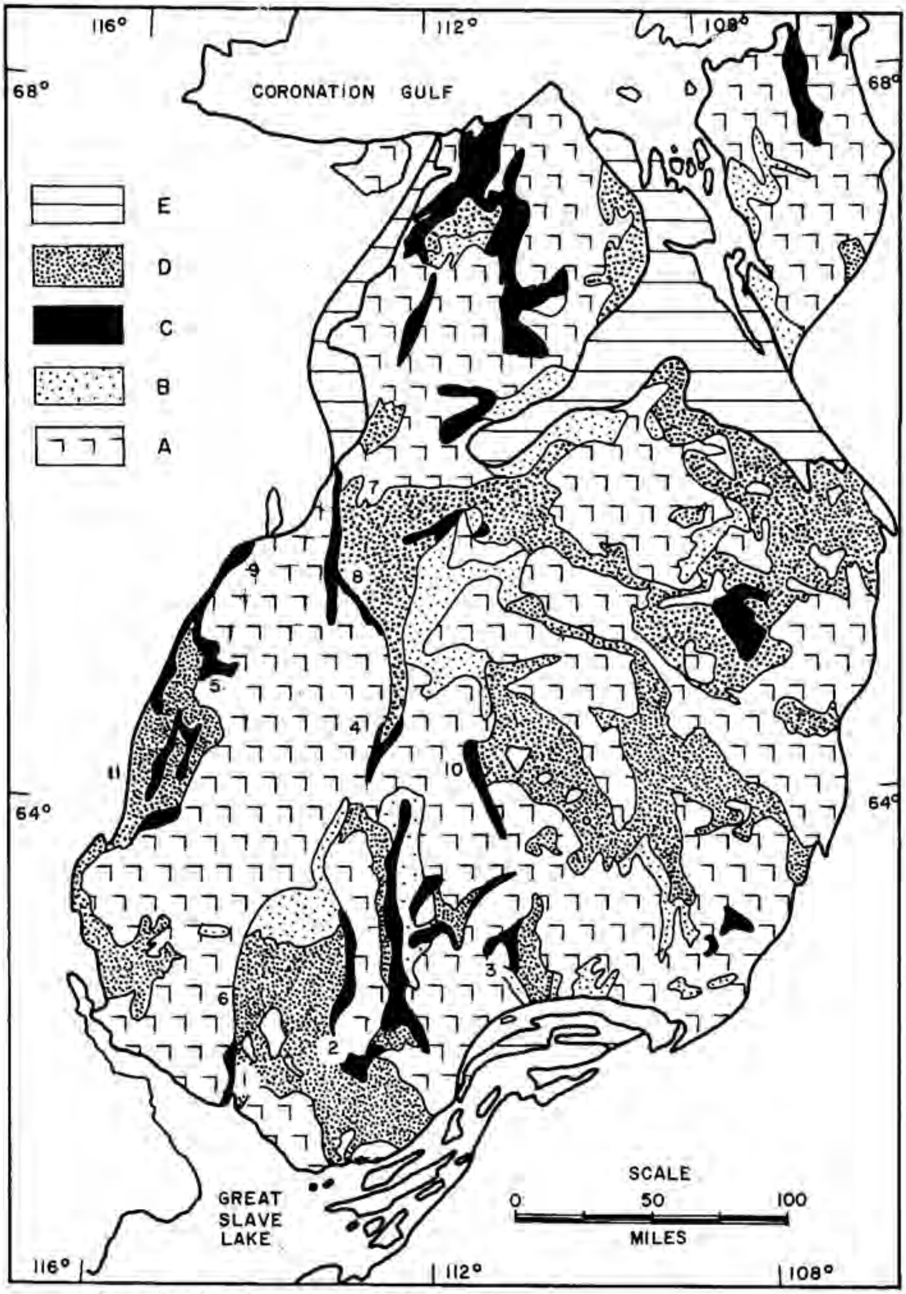
It is suggested that the Yellowknife rocks were deposited on a continuous, highly mobile sialic crust. The mafic volcanics were extruded from linear fractures paralleling contacts between positive and negative areas of the crust. Sediment derived mainly from the positive sialic areas was poured into the intervening negative basinal areas. Ultimately, extensive deformation and diapiric granitic intrusion at the margin of the basins has resulted in the situation seen today.

INTRODUCTION

The Slave Structural Province underlies some 75,000 square miles of the northwestern part of the Canadian Shield. The rocks of the province are for the most part Archean in age. To the northwest, southeast and northeast, the Archean rocks are unconformably overlain by Aphebian sediments of the Snare and Epworth Groups, the Great Slave Supergroup and the Goulburn Group respectively.

Of the total area, somewhat less than two-thirds of the Slave Province is composed of Archean sedimentary and volcanic rocks of their metamorphic equivalents (Davidson, 1967). The remainder consists of extensive granitic batholiths. Between 15 and 20 per cent of the stratified rocks are volcanic in origin while the remaining 75 to 80 per cent are of sedimentary origin. In this respect, the Slave Province differs markedly from the much larger Superior Province where Archean volcanic rocks are considerably more abundant than the associated sedimentary rocks.

The Archean stratified rocks of the Slave Structural Province all belong to the Yellowknife Supergroup (J.B. Henderson, 1970b). On a map of the province (Fig. 1), the Yellowknife rocks can be seen to occur in three major areas: (1) along the western margin of the province, (2) in a belt extending north from north of Great Slave Lake to the central part of the province and (3) in the large, more or less equidimensional area in the northeastern part of the province. These areas of volcanics and sediments are separated by large



A - Predominantly granitic rocks.
B - Mixed granitic and metasedimentary and metavolcanic rocks.
C - Volcanic rocks of the Yellowknife Supergroup.
D - Sedimentary rocks of the Yellowknife Supergroup.
E - Aphebian cover on the Slave Province.

1. Yellowknife area.
2. Ross Lake, Cameron Lake, Bealieu River areas.
3. Benjamin Lake area.
4. Winter Lake area.
5. Arseno Lake area.
6. Giauque Lake area.
7. Itchen Lake area.
8. Point Lake area.
9. Mesa Lake area.
10. Courageous Lake-Matthews Lake.
11. Ingray Lake area.

Figure 1: Generalized geological map of the Slave Structural Province

granitic batholiths and high grade metamorphic rocks. In addition there are several smaller occurrences of Yellowknife sediments and volcanics, notably north of the East Arm of Great Slave Lake and south of the Arctic Coast.

With a few exceptions, geological work in the Slave Province has been primarily of a reconnaissance mapping nature (Lord, 1942; J.F. Henderson, 1944; Fortier, 1949; Folinsbee, 1949; Fraser, 1969). Recently, however, more attention has been given to detailed studies of the geochemistry of the volcanics (Boyle, 1961; Baragar, 1966), geochronology (Green *et al.*, 1968), sedimentology (J.B. Henderson, 1970a), structure and metamorphism (Davidson, 1967) in particular areas of the province.

STRATIGRAPHY

The stratigraphy of the Yellowknife succession is broadly similar throughout the province and indeed is similar to the rather restricted assemblage that characterizes Archean stratified rocks throughout the world. This consists of a thick sequence of predominantly mafic volcanics followed by a sedimentary series of typically greywackes and shales. In the transition zone between these two dominant lithologies, minor amounts of conglomerate, more mature sandstones, limestones and tuffaceous sediments may or may not occur.

The Yellowknife rocks have almost always been divided into at least two units: a predominantly mafic volcanic unit and a sedimentary unit. In the past they were rarely named, but were commonly referred to by map-legend number, or as the Yellowknife sediments or volcanics. In some areas the sequence was divided into "divisions" (Jolliffe, 1942). For example, at Yellowknife the mafic volcanic rocks were referred to as Division A, while the more felsic flows and sediments were placed in Division B and C.

In some of the areas studied in greater detail it was found that the basic sedimentary and volcanic units could, in turn, be further subdivided into mappable lithological units. Thus, in some of the volcanic sequences, where the top of the sequence is more silicic in composition or contains a greater abundance of sedimentary material, one or more additional units were added. Similarly, with the sediments, conglomeratic or shallow-water crossbedded sandstones were separated from the predominant greywacke-mudstone unit. In the Yellowknife area, where the stratigraphy has been recently formally described, the Yellowknife Supergroup can be subdivided into six formations - three of volcanic origin and three of sedimentary origin (Henderson, 1970b).

It is generally accepted that the major basic volcanic units of the Yellowknife Supergroup are older, on the whole, than the sedimentary formations. Wherever the sedimentary-volcanic contact is observed, the sediments overlie the volcanics except at Mesa Lake (Ross, 1962) where the major basic volcanic unit is found between two thick sedimentary sequences. The sediments are typically conformable with the volcanics and, commonly, minor flows and tuffaceous beds are found immediately above the contact, suggesting that there was little or no time lapse between volcanism and sedimentation. At Yellowknife, however, sediments overlie the basic volcanics with angular unconformity. It is felt that this unconformity does not represent a major time break, but possibly a tilting of the volcanic pile on the thin relatively unstable Archean crust.

Where conglomerates occur in the sedimentary sequence, they almost invariably are at the contact between the volcanic and sedimentary units. This is certainly true at Yellowknife where the conglomerate fills large depressions on the surface of the unconformity. In most other areas where a conglomerate is reported, it lies with apparent conformity between volcanics and finer grained sediments. Examples of this relationship can be seen in the Arseno Lake (McGlynn and Ross, 1963), Beaulieu River (J.F. Henderson, 1938) and Winter Lake (Fraser, 1969) map-areas. In the Courageous-Matthews Lakes map-areas (Moore, 1956) conglomerate is reported within the major basic volcanic unit. At Point Lake conglomerate is interstratified with basic volcanics and in part overlies an older granitic body (Stockwell, 1933).

It is interesting to note the map distribution of the various units of the Yellowknife Supergroup within the Slave Province. As already mentioned, rocks of the Yellowknife Supergroup occur in three major zones with, in some cases, rather irregular diffuse borders. The sedimentary units underlie by far the greatest area and occupy the central parts of these zones. The volcanics, where present, are restricted to the margins of the zones. This is best exemplified by the southern part of the central zone (Fig. 1), where the volcanic rocks occur along the margin of the basin at Yellowknife in the west and along the Cameron River to the east. This suggests that these zones of Archean stratified rocks may represent basins of accumulation that were, in part, bordered by volcanic piles.

Table 1 shows estimated comparative thicknesses of the various units of the Yellowknife Supergroup in several different parts of the Slave Province.

Table 1

Comparative Unit Thicknesses in the Yellowknife Supergroup

Area	Basic Volcanics	Intermed. & Felsic Volc.	Cong. & Assoc. Seds.	Sediments
Yellowknife	40,000' ±	4,000'	0-800'	15,000' +
Benjamin Lake	2,000'	1,000'	-	?
Courageous Lake-Matthews Lake	12,000 - 2,000'	0-8,000'	0-400'	5,000'
Arseno Lake	18,000 - 1,500'	-	0-1,500'	2,000-3,000'
Mesa Lake	1,000'	-	-	2,000-3,500'

Estimates of thicknesses are often very difficult to make, due to extreme structural complexities. This is particularly true for sediments, as the top of the section is rarely defined.

BASEMENT TO THE YELLOWKNIFE SUPERGROUP

Despite the large areas of the province underlain by rocks of the Yellowknife Supergroup, very few observations have been made of what can be considered basement to the Yellowknife. Most contacts with non-Yellowknife rocks appear to be intrusive. However, a few possible documentations of basement have been made.

Baragar (1966) points out that a large zone of amphibolitic dykes in the granodiorite east of Ross Lake, on the eastern margin of the central area of Archean strata may represent feeders to the Yellowknife volcanics in the area. The dykes cut both the granodiorite and the volcanic strata but apparently not the overlying Yellowknife sediments. Thus, despite the fact that other small bodies of intrusive granitic rocks are present in the contact area, it is possible that much of the granodiorite in this particular area might represent basement.

Davidson (1967) using a similar argument describes a small granitic body exposed in the core of an anticline in the Benjamin Lake area, that may also be basement. Metamorphosed basic dykes that cut the granitic body and the possible pre-Yellowknife sediments that overlie it, were not traced into the overlying Yellowknife basic volcanic flows.

Another possible area of basement is located in the Beaulieu River area (J.F. Henderson, 1938) where a small, oval, highly altered and granulated chlorite granite body is surrounded by basic Yellowknife lavas. Along one contact between the granite and the volcanics is a fragmental zone that could represent a granitic conglomerate above the basement, although Henderson also suggested that it may represent a tectonic breccia zone.

The first discovered and best documented case for a basement to the Yellowknife Supergroup is described by Stockwell (1933) at Point Lake, in the north-central part of the Slave Province. Again, the rock in question is a highly altered chloritized granite that is intruded by other granitic bodies that cut both it and the overlying Yellowknife lavas and sediments. Immediately overlying the chloritized granite is a conglomerate that dips away from the granite body and whose cobbles are identical to it. Within the conglomerate there is a cobble size gradation away from the granite source and there appears to be no metamorphic gradient, as would be expected if the granite was intruded into the conglomerate. As well, basic dykes, bearing a close resemblance to some of the overlying lavas interstratified with the conglomerate, cut the older chloritized granite but were not observed to cut the younger granites in the area. It io interesting that this possible basement granite contains inclusions of a "dark green schist", possible evidence of the crustal material into which the Yellowknife "basement" granite intruded.

In all instances cited, the basement material is a strongly altered granitic rock. In addition these granitic rocks are all part of, or at least very close to the large batholithic areas of granodiorite that separate the various areas of Archean lavas and sediments. Although the number of instances of possible basement is very meager, the fact that they all are at, or at least close to, the margins of the supracrustal outcrop area supports the suggestion that these zones do indeed represent basins of accumulation and are not just fortuitous downfoldings of an originally much more extensive area. If this is true, the present day margins are a fairly close approximation to the original margins.

VOLCANIC ROCKS OF THE YELLOWKNIFE SUPERGROUP

Volcanic rocks of the Yellowknife Supergroup underlie about five per cent of the Slave Structural Province. They occur for the most part in 18 or 19 discreet belts that trend north to northeast. Most of the volcanic belts occur along the margins of areas underlain by sediments and are most abundant in the western half of the province.

The volcanic rocks lie at the base of the Yellowknife succession as discontinuous belts, except at Mesa Lake (Ross, 1962) where sediments occur below volcanic rocks. In some areas, thin layers of volcanic rocks occur within the overlying sediments and may represent resumption of volcanism or may be contemporaneous with volcanism occurring in other parts of the volcanic complex.

The volcanic strata comprise massive mafic lavas (basalt or andesite), pillowed basic lavas and fragmental rocks, and intermediate to acidic lavas and tuffaceous rocks that include dacites, latites and quartz latites. The intermediate to acidic strata occur in the upper part of individual piles. Sills, dykes, and irregularly shaped masses of gabbro and diorite occur in the volcanic rocks and, less commonly, in overlying sediments. They are thought to be intrusive equivalents of the lavas. Dykes of acidic porphyry also occur in the lavas, and probably represent intrusive equivalents of the acidic lavas.

The stratigraphy or anatomy of only a few complexes is known in any detail. At Yellowknife (J.F. Henderson and Brown, 1966; Baragar, 1966) 40,000 feet of volcanic strata are exposed. The succession comprises massive, pillowed and variolitic basalts and andesites, with intermediate and acidic layers at about mid point, and at the top of the sequence. Latite, dacite and quartz latite comprise 15 to 20 per cent of the strata (Baragar, 1966). In the Cameron River belt (Baragar, 1966) where about 7,000 feet of volcanic rocks are exposed, massive and pillowed basalts and andesites form the bulk of the succession and a discontinuous layer, up to 1,400 feet thick, of latite, dacite and quartz latite flows and fragmental strata occurs in the upper part of the unit. At Arseno Lake (McGlynn and Ross, 1963) 18,000 to 1,500 feet of lava are composed of basic massive and pillowed basalts and andesites with minor amounts of more acidic material. In the Benjamin Lake area, 2,000 feet of massive and pillowed flows are overlain by about 1,000 feet of latites and acidic or intermediate tuffs and breccias. Available data on other volcanic belts in the Slave Structural Province suggest that relative proportions of basic and acidic lavas, and stratigraphic position of the acidic layers are similar to the successions known in greater detail.

Baragar (1966) in his study of the geochemistry of volcanic rocks at Yellowknife and along the Cameron River recognizes two differentiation trends, namely a calc-alkali trend, involving enrichment in alkaline (plus quartz) components, and a tholeiitic trend, characterized by enrichment in iron at the expense of magnesium. The lavas, therefore, have mixed tholeiitic and calc-alkaline characteristics. Baragar postulates that tholeiitic basalts represent the fundamental magma type, with calc-alkaline trends resulting from sialic wallrock contamination.

The thickness of the numerous volcanic belts is variable. The thickest measured section is about 40,000 feet and as much as 18,000 feet has been estimated in other belts; however, maximum thicknesses of 5,000 to 10,000 feet are more representative. In most belts, the thickness varies considerably and appears to decrease along strike. In longitudinal section, a pinch and swell effect would be evident, with the belts thinning out away from centers of volcanism. In addition, it is thought that the lavas wedge out across their strike under the sediments, and that the sedimentary basins are not everywhere underlain by volcanic rock. The volcanic rocks were extruded from centers along linear zones of weakness in the crust, and thinned out in all directions from these centers.

SEDIMENTARY ROCKS OF THE YELLOWKNIFE SUPERGROUP

Archean sedimentary rocks or metamorphic rocks derived from sedimentary rocks underlie a large proportion of the province, and certainly greatly exceed the area underlain by volcanic rocks. Yet despite their great abundance, they are quite restricted as to type, being predominantly immature greywackes and mudstones with only minor amounts of conglomerate, limestone and more mature sandstones. These sediments show for the most part a remarkable similarity throughout the province, both within and between the main basins of accumulation.

For discussion purposes the sediments will be broken down into three groups: (1) conglomerates, (2) greywacke-mudstones, and (3) other sediments.

Conglomerates

Conglomerates where present, are characteristically found in rather thick sections between the thick basic volcanic sequences and the greywacke-mudstone facies. In most areas, the transition from volcanism to sedimentation is not marked by a conglomerate. Good examples of the conglomerate occur in the Arseno Lake and Winter Lake areas and at Point Lake, Ross Lake and Yellowknife.

Conglomerates occur in elongate lenses ranging in thickness from only a few feet to over 1,500 feet in some cases, and can be traced laterally for over ten miles. Some occurrences, however, are much more discontinuous, with other sediments occurring between the pockets or lenses of conglomerate. In some instances, conglomerates occur in areas where the associated mafic volcanic sequence thins. At Ross Lake, the Cameron River volcanic belt thins out from a maximum thickness of 8,000 feet and disappears over a distance of about four or five miles. It is in this area of thinning and south of it that the conglomerate is found. A similar situation is seen at Point Lake. This may suggest that the conglomerates tend to be deposited on the flanks of the volcanic piles.

Boulders in the conglomerates are mainly of volcanic origin. Their angular nature and compositional similarity to the underlying mafic volcanics suggest local derivation. The deposits commonly grade upwards from coarse angular mafic blocks, often several feet in length, to finer and commonly more felsic varieties. Cobbles and pebbles of greywacke, shale, quartz and chert are minor components but locally can be abundant. In addition, many of the conglomerates contain well rounded pink to grey granitic cobbles - a minor but quite spectacular component. The size of the clasts varies from less than an inch to several feet in the case of the angular basal boulders. The average size ranges between three inches and one foot, and except for the basal clasts, the cobbles, in particular the granitic boulders, show considerable rounding.

The matrix of the conglomerates is quite variable, ranging from a mixture of quartz and finely comminuted basic volcanic rock fragment through quartz and silicic volcanic rock fragments, to greywacke and, in one case, a calcareous and quartzose mixture. Again, the preponderance of volcanically derived material is of note. In many of the conglomerates thin discontinuous lenses or patches of sediment identical to the matrix of the conglomerates is found. These frequently show evidence of vigorous current activity with abundant cross-bedding, scours and ripples.

The conglomerates for the most part are conformable with the underlying volcanics and at Point Lake and Courageous-Matthews Lake the conglomerate is interbedded with volcanic flows. One exception is the conglomerate at Yellowknife which overlies the mafic volcanics with angular unconformity and fills depressions in the erosional surface that, in some cases, are several hundred feet deep. The general consensus of opinion is that although there was some erosion (most of the clasts in the conglomerate are locally derived), the conglomerate does not represent a long period of erosion after the major volcanic event prior to the principal period of sedimentation. The presence of a few granitic boulders in the conglomerate does not necessarily require an orogeny involving the intrusion and subsequent unroofing of great tracts of granitic rock prior to sedimentation. The contrast between the hard, very well rounded granitic cobbles and the relatively soft, angular, apparently locally derived, volcanic clasts suggests derivation of the granitic cobbles from a more distant source. This would preclude derivation from local leucocratic stocks in the volcanic pile as is seen in some recent volcanic sequences such as in the Aleutian Islands (Coats, 1956). There, conglomerates contain angular volcanic and granitic clasts both of local origin. A more likely source for the granitic cobbles in the Archean would be a granitic basement area.

Greywacke-Mudstone Facies

The greywacke-mudstone couplets make up by far the largest proportion of the sediments and, indeed, the greatest part of all the Archean stratified rocks. They typically overlie or apparently overlie the thick sequences of mafic volcanics and, where present, the conglomerates. The contacts are conformable and are commonly gradational; the volcanics giving way through a series of interbedded flows and tuffaceous beds to the typical greywackes and mudstones. Where associated with conglomerates, gradation is through a decreasing number of beds and lenses of ever finer grained conglomerate and cleaner sandstones.

Regionally these sediments occupy the large central areas of basins rimmed in part by the mafic volcanic sequences, and surrounded by granitic areas. Because of structural complexity, lack of good marker horizons, and difficulty in defining the true top of the unit, thicknesses are difficult to determine. Estimates range from one to two thousand to in excess of 15,000 feet.

This facies is made up of interbedded greywacke and mudstone with greywacke greatly predominating in most cases. The individual greywacke-mudstone couplets range in thickness from less than an inch to thick massive units measurable in tens of feet. It is commonly found, however, that many of the excessively thick units are composite in nature, with two or more greywacke units occurring together without the intervening mudstone.

The less metamorphosed sediments present a magnificent array of sedimentary structures. A high proportion of the greywacke beds are graded - a very useful feature for determining the complex structure of these areas. In addition the Bouma sequence of sedimentary structures (Bouma, 1962), so commonly noted in turbidites is also present in these rocks. This sequence includes a basal graded division followed by a parallel laminated division above which is a current ripple-laminated division. A second or upper division of parallel lamination is next, followed finally by the uppermost pelitic division.

The base of each individual bed is invariably sharp but is not necessarily straight or even. Many beds show evidence of minor scouring due to current activity or small channels that cut down through underlying beds. In addition, the bases of some beds made very irregular by the development of load and flame structures due to post-depositional soft sediment deformation.

Individual beds are laterally continuous. Pinching out of beds is rarely observed, although in places a pair of greywackes can be traced laterally into a single amalgamated bed. Even the thinnest of beds can be traced across very large outcrops (up to several hundred feet). Tremblay (1952) in the Giauque Lake area reported that when groups of thicker greywacke beds were traced along strike over several thousand feet, they sometimes are seen to grade into much more thinly bedded greywackes with a higher proportion of mudstone.

The sediments have been classified according to the thickness and proportion of greywacke to mudstone. Although the various types are commonly mixed, they do sometimes occur in groups of predominantly one type. In some sections, a zone of thin beds with a high shale to sand ratio will be followed by a group of very thick massive greywackes.

The greywackes are composed of a sand-sized fraction of quartz, rock fragments and feldspar in a matrix of chlorite, muscovite, and very fine grained quartzo-feldspathic material. The mudstones have a composition similar to the greywacke matrix. Depending on the relative proportions of the three major components, the sandstones have been variously called impure quartzites, greywackes and arkoses. At Yellowknife, the greywackes have a relatively high proportion of silicic and intermediate volcanic rock fragments which suggest that silicic volcanics were an important part of the source area. In addition, a minor but persistant component of granitic rock fragments is present. The compositional make-up of the greywackes is similar at Gordon Lake, suggesting again that a silicic volcanic source was a very important contributor to these sediments.

At Yellowknife and in the Mesa Lake area, paleocurrent measurements have been made in these sediments. For the most part measurements have been made on small-scale cross-laminations of the ripples or small cross-laminated scour fills which, when rotated back to overcome orientation errors caused by folding, give a good approximation of the depositional current sense. Considering the extreme degree of deformation in these Archean sediments, too much significance should not be placed on the exact current azimuths that result as there are bound to be considerable errors in the unfolding of the beds. The results are probably significant if not taken to indicate more than the general quadrant from which the currents came. At Yellowknife, the sediments appear to have been derived from a source to the west - a direction more or less perpendicular to the present-day margin of the basin, suggesting that the sediments were being shed from a land mass in the position now occupied by the western granodiorite. In the Mesa Lake area (Ross, 1962) in another Archean basin west of the same large granitic area, the paleocurrent trend is towards the southwest, away from the granodiorite area, but at more of an angle than was the case at Yellowknife. Ross suggests the direction may indicate the axis of a trough along which the currents were flowing. More information of this type would certainly be useful in outlining the shape of the original Archean basins.

The great proportion of the Yellowknife sediments are greywackes, believed to be deposited as a result of turbidity currents. The fact that the sediments are mainly greywackes composed predominantly of quartz, rock fragments and feldspar strongly suggests a source area of sialic composition. Indeed the few chemical analyses of these greywackes that are available show close similarities to those of a granodiorite. Greywackes being immature sediments, have undergone relatively minor sorting and concentration and probably represent a close approximation of the composition of their source area. Such paleocurrent evidence that is available indicates that the greywackes are derived from areas now occupied by granodiorite masses.

Other Sediments

Although the greywacke-mudstone assemblage makes up by far the greatest proportion of the sediments, other relatively minor sediment types are present.

Many of these other sediments occur close to the contact between the volcanics and the greywackes. Tuffaceous beds are most prevalent in this area and mark the transition from predominantly volcanic activity to sediment deposition. Thick tuffaceous units may also occur higher in the sedimentary section as is seen at Yellowknife and Arseno Lake.

A unit which is frequently reported as an impure quartzite is commonly found in the contact area. This rock occurs in well defined beds that are either massive or exhibit large-scale tangential crossbedding. In some cases this unit is interbedded with the normal greywacke-mudstone facies. At Giauque Lake (Tremblay, 1952), a thick section of this impure quartzite is interbedded with minor amounts of greywacke while at Benjamin Lake tangentially crossbedded units between six inches and five feet thick are a minor occurrence in the greywacke-mudstone section (Davidson, 1967). At Yellowknife a sandstone composed of quartz and silicic volcanic rock fragments together with a patchy conglomerate overlies the mafic volcanic pile.

Where present, calcareous sediments are also found near the contact with the volcanics. Lord (1942) in the Ingray Lake area reports a transition from mafic volcanics through tuffaceous beds into ninety feet of crystalline siliceous limestone. In the Ross Lake area, the conglomerate has a calcareous matrix and a unit of impure limestone occurs below the transition to the normal greywacke-mudstone facies.

One notable feature of Archean sediments in the Slave Province is the general lack of iron formation. This is in strong contrast to the Superior Province where iron formation is a common, although usually volumetrically small facies in the Archean sediments. An exception to this paucity of iron formation is the Itchen Lake area (Bostock, 1967) where siliceous magnetic iron formation occurs locally within the greywacke facies.

STRUCTURE OF THE SLAVE PROVINCE

Rocks of the Yellowknife Supergroup have been folded, metamorphosed and intruded by granitic rocks. This activity is considered by Stockwell (1964) to result from an orogeny that has been assigned to the Kenoran orogeny. Most of the K-Ar age determinations on minerals formed during this orogeny fall in the range of 2,300 to 2,600 m.y. and their mean value is 2,490 m.y.

Rocks of the Yellowknife Supergroup were deformed into northerly to northeasterly trending similar folds that plunge moderately north and south. The more northerly trends are common in the western half of the area whereas the more easterly trends occur in sediments of the eastern part of the province. These folds in many areas were mofidied by steeply plunging cross-folds. The sedimentary rocks are generally tightly and intricately folded but volcanic rocks, probably because of their greater competency, behaved differently. Thin volcanic successions are folded into broad folds parallel to those in the sediments but thicker successions occur as homoclinal sequences that face away from bounding granitic rocks that have been presumed to be younger. An alternative interpretation is that the homoclinal successions are in contact with granitic rocks that were part of the basement and that have been somewhat remobilized during deformation of the supracrustal rocks.

The Yellowknife strata, particularly the volcanic sequences, are cut by chloritic shear-zones that may be contemporaneous with folding and also by later high-level faults that strike north, northwest and northeast and seem to be of several ages.

About two-thirds of the Slave Structural Province is underlain by granitic rocks of which possibly as much as half are migmatite, mixed gneiss, banded gneiss and granitic gneiss, with the remainder consisting of batholiths of massive or slightly gneissic granitic rocks of various compositions. In a number of areas, Yellowknife sediments are separated by zones of mixed gneisses that were derived, at least in part, from the supracrustal rocks. It seems likely that they were parts of single larger basins.

Yellowknife volcanic and sedimentary rocks have been metamorphosed to degrees varying from low greenschist to amphibolite facies.

In the areas where detailed work has been completed it can usually be shown that deformation, metamorphism and intrusion of granites were elements of a sequence of overlapping events that include folding and low grade metamorphism followed by granitization, intrusion of syntectonic granitic rocks and associated metamorphism that overlaps in time with the early folding, later cross-folding and intrusion of high-level late tectonic granodiorites and pegmatites.

CRUSTAL MODEL FOR THE SLAVE STRUCTURAL PROVINCE

The authors' model of Archean deposition in the Slave Structural Province has been strongly influenced by the work of Martin (1969) and Anhaeusser *et al.* (1969) among others.

A model to explain evolution of this province during the Archean must take into account the following features:

(1) The areas of Archean volcanic and sedimentary rocks are separated by large granodioritic areas.

(2) Where volcanics are present they tend to occur at the margins of areas underlain by stratified rocks.

(3) The sedimentary rocks, which greatly exceed the volcanics in volume, have on the average a granodioritic composition.

(4) All areas that have been suggested as possible basement to the stratified rocks are granitic in nature.

(5) Areas of sedimentary and volcanic rocks have a dominant north to north-northeasterly trend, particularly in the western, central and southern parts of the province.

The zones of Yellowknife supracrustal rocks probably represent basins of accumulation and the present-day borders to these areas are a close approximation to the original margins. That is to say these areas do not merely represent preserved "keels" of large synclines that originally covered a much more extensive area. Between these negative basinal areas must have been positive areas that are now occupied by extensive granitic batholiths. Volcanic activity occurred in large part at the contacts between positive and negative areas in the crust. Only rarely is there any evidence of volcanic activity in the central part of the basins.

The meager paleocurrent evidence that is available suggests that the sediments were derived from the interbasinal positive areas. The sediments are predominantly greywackes and mudstones; chemically they approximate a granodioritic composition indicating that the source area was sialic. Further evidence that sialic material was available is indicated by the scattered, well rounded granitic cobbles found in the conglomerates at the margins of the basins. It is possible that the granitic cobbles could have been derived from local minor intrusions into the mafic volcanics that typically underlie them. However, the high degree of rounding of these granitic boulders, in contrast with the rather angular, locally derived, volcanic clasts would suggest a somewhat more distant source. In this regard it is significant that in all places where possible basement to the Yellowknife Supergroup has been described, basement rocks are granite.

The marginal volcanic sequences are not regarded as a major contributor of sediment although erosion of the comparatively thin silicic top of some of the volcanic piles is of local importance. The composition and volume of the sediments would require tremendous quantities of silicic material - far beyond the capabilities of the essentially mafic volcanic belts.

It appears likely that a positive terrain of sialic composition did exist during Archean times and from the evidence of the relict basement and granitic cobbles in the Yellowknife Supergroup that at least some of the basement was of plutonic or hypabyssal aspect. This terrain could conceivably have consisted of intruded and deformed remnants of pre-Yellowknife sedimentary and/or older volcanic sequences. With the present information available it is not possible to precisely define such areas but it is felt that the granitic terrain west and north of Yellowknife up to the Arctic coast and the area of granitic rocks between the Cameron River and Benjamin Lake north of the East Arm of Great Slave Lake may be two examples.

It is believed that the Yellowknife basins were completely underlain by sialic crust throughout their history and hence contrast with many of the more recent geosynclines that have developed at continental-oceanic crustal junctions. The relatively small size of many of the zones that are surrounded by granitic batholiths makes it difficult to conceive of them being underlain by simatic crust if the batholiths represented sialic areas at the time of basin filling. The fact that the volcanic sequences show evidence of sialic contamination would also suggest intrusion through a sialic crust. It seems unlikely therefore that the Archean sequences in the Slave Province are analogous to island arc systems.

In Archean time the Slave Structural Province is thought to have been underlain by a continuous highly mobile sialic crust. This mobility may have been a result of the crust being much thinner than present day continental areas or possibly due to a much steeper thermal gradient resulting from a higher rate of radioactive decay (Urry, 1949). Parts of the crust were uplifted, forming ridges and concomitant negative areas that have been preserved in the northerly trends seen in the province today. Deep fractures or zones of weakness developed parallel to the positive areas and along these the volcanic rocks were extruded. In response to the uplift of the positive areas, silicic debris was eroded and deposited in the basinal areas between the positive zones. Volcanic activity was concentrated along parts of the margins of the basins while the sediments flowed into the basin between volcanic areas in a series of large fans (Fig. 2).

With continued accumulation in the basins these areas were depressed into the mobile crust, resulting in the eventual deformation of the supracrustal rocks. Diapiric granitic bodies were intruded preferentially between positive and negative regions of the crust, which resulted in intrusive relationships between the Yellowknife rocks and the batholithic areas (Fig. 3). Deformation and intrusion took place at relatively shallow depths in the crust (possibly as a result of its mobility) as shown by the low pressure metamorphic effects (Davidson, 1967). In some instances deformation and intrusion were much more extensive, resulting in the partial obliteration of parts of basins. Thus the depositional areas may have been much more extensive than is indicated now.

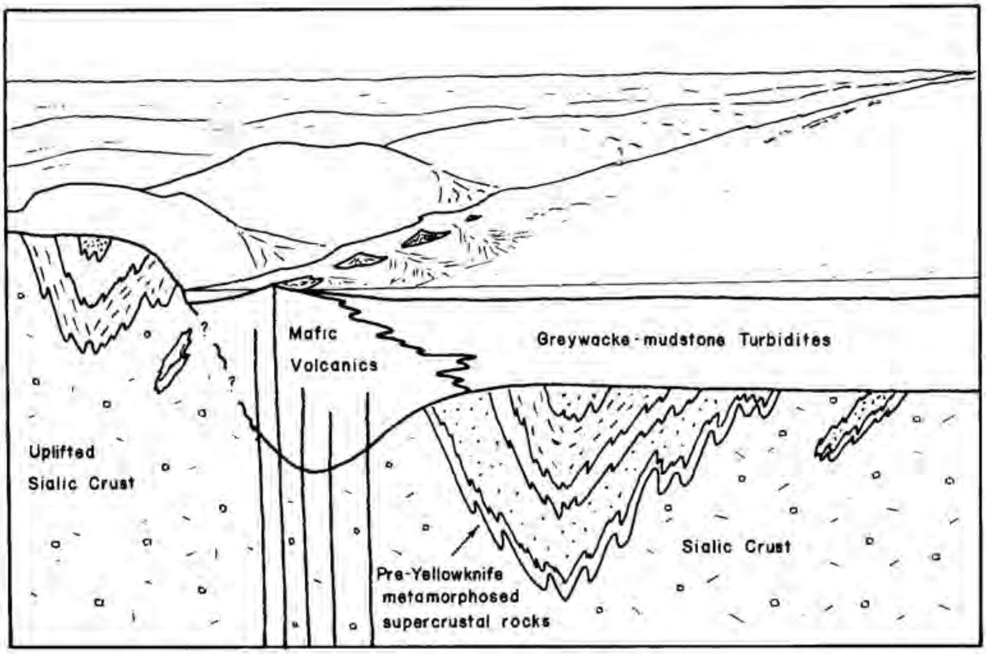


Figure 2. Cross-section through part of an Archean basin prior to deformation.

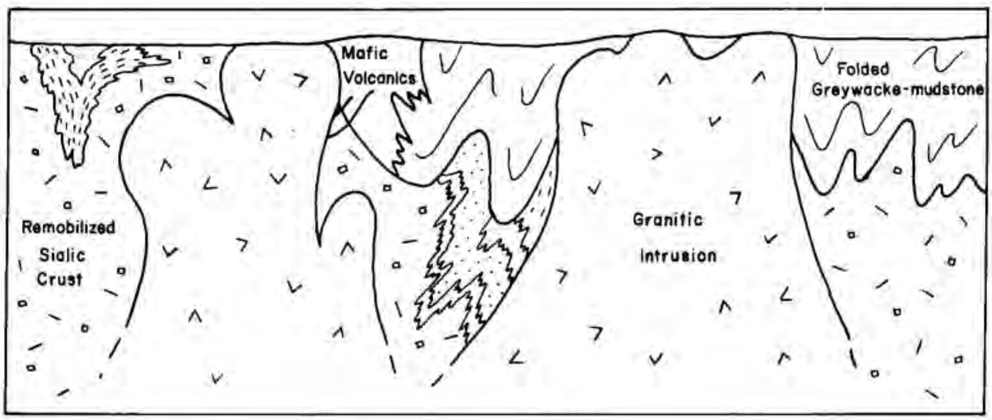


Figure 3. Cross-section through the same basin after deformation.

This cycle may have been repeated over a long period of time with basins being filled, deformed and eventually cannibalized to fill newly developing basins. As suggested by Martin (1969) the 2,500 m.y. number that keeps appearing in Archean terrains may reflect the achievement of stability for unknown reasons by the up-to-then mobile crust. What we see preserved in the Archean is the last of a long series of cycles that operated under conditions that have never again existed on the earth.

The exercise of developing this model or synthesis of the Archean geology of the Slave Province, based as it is on rather scattered evidence, emphasizes the need for further work in this area. Particularly the granitic areas in the province need re-studying to test for the presence of basement rocks, to outline their extent, and to study the nature of younger granitic rocks and thereby place them in an orogenic sequence. The composition and stratigraphy of many volcanic complexes needs to be determined; and sedimentological and paleocurrent studies in the sedimentary rocks of the Yellowknife Supergroup are required to define the extent and nature of the original basins.

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THE FILLING OF THE CIRCUM-UNGAVA GEOSYNCLINE*

E. Dimroth, Department of Natural Resources, Quebec
W.R.A. Baragar, Geological Survey of Canada, Ottawa
R. Bergeron, Department of Natural Resources, Quebec
G.D. Jackson, Geological Survey of Canada, Ottawa

Editor's Note: The size and scope of this paper reach well beyond the framework set for presentations in this workshop. Rather than asking for a drastic modification of the manuscript, the editor has accepted it because it presents abundant unpublished information and is the first, most comprehensive report of its kind on this major Proterozoic geosyncline.

A.J.B.

Abstract

The Labrador Trough, Cape Smith Belt and Belcher Fold Belt are erosional remnants of an Aphebian geosyncline that surrounds the Superior Province of Quebec (named here the Ungava Craton) in a wide arc open to the south. Its filling rests on an Archean basement. Ideally the filling is strongly asymmetric, consisting mainly of sedimentary rocks (comprising much orthoquartzite, dolomite and ironstones) in the zone towards the craton, whereas mafic volcanic and intrusive rocks predominate in the zone away from the craton. Erosion has removed large parts of the filling so that only the mainly volcanic zones remain in the Cape Smith Belt and in the northernmost part of the Labrador Trough.

The geosynclinal filling tends to be cyclic. In the Central Labrador Trough cycles begin with deposition of quartzites and precipitate sediments. They culminate with deposition of shales and greywackes in the frontal zone (i.e. towards the craton) and in intense basaltic volcanism in the distal zone (away from the craton). Facies and stratigraphy of the Cape Smith Belt and of the northernmost Labrador Trough correspond to the facies of the distal zone farther south. In the Belcher Fold Belt cycles begin with volcanism and terminate with deposition of quartzitic and precipitate sediments in what is probably the frontal part of the geosyncline. The distal zone may be to the northwest in Hudson Bay.

The paleogeography of the oldest formations in the Central Labrador Trough is examined in some detail. Much clastic material is probably derived from a source area extending east of the geosyncline. A ridge emerged later in the median part of the Trough, and shales and greywackes were deposited in two separated basins in the west and in the east of the Trough. Faults formed. Subsequently the central and eastern zones of the geosyncline subsided rapidly, and were filled by ophiolites. Finally conditions reverted to deposition of orthoquartzites, dolomites and iron-formation of the second cycle. These rest unconformably on the older rocks. There was no orogeny between cycles.

In the Belcher Belt the centres of volcanism lay first to the north and later to the west of the Belcher Islands. In the eastern part of the belt much of the clastic filling was probably derived from craton to the east.

The volcanic filling of the Labrador Trough is extraordinarily "primitive", and except for moderately low alumina approaches oceanic tholeiites in composition. It is more basic than the mafic fraction of Archean volcanic

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belts, and contains almost no rhyolites and few intermediate rocks. A highly feldspathic variety (leopard rock) is a characteristic component of the assemblage. The volcanism of the Cape Smith and Belcher Island segments of the geosyncline seems to have been similar. About half of the volcanic filling of the geosyncline occurs as sills, which may be attributed to the lateness of the bulk of the volcanism.

The Labrador Trough segment of the Circum-Ungava geosyncline provides one of the best and most complete examples available of a Precambrian geosyncline, and is similar to certain Alpine geosynclines.

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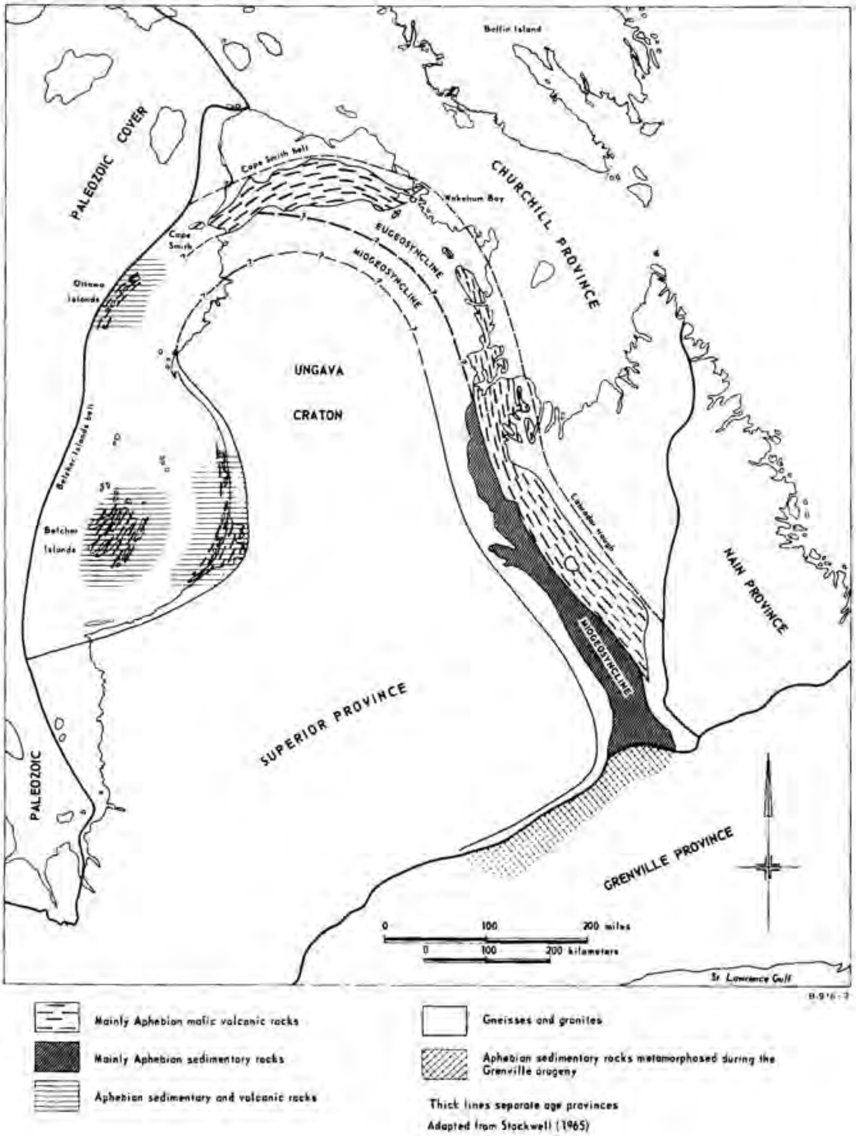


Figure 1. Index map

INTRODUCTION

The Circum-Ungava geosyncline surrounds the Superior Province of Quebec in a wide arc open to the south (Fig. 1). It rests on an Archean basement, and has been folded and metamorphosed during the Hudsonian orogeny. It is therefore of Aphebian (Early Proterozoic) age.

Post-Hudsonian uplift and erosion have removed large parts of the geosynclinal filling, so that only three separate geological units now remain. These are:

1. The Belcher Fold Belt, comprising the Belcher and Ottawa Islands, and the adjoining parts of the mainland at Richmond Gulf. This belt trends roughly north. Inliers of Proterozoic rocks on the west side of Hudson Bay mark the westward extension of this belt beneath Paleozoic strata of the Hudson Bay lowlands.

2. The east trending Cape Smith-Wakeham Bay Belt in the northern part of Ungava Peninsula.

3. The Labrador Trough extending in a general south-southeasterly direction from Payne Bay to Wabush and beyond.

All three units have a common foreland composed of Archean gneisses and granites of the Superior Province. This foreland will be referred to as the Ungava craton. The hinterland north of the Cape Smith Belt and east of the Labrador Trough is also mainly composed of Archean gneisses that underwent a second phase of metamorphism and deformation during the Hudsonian orogeny. The hinterland of the Belcher Fold Belt is presumably submerged below Hudson Bay and possibly below the Paleozoic rocks of the Hudson Bay lowland.

In order to describe the relative positions of facies zones in respect to the surroundings of the geosyncline we have adopted the following terminology: The part of the geosyncline close to the craton is named "frontal" and the part close to the hinterland "distal". A "median" part is distinguished where necessary. The terms "external" and "internal" designate the equivalent zones in Alpine chains, but appeared to be ambiguous because the craton adjoins the inner arc of the Circum-Ungava geosyncline, not the outer arc as is the case in the Alps.

The filling of the geosyncline tends to be cyclical. Well washed sandstones and chemical sediments form the base of each cycle in the central and northern Labrador Trough and in the Cape Smith Belt. Shales and greywackes follow upwards and, they interfinger with mafic extrusive and intrusive rocks towards the distal part of the geosyncline. The shales grade upwards into a new precipitate phase in the centre of the basin and the orthoquartzite-precipitate phase of the next cycle overlaps the older stratigraphic units at the margin. Such "ideal" cycles are variously modified in other parts of the geosyncline depending on the local paleogeography.

All three belts contain lithologies that are absent or rare in Archean belts, in particular well-sorted sandstones, dolomites and cherty ironstones.

The Labrador Trough is by far the best preserved of the three belts, and, because it has also been more intensely studied than the other units, it will be described first. Descriptions of the Cape Smith Belt, and of the Belcher Fold Belt will follow.

Previous Investigations

Bergeron (1957a, b, 1965), Fahrig (1957), Gastil *et al.* (1960), Jackson (1960), Baragar (1967), Dimroth (1968b), Sauvé and Bergeron (1965) and Hoffmann and Jackson (1969) have outlined the history of exploration of the various parts of the Circum-Ungava geosyncline. The Geological Survey of Canada mapped the Belcher Islands and most of the Labrador Trough south of 58°00', generally at a scale of one inch equals four miles. The Department of

Natural Resources of Quebec mapped parts of the Cape Smith Belt, the parts of the Labrador Trough north of 58°00' and between 56°30' and the height-of-land (ca. 55°00'), and the parts of the Southern Labrador Trough within Quebec, generally on a scale of 1 inch equals 1 mile. Much geological work has been done by mining companies, particularly in the Southern and Central Labrador Trough. A great number of publications and of unpublished theses are concerned with aspects of the geology of the region; they will be referred to in the following pages.

Present Study

Contributions of each author to this paper are as follows: E. Dimroth wrote the introductory paragraphs, the section on the Central part of the Labrador Trough, the comparison between the three belts and the comparison with Alpine geosynclines. R. Bergeron described the Northern Labrador Trough and the Cape Smith Belt. W.R.A. Baragar contributed the sections on the igneous succession, on the eugeosynclinal-miogeosynclinal relationships, the chemical characteristics of the ophiolites and the comparison with Archean belts. G.D. Jackson described the Belcher Fold Belt and the Southern Labrador Trough. The descriptions are based on all material, published or unpublished, that is presently available.

THE LABRADOR TROUGH

General Features

The Labrador Trough extends for some 600 miles in a north-northwesterly direction from the Grenville Front at Wabush (53 N, 67 W) to Lac Roberts (61 N, 71 W). Two lithotectonic zones are continuous for most of its length; they are a western zone composed mainly of sediments and an eastern zone that is underlain predominantly by mafic intrusive and extrusive rocks. Both zones overlie Archean basement gneisses (Figs. 1 and 16). Southwest of the Grenville Front the metamorphosed equivalents of rocks of the Labrador Trough can be recognized for an additional 150 miles. This segment is called the Southern Labrador Trough in this paper following the practice of Gastil *et al.* (1960).

A broad anticlinorium, characterized by domes of granitoid basement gneisses separated by generally narrow synclines of metamorphosed Proterozoic rocks (para-gneisses, meta-quartzites, calc-silicate rocks, marbles, amphibolites and ultrabasic rocks) extends east of the Trough; the western edge of the anticlinorium defines the eastern boundary of the Labrador Trough.

The gneisses that form the basement of the western zone of the Trough did not undergo penetrative deformation during the Hudsonian orogeny. They therefore give Archean ages by the K-Ar method; their Archean age is undisputed.

The granitoid basement gneisses east of the trough, underwent intergranular and intragranular deformation during the Hudsonian orogeny. They were remetamorphosed and therefore give Hudsonian ages by the K-Ar method. De Roemer (1956) and Taylor (1968, 1969) assume them to be metamorphosed equivalents of the Labrador Trough rocks. The writers believe that they are Archean. Their arguments are:

1. The domes of granitoid gneisses are mantled by paragneisses continuous with Labrador Trough rocks (Gélinas, 1965; Dimroth, unpublished). The granitoid gneisses are therefore stratigraphically below Aphebian rocks. Arkoses and conglomerates containing gneiss fragments occur at the base of the Aphebian sequence on the eastern margin of the Trough east of Romanet Lake (56°15'N) (Dimroth 1964, and unpublished).

2. The granitoid gneisses of some of the domes suffered a retrograde metamorphism that correlates with the prograde metamorphism of the Labrador Trough rocks (Dimroth, 1964; Wynne-Edwards, oral communication). K-Ar ages presumably date this retrograde metamorphism. The retrograde metamorphism is superposed on a previous migmatization that is absent from the Labrador Trough rocks. The granitoid gneisses therefore suffered a high grade metamorphism in an orogeny previous to the Hudsonian Revolution.

3. Rb-Sr whole rock ages of the granitoid gneisses assuming an initial ratio of .710 vary from 2,700 to 2,100 million years. An isochron of all such samples yields an age of 2,400 million years (Beall *et al.*, 1963). These ages presumably date a Pre-Hudsonian metamorphism of the basement gneiss.

The Aphebian sequence of the Labrador Trough comprises two, or possibly three sandstone-precipitate-shale cycles. A marginal unconformity separates the lower two cycles, and erosional unconformities also occur within them. A continental red bed sequence deposited in a post-Kenoran fault basin underlies the geosynclinal sequence in the central part of the Trough.

The Labrador Trough has been subdivided into three segments for the purpose of this paper: The stratigraphy of the first cycle, and its relations to the second, are well exposed in the central part of the Trough, between 55° and 57°, and this segment will be described first. The rocks of the second cycle predominate in the section south of latitude 55°, including the Southern Labrador Trough within the Grenville Province. Units that appear to belong to a third cycle are best preserved in the part of the Northern Labrador Trough situated between 57°00' and 59°00'.

A tentative stratigraphic correlation along and across the Labrador Trough is presented in Table I. The correlation of formations between 54° and 56°30' is based on much detailed work (Harrison, 1952; Baragar, 1967; Dimroth, 1969; and unpublished), and appears to be well established. The sequence and correlation between 56°30' and 58°00' is generally based on work at the scales of one inch equals two or four miles (Bergeron, 1954; Fahrig, 1955, 1956a, b; Roscoe, 1957), and must be regarded as tentative. The assumed presence of a third stratigraphic cycle is based on the work done in the northern segment of the Labrador Trough. The sequence in the area north of 58°00' is again based on detailed work (Bérard, 1956; Sauvé and Bergeron, 1965), but correlation across the Trough is hampered by rapid facies variation and faulting. It is generally possible to relate the sequence in the southern part of the Central Labrador Trough and in the Southern Labrador Trough to that farther north, but due to extreme structural complications there may be some doubt about the stratigraphic position of a few units.

The Central Labrador Trough

Stratigraphy

Introduction

The stratigraphical sequence of the Central Labrador Trough is presented as Table II based on Harrison (1952), Frarey and Duffell (1964), Baragar (1967), Dimroth (1968b, 1969), and on unpublished work. Work in this central zone is continuing and some details of the section are tentative. Dominant lithologies of the formations are indicated in the table. Thicknesses of some formations are shown on Figures 2 to 7. All stratigraphic units below the Wishart Quartzite belong to the first cycle; the second cycle comprises most of the units above the Wishart Quartzite (*see* Table I). A lenticular iron formation occurring in the Thompson Lake Slate may possibly correlate with the basal precipitate phase of the third cycle. The Thompson Lake Slate and Willbob

Table I: Tentative correlations along the Labrador Trough

Between 59° and 58°		Between 58° and 57°			Between 57° and 54°		
west (Bérard 1965)	east (Sauvé and Bergeron 1965)	west (Bergeron 1954, Fahrig 1955)	east (Fahrig 1955, 1956a)		west (Harrison 1952, Dimroth 1969)	east (Baragar 1967, Dimroth 1969)	
		conglomerate			Sims Quartzite		POST-HUDSONIAN COVER
	Thévénet Slate						CYCLE III
	Bellancourt Basalt		basalt			Willbob Basalt	
Larch River Slate		Larch River Slate	slate			Thompson Lake Slate and Irene Lake iron formation	
		upper iron formation	upper iron formation	locally basalt and pyroclastics			
Abner Dolomite		Abner Dolomite	dolomite	?	?	?	
?		?					
						Murdoch Pyroclastics	CYCLE II
Chioak Formation (slate, conglomerate, greywacke)	Baby Schist	Chioak Formation (slate, conglomerate, greywacke)	basalt and pyroclastics			basalt	
Dragon Formation (slate, siltstone)			slate		Menihek	Menihek Slate	
Fenimore Iron Formation	iron formation	Fenimore iron formation	iron formation (local)		Sokoman Iron Formation	Sokoman Iron Formation	
Lower Slate	Baby Schist				Ruth Slate		
Allison Quartzite	Harveng Dolomite	Allison Quartzite	quartzite		Wishart Quartzite	Wishart Quartzite	
NOT PRESENT		NOT PRESENT			Attikamagen IV Slate		CYCLE I
					Denault Dolomite	Denault Dolomite	
					Attikamagen III Slate	basalt-shale sequence (Attikamagen II)	
					Swampy Bay Subgroup (Attikamagen I)	Swampy Bay Subgroup	
					Pistolet Subgroup	Pistolet Subgroup	
					Seward Subgroup	Seward Subgroup	
ARCHEAN		ARCHEAN			ARCHEAN		BASEMENT

Basalt might therefore be parts of the third cycle although it is equally possible that the relatively unimportant slate and the local iron formation of the Thompson Lake Formation mark only a brief interruption of volcanic activity.

Formations of the First Cycle

Some of the stratigraphic relations of the first cycle have been previously discussed (Dimroth, 1968b, 1969) and need not be repeated in detail. In brief it was concluded that the Seward and Pistolet Subgroups were continuous across the Trough. The succeeding sequence (Swampy Bay Subgroup) was deposited in basins situated in the west and east of the geosyncline. A rising geanticline separated both basins. Later the geanticline subsided again and was buried beneath a succession of basalts and shales (basalt-shale sequence) which extended over the eastern basin. Strata of the eastern basin were tectonically exposed in a fault wedge in the vicinity of Romanet Lake (56°15'N; 67°40'W). They are probably present below the younger rocks elsewhere.

Relationships in the Pistolet and Seward Subgroups are well established. Contacts between all formations of the two subgroups have been observed. Horizontal correlations are based in part on continuous mapping, on lithological comparison of characteristic rock suites comprising many members, and on mappable facies variations.

Relationships in the predominantly argillaceous formations overlying the Pistolet Subgroup are more uncertain. The rocks lack characteristic key beds, and poor outcrop further complicates mapping of different lithological units. The correlation across and along the geosyncline is therefore based on indirect evidence.

The terminology of the argillaceous unit is complicated by the use of different names for the same unit or parts of it in separated regions of the Trough. The uppermost part of the argillaceous unit was called Attikamagen Formation by Harrison (1952) in the Schefferville region. Later Frarey and Duffell (1964) and Baragar (1967) following geologists of Iron Ore Company of Canada (private reports), extended the term so as to include all predominantly pelitic rocks that are below the Wishart Formation in the area south of latitude 56°N. Dimroth (1969) introduced the term Swampy Bay Subgroup for a well defined slate-greywacke suite that overlies the Pistolet Subgroup in the area north of 56°N. During the 1968 and 1969 field seasons Dimroth (unpublished) followed the units of the Pistolet and Swampy Bay Subgroups southwards. The Swampy Bay was found to be the equivalent of the lower part of the Attikamagen as redefined by Frarey and Duffell (1964). During this work Dimroth subdivided the Attikamagen into four gradational units of formational rank that have not yet been named and that are informally numbered in this paper. The lowermost of these, here called Attikamagen I, corresponds to Swampy Bay Subgroup. The distribution of the major lithological units of the Attikamagen is shown on Figure 2.

A graphitic slate (Hautes Chutes Formation) is at the base of the Swampy Bay Subgroup in the west of the Labrador Trough. It is overlain by laminated slate (Savigny Formation) that grades upwards and eastwards into the slates and greywackes of the Otelnuk Formation. Assemblages of slate, greywacke, grit and open work conglomerates (du Chambon Formation, Romanet Formation) are equivalent to the Swampy Bay in the separate, eastern segment of the Trough (Dimroth, 1968b).

South of 56°N in the western part of the Trough the Attikamagen I Formation is equivalent to the Swampy Bay Subgroup. It consists of graphitic slate, overlain by laminated slate, siltstone, and fine-grained sandstone that grade upwards and eastwards into open work sandstone, grit, and greywacke.

The Attikamagen III Formation overlies and is gradational into the Attikamagen I Formation in the western part of the Trough. It comprises grey and locally red and green silty slate with poorly visible lamination. In its upper part it contains some carbonate and not uncommonly beds and lenses of dolomite at the top of the formation. The Attikamagen III is overlain by the Denault Dolomite.

The correlation of Attikamagen III Formation with units in the central and eastern regions of the Trough is complicated by facies changes, structural discontinuity, and the presence of numerous dolerite sills. Nevertheless the relationships may be established indirectly. The basalt-shale sequence of Dimroth (1968b) consists of four litho-stratigraphic subunits: (1) 100-300 feet of basal sandstone (locally of grit and conglomerate); (2) about 1,000 feet of grey laminated slate and siltstone, locally with minor flows of pillowed basalt and some pyroclastics; (3) 0-2,000 feet of pillowed and massive basalt which lenses out westward; (4) grey laminated slate and siltstone, with a few local dolomite beds, and with beds of quartz wacke and of impure sandstone. Pillowed basalt and pyroclastics occur locally. Quartz wackes and impure sandstones form a characteristic local horizon close to the top of this member at a number of localities. Unit 4 is about 1,000 feet north of 56°00'. It is more than 3,000 feet thick at latitude 55°. The distribution of these subunits is shown in Figure 2.

The basal sandstone of the basalt-shale sequence rests on top of the Chakonipau, du Portage, Dunphy, or Lace Lake Formations. Member 2 is in fault contact with a complete section from the Chakonipau to the Romanet and du Chambon Formations at Romanet Lake (lat. 56°15', long. 67°45'). The member, as well as members (3), (4) and (5) are continuous over the Romanet fault wedge. It follows that the member 2 is younger than at least part of the Attikamagen I. Member 4 of the basalt-shale sequence plunges below the Denault Dolomite in a few folds south of 56°00' where contacts are clearly not faulted. It is therefore believed (a) that the four lower members of the basalt-shale sequence correspond to the Attikamagen III and possibly to upper parts of the Attikamagen I, and (b) that an erosional unconformity separates the basalt-shale sequence from the rocks below. Accordingly, the basalt-shale sequence is named the Attikamagen II Formation.

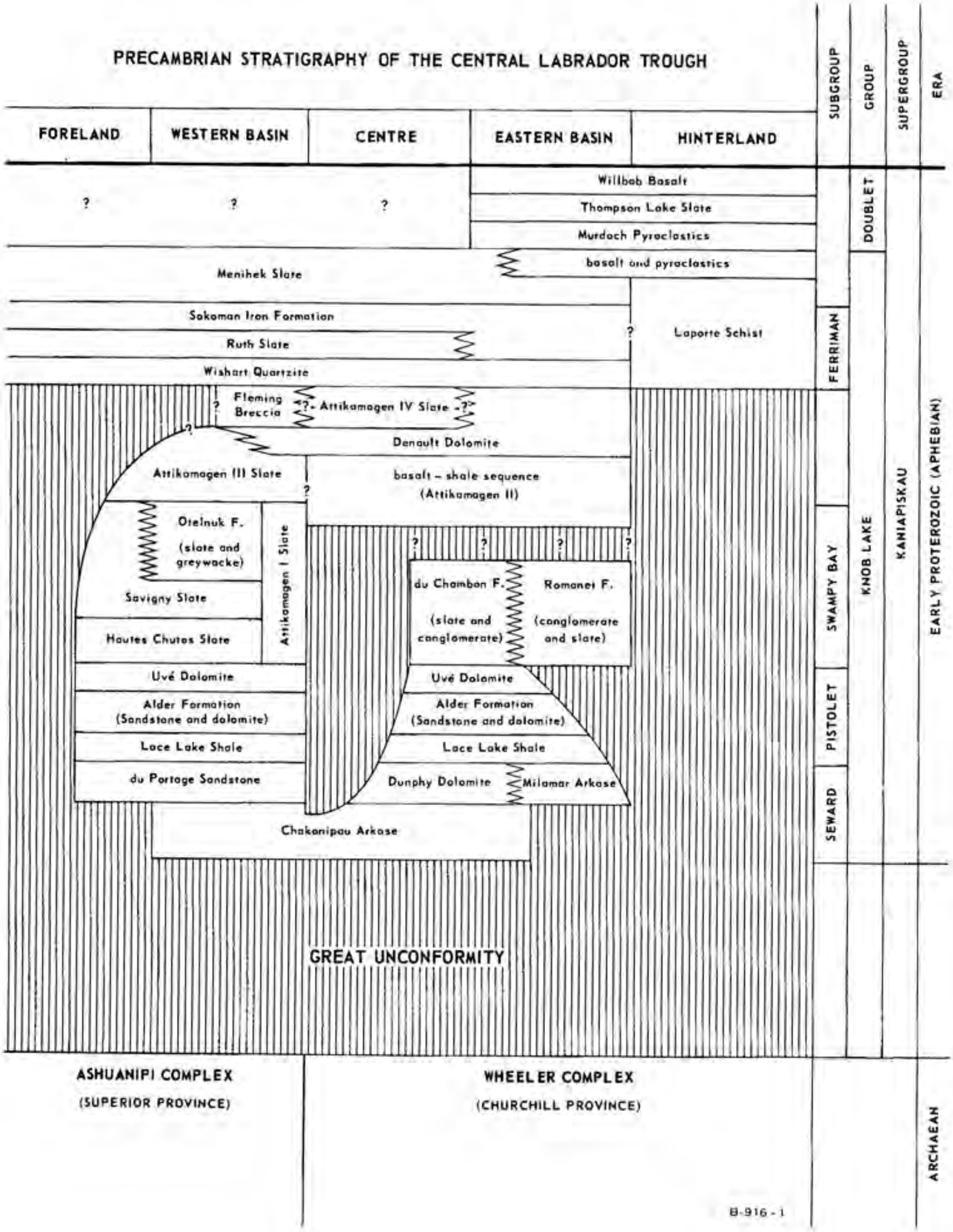
Some fifty gabbro sheets are intercalated with shales of the Attikamagen II. They generally have concordant relationships, and were therefore regarded by Dimroth (1969) as thick basalt flows, whereas Fahrig (1955, 1956a, 1956b), Roscoe (1957) and Baragar (1967) considered them to be intrusive. Discordant relations of a few sheets at the base of the sequence have since been recognized, and these must be intrusive. Lateral gradations of gabbroic rocks into basalts, on the other hand, appear to occur at a few localities (Dimroth, unpublished). These relations will be discussed in a separate paper.

The Denault Dolomite has locally gradational and/or interfingering contacts with the upper part of the Attikamagen III towards the basin margin, and with the lower part of a similar argillaceous unit towards the basin centre. The latter is called the Attikamagen IV Formation in this paper but is to be formally named the Dolly Formation by Harrison *et al.* (in press). It is indistinguishable from the Attikamagen III, and appears to be vertically continuous into the Attikamagen III in the southeast of the Menihek Lake map-area (Frarey, 1961), and in parts of Michikamau Lake area (Wynne-Edwards, 1960), where only lenticular bodies of the Denault Dolomite have been mapped. The Fleming overlies the Denault and apparently interfingers with the Attikamagen IV to the east. The Wishart Formation disconformably overlies the Attikamagen III and all older units at the basin margin, whereas it appears to be conformably on top of the Fleming and Attikamagen IV Formations in the basin centre.

Formations of the Second Cycle

The second cycle comprises the Ferriman Subgroup, and all overlying stratigraphic units, with the possible exception of the Thompson Lake and Willbob Formations.

Table II


PRECAMBRIAN STRATIGRAPHY OF THE CENTRAL LABRADOR TROUGH
FORELAND
WESTERN BASIN
CENTRE
EASTERN BASIN
HINTERLAND
SUBGROUP
GROUP
SUPERGROUP
ERA
Willbob Basalt
Thompson Lake Slate
Murdoch Pyroclastics
basalt and pyroclastics
Menihek Slate
Sokoman Iron Formation
Ruth Slate
Wishart Quartzite
Laporte Schist
Fleming Breccia
Attikamagen IV Slate
Denault Dolomite
Attikamagen III Slate
basalt – shale sequence (Attikamagen II)
Otelnuk F. (slate and greywacke)
Attikamagen I Slate
Savigny Slate
Hautes Chutes Slate
du Chambon F. (slate and conglomerate)
Romanet F. (conglomerate and slate)
Uvé Dolomite
Alder Formation (Sandstone and dolomite)
Lace Lake Shale
du Portage Sandstone
Dunphy Dolomite
Milamar Arkose
Chakonipau Arkose
GREAT UNCONFORMITY
DOUBLET
FERRIMAN
SWAMPY BAY
PISTOLET
SEWARD
KNOB LAKE
KANIAPISKAU
EARLY PROTEROZOIC (APHEBIAN)
ASHUANIPI COMPLEX
(SUPERIOR PROVINCE)
WHEELER COMPLEX
(CHURCHILL PROVINCE)
ARCHAEAN
B-916-1

The Ruth and lower Sokoman Formations interfinger as shown on Table III . The Sokoman is sharply overlain by shales, and greywackes of the Menihek Formation. Contacts between the Sokoman and the Menihek appear to be generally conformable (Harrison, 1952; Dufresne, 1952); but slightly discordant relations are indicated locally (Sauvé, 1953; Dufresne, 1952; Fahrig, 1955).

Table III: Stratigraphy in the central Labrador Trough of the Ruth and Sokoman Formations

	Zone I	Zone II	Zone III		
SOKOMAN FORMATION	200' upper silicate-carbonate iron formation and/or lean cherts	80'-150' upper silicate-carbonate iron formation and lean cherts	50'-100' upper silicate-carbonate iron formation and lean cherts	Upper ferrous zone	SOKOMAN FORMATION
	200' upper hematite iron formation	300'-500' hematite iron formation	200'-300' hematite iron formation	Ferric zone	
	150' lower silicate-carbonate iron formation (local)				
	150' lower hematite iron formation				
RUTH FORMATION	35' iron siltstone	100'-150' lower silicate carbonate iron formation	150' lower silicate-carbonate iron formation	Lower ferrous zone	
	20'-50' iron shale	10'-30' iron shale			
	5'-20' black chert or jaspilite	5'-20' black chert ?	15'-20' black chert or jaspilite	Chert	
	0'-30' shale or siltstone	black shale ?	0'-15' black shale	RUTH	
	Wishart Sandstone	Wishart Sandstone	Wishart Sandstone (partly ferriferous)		

Zone I West of Castignon, Otelnuk and Ritchie Lakes.
Zone II Main ore zone between Eclipse Lake and Attikamagen Lake (western shore).
Zone III Eastern zone between Low Lake and Attikamagen Lake (eastern shore). Upper members of Sokoman absent NE of Low Lake.

The Wishart, Ruth and Sokoman Formations are absent in the extreme east of the Trough south of 56°N, and in the centre between 56°N and 57°N, as shown on Figure 7.

On the west side of the Trough between 55°30'N and 56°N dolomite overlies the Ferriman Subgroup and is overlain in turn, by shales. The dolomite was called the Purdy Formation and was thought to succeed the Sokoman Formation unconformably (Private Reports, Iron Ore Company; Frarey and Duffell, 1964; Baragar, 1967). Nearly continuous mapping, lithological comparisons, and structural considerations suggest that the dolomite is part of a horizontal thrust sheet involving the Alder and Uvé Formations, and that the overlying shales correlation in part with the Attikamagen I, and in part with the Attikamagen III. The term Purdy Formation should therefore be dropped.

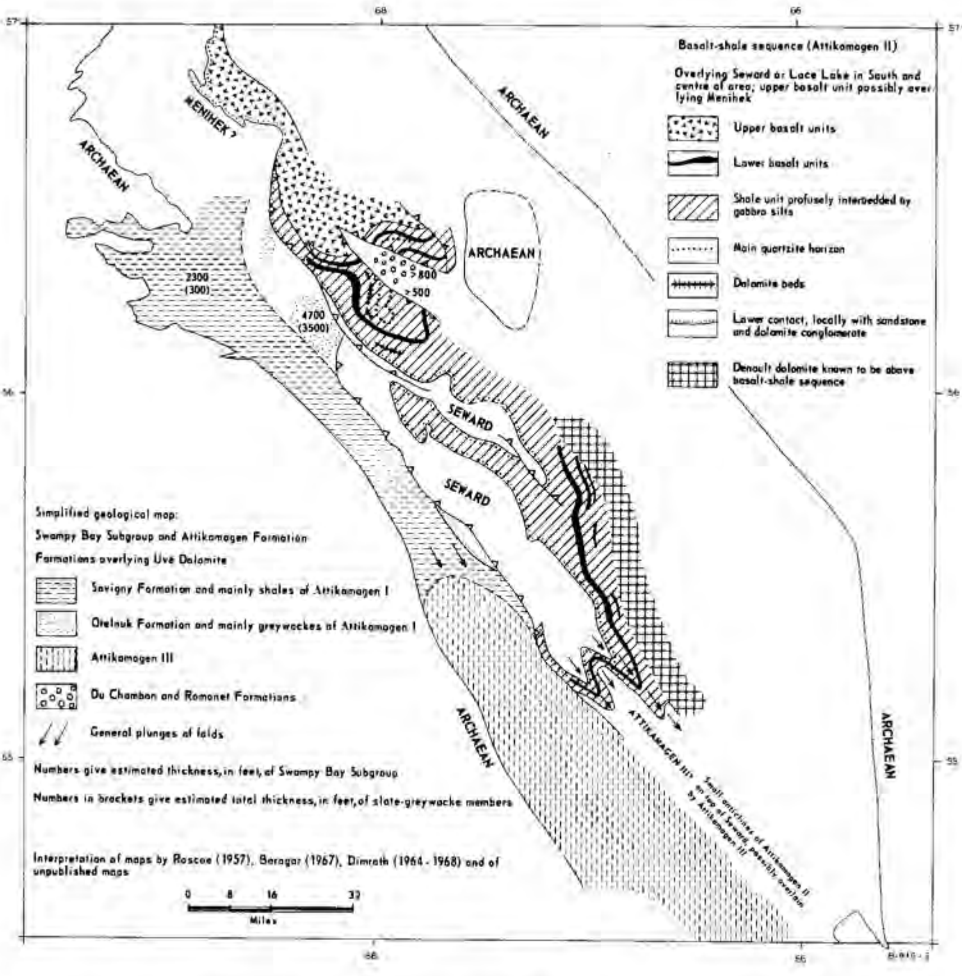


Figure 2. Generalized distribution of the Attikamagen subunits.

A thick (10,000 feet ±) volcanic unit that overlies the Attikamagen II Formation with apparent conformity between latitudes 56°N and 56°30'N (Fig. 2) presents some uncertainties in its correlation. At Wapaniskan Lake (about lat. 57°N) the northward continuation of the same unit overlies post-Sokoman slate (Menihek?) and older rocks with apparent discordance (cf. Fahrig, 1955, 1956a; Roscoe, 1957; Hashimoto, 1964). It comprises pillowed basalts overlain by mafic pyroclastic rocks containing minor acidic components. These features suggest to us that the unit is the correlative of the Menihek basalts of Wakuach Lake area (lat. 55°N-56°N) which are similarly overlain by a pyroclastic sequence containing minor acidic pods (Baragar, 1967). However, the absence of the Denault, Wishart, and Sokoman Formations between this volcanic unit and the concordantly underlying Attikamagen II Formation is puzzling. Possibly they were not deposited in this region.

Formations of the Doublet Group overlie the Menihek Formation in the east of the Trough. These relations are fairly well established (Frarey, 1967; Baragar, 1967). Biotite schists and garnet, staurolite and kyanite bearing schists named Laporte Schist (Fahrig, 1951) overlie the Archean gneisses in the extreme east of the Trough. The relations in the Fort Chimo region (Gélinas, 1965), and to the northeast of Effiat Lake (Dimroth, unpublished) suggest that the Laporte Schist corresponds to part of the sequence below the Murdoch Formation, and above the Pistolet Subgroup. It may correspond to parts of the Attikamagen and Menihek Formations.

Facies and Paleogeography

Introduction

Facies and paleogeography of the part of the sequence (below the Wishart Formation) have been described in some detail by Dimroth (1968b), and only features that appear to have major implications will be discussed. A considerable amount of work has been done during the last two years and its results are also included here.

Rocks of the First Cycle

Introduction: The cycle began with the rapid deposition in a continental basin of red arkoses and conglomerates (Chakonipau Formation). The facies and distribution of these deposits were strongly influenced by an active, easterly trending fault graben. Subsequently the intensity of faulting decreased, but faults continued to exert some influence during the whole of the cycle. A sequence of predominantly sandstone and dolomite (upper Seward, Pistolet Subgroup) followed. These rocks are mainly marine. Continental influence which predominated at the beginning of the cycle, declined rapidly during this period. Marginal facies may be recognized in most of the units deposited at this time on both west and east sides of the Trough. After deposition of euxinic black shale (Hautes Chutes Formation) followed sedimentation of shales, quartz wackes and grits of the main part of the Swampy Bay Subgroup and of the Attikamagen I in separated basins in the west and east of the Trough. Much of the sedimentary material deposited at this stage is derived from a geanticlinal area in the centre of the Trough. Renewed faulting occurred, and block conglomerates were deposited locally at the base of fault scarps (Romanet Formation). Gradually the central geanticline subsided again and was buried beneath a younger shale-quartz wacke sequence and basalts (Attikamagen II). Stabilization of the Trough at the end of the first cycle resulted in widespread carbonate deposition (Denault Formation).

Chakonipau Formation: The Chakonipau Formation shows all the features of a unit deposited rapidly in a semi-dry continental basin. Figure 3 shows the facies distribution of the Chakonipau Formation. The basin apparently had a limited extent, and was restricted to a zone in the centre of the geosyncline with a westward trending appendix north of 56°. The sedimentary material appears to be derived from three different source areas. Quartz pebble conglomerates characteristic of the formation west of 69° and north of 56° are likely derived from coarse-grained granites extending of the northwest (private reports, Dennison Mines Ltd.). North of 56°00' and east of 69° most of the formation is derived from a terrain underlain by medium-grained biotite-hornblende-plagioclase gneisses and by andesites. Rapid coarsening of conglomerates suggests a source area farther south. A fault, named Cambrien-Otelnuk fault has been inferred to be the southern boundary in this part of the basin (Dimroth, 1968b). Baragar (1967) has inferred that granitic rocks to the west are the source of the material deposited south of 56°N. This material was then redistributed by longitudinal currents in the centre and east of the basin.

Explosive andesitic and basaltic activity occurred during the Chakonipau deposition. Abundant andesitic fragments are common in the arkoses and conglomerates at Chakonipau and Otelnuk Lakes (lat. 56°15', long. 68°20')

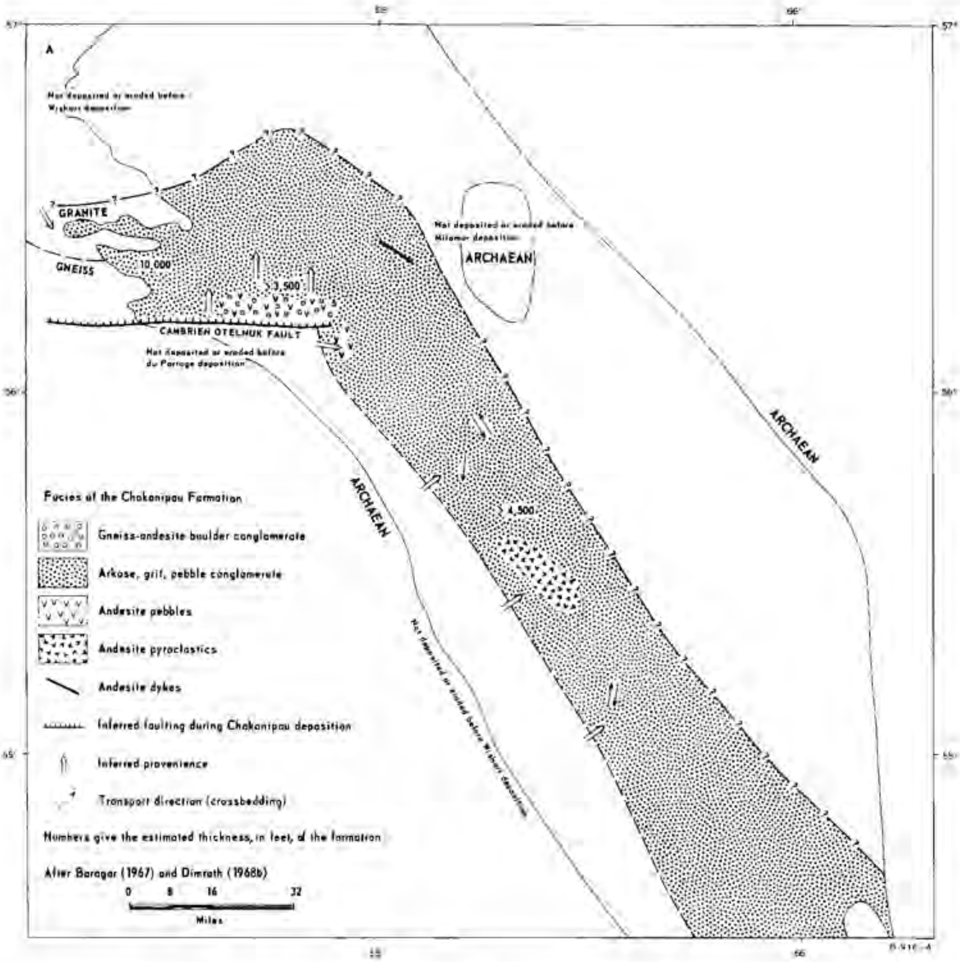


Figure 3. Facies of the Chakonipau Formation.

close to the inferred Cambrien-Otelnuk fault. Baragar (1967) has described basaltic agglomerates, breccias, tuffs, and flows southeast of lac Musset (lat. 55°30', long. 67°15'). Andesitic and quartz-dioritic dykes were observed at Romanet River (lat. 56°25', long. 67°15'). These are somewhat younger, because they intrude the Dunphy Formation.

Upper Seward and Pistolet Subgroups: The facies of the upper part of the Seward and of the Pistolet indicates on the whole a waning continental influence. Subsidence of the basin was not a continuous process, but rather appears to be the result of two interfering processes:

1. Subsidence of the basin and slow uplift of relatively distant source areas east and west of the geosyncline. Chemical weathering in the source area. Coarse clastics remained on the immediate forelands and hinterlands whereas pelitic material was deposited in the marine basin.
2. Continuing subsidence of the basin and simultaneous emergence of geanticlines in the immediate foreland and particularly in the hinterland of the geosyncline. Sands were transported from the uplifted zones into the

basin centre and marginal overlaps formed, especially in the east of the Trough. Subsidence rates later appear to decrease; dolomites were deposited in the basin centre and extended towards the basin margins during the waning stage of this phase.

This rhythmical interplay of uplift and subsidence occurred three times with increasing intensity of the subsiding phase. The basin centre migrated westwards during this part of the evolution of the Labrador Trough.

The facies distribution of the upper Seward Subgroup is shown on Figure 4. Continental influence still prevailed during deposition of this unit. It has been argued (Dimroth, 1968b) that the fine grain size and good sorting of the very fine grained arkosic sandstones characteristic of the du Portage Formation indicate wind transport. Beds of very fine grained arkosic sandstone alternate with beds of coarse grained quartz sandstone that show textural and compositional maturity. The quartz sandstone grades laterally into dolomitic sandstone, calcarenite, and stromatolitic dolomite that are most likely beach or shallow marine deposits. Silt and hematite dust occurring in the marine, stromatolitic Dunphy Dolomite, support the inferred participation of wind transport during deposition. The Dunphy Dolomite grades eastwards into arkoses, conglomerates, and quartzites of the Milamar Formation.

Thicknesses of the formations of the upper Seward reach a maximum of 2,000 feet in the east-central part of the Labrador Trough.

The Lace Lake Shale is a laminated shale with beds and lenses of dololutites. Its grey colour, tabular laminations, and graphite content suggest marine deposition below wave base. In the extreme west of the Trough a marginal facies of the formation is characterized by coarser grain-size (siltstones, argillites, some very fine grained sandstones), by red and green colours, by flaser textures of the siltstones, by cross-bedded dolosiltites, and by intraclastic dolomites. In the extreme east of the Trough the formation contains a higher proportion of very fine grained sandstone. The formation is of remarkably constant thickness over the whole width of the Trough.

Five facies have been recognized in the Alder Formation (Fig. 5). They are:

1. A marginal facies, mainly composed of sandstone, in the west and east of the Trough.
2. A northwestern marginal facies of somewhat uncertain stratigraphic position occurs in a limited area southwest of Hematite Lake (56°30'N; 68°50'W). It is similar to the marginal facies of the Lace Lake Formation but contains numerous sandstone beds, five to twenty feet thick.
3. A submarginal facies of stromatolitic dolomite or calcarenite extends in a narrow zone in the west.
4. A dolomitic basin facies characterized by a sequence of alternating beds of stromatolitic dolomite and sandstone.
5. A sandy basin facies represented by thick sandstones. This facies contains much grit and some conglomerate northeast of Wakuach Lake (lat. 55°45', long. 67°45').

The marginal facies is mainly composed of coarse-grained sandstone in the west of the Trough; local interbeds of grits containing shale granules, of sandstone conglomerate and of gneiss pebble conglomerate occur in the east of the Trough. An overlap of the Alder over older rocks in the east of the basin is therefore inferred. The thickness of the Alder Formation increases rapidly towards a maximum in the west-centre of the Labrador Trough.

The lower member of the Uvé Formation is similar to the Lace Lake shale. The boundary between the marginal and the basin facies in this member

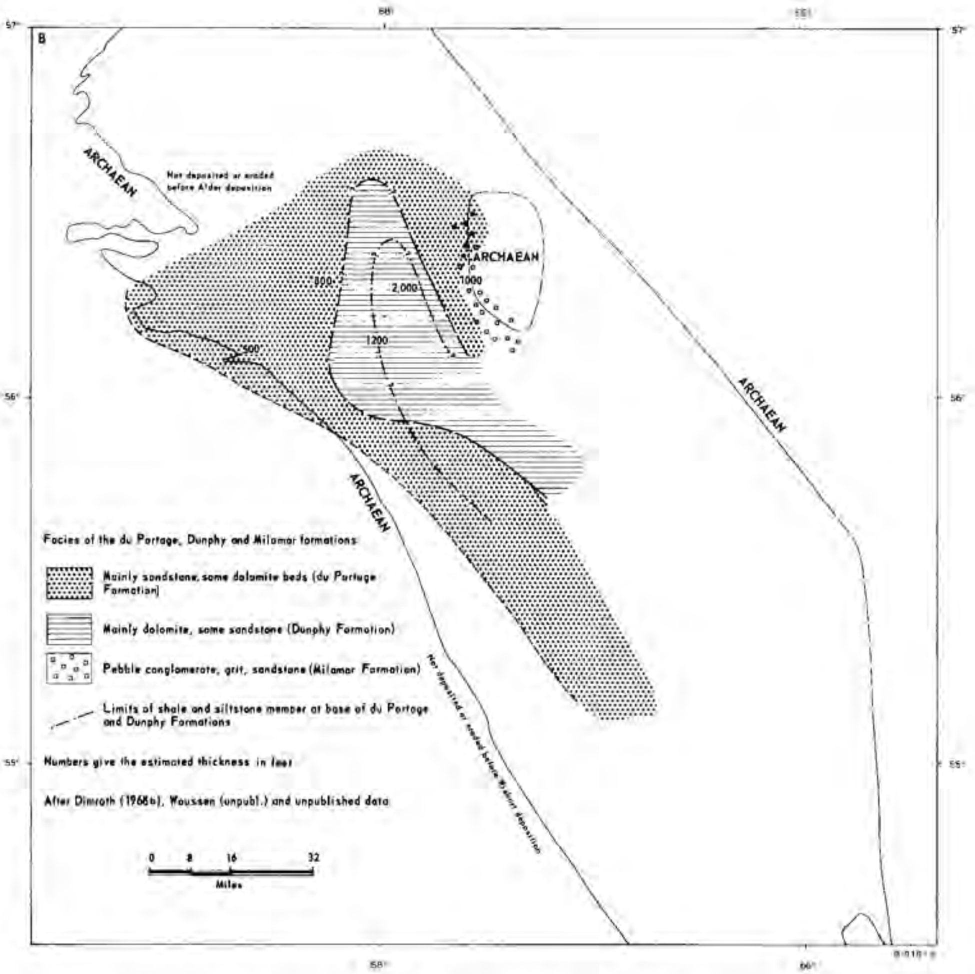


Figure 4. Facies of the upper part of the Seward Subgroup.

is also shown on Figure 5. The upper member, formed mainly of beds of massive dololutite alternating with some beds of sandstone and shale, shows little facies differentiation. It commonly shows traces of synsedimentary deformation.

Swampy Bay and Attikamagen Subgroups: Units of the Attikamagen Subgroup fall roughly into older (Attikamagen I-Swampy Bay Subgroup) and younger (Attikamagen III, Attikamagen II) divisions (Table II). The relationship between these subunits has already been discussed (pp. 9-12). Facies maps of both age units are presented as Figures 6 and 7. Although time relations between contrasting lithological units deposited in different parts of the basin are not known with precision, a pattern is apparent that leads to the following hypothetical sequence of events:

1. Deposition of the Hautes Chutes Shale in an euxinic basin.
2. Emergence of geanticlinal source areas in the centre and in the east of the basin. Tectonic activity and faulting. Deposition of shales and

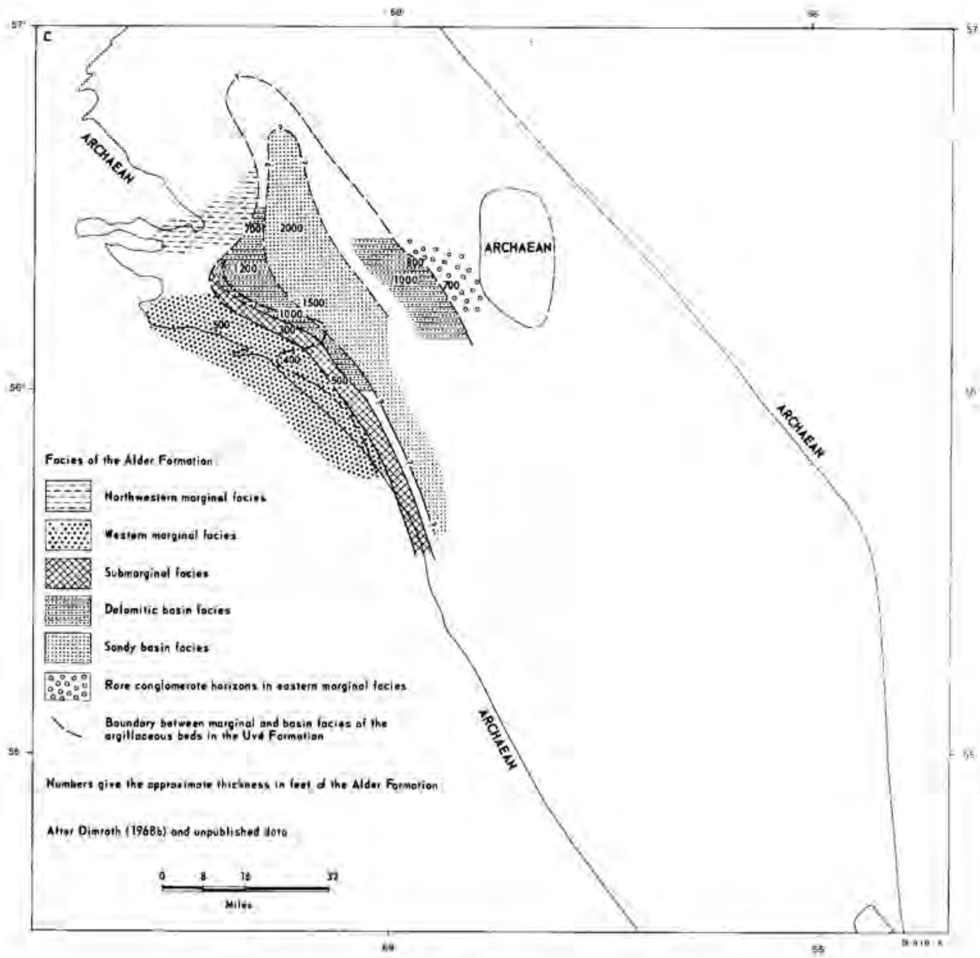


Figure 5. Facies of the Alder Formation.

greywackes, with some turbidite structures (graded bedding, Bouma cycles, flute casts, convolute laminations) in a western basin in the Labrador Trough, and of conglomerates, grits, sandstones, quartz wackes and shales in an eastern basin. Some block conglomerates are probably slumped masses deposited at the base of fault scarps. This phase is discussed in detail in Dimroth (1968b). A facies map of the lower subunits of the Attikamagen that overlie the Uvé Dolomite conformably is presented as Figure 6.

3. Renewed subsidence of the whole geosyncline. Basaltic volcanic activity in the east with deposition of a basalt-shale sequence (Attikamagen II) in the east, and of silty shales (Attikamagen III) in the west.

Volcanic rocks, and gabbro sills are thickest in the eastern basin of the Trough. They lens out gradually to the west, but decrease abruptly in thickness at the eastern margin of the Trough. A relatively thin suite of metamorphosed silty shale and impure sandstone, with interbeds of sandstone, dolomitic sandstone, silty marl, and rarely dolomite occurring on the hinterland at

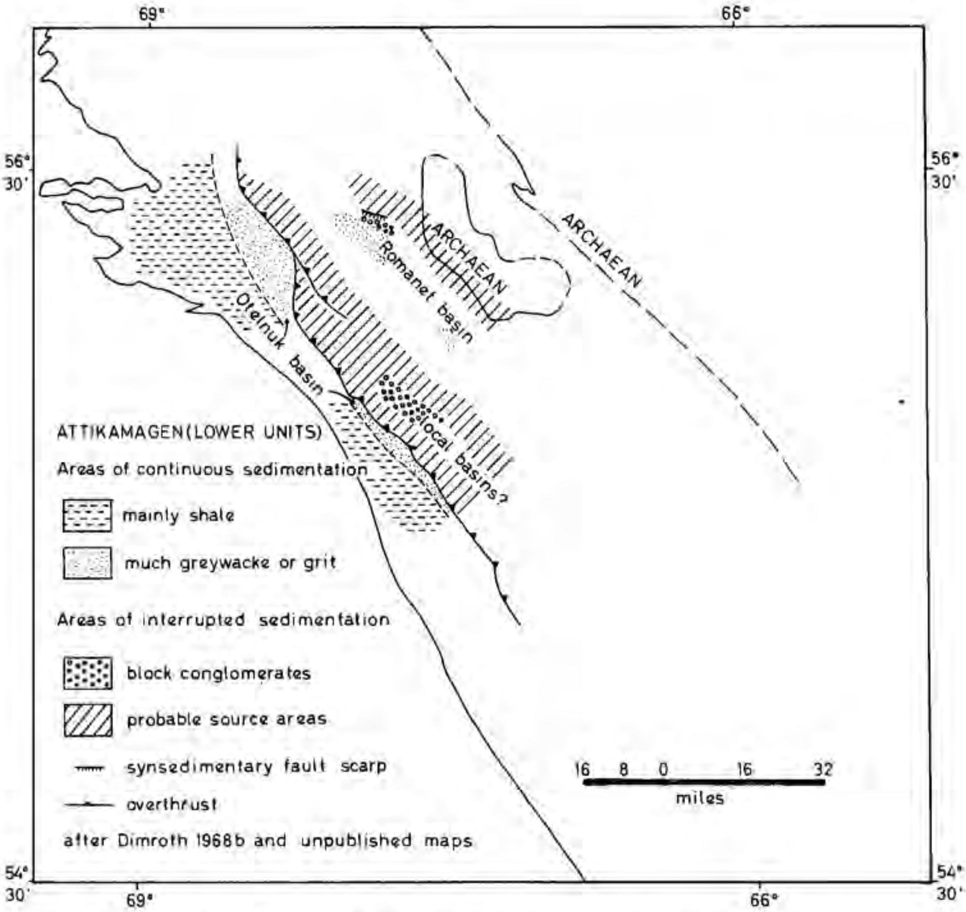


Figure 6. Facies of the lower subunits of the Attikamagen.

about 56°N (Laporte Schist), is probably equivalent to the Attikamagen and Menihek Formations. If so the Attikamagen Subgroup on the hinterland at this locality is thin, is nearly devoid of volcanic rocks, and comprises a higher proportion of mature sandstones and calcareous rocks than it does in the Trough to the west. Figure 7 presents a facies map of the upper division of the Attikamagen.

The Denault Formation and its relation to other units: The Denault Dolomite occurs only in the southern part of the Central Labrador Trough. Its distribution is shown on Figure 8. The formation interfingers with the underlying Attikamagen III and the overlying Attikamagen IV (Dolly) Formations (pp. 12-13) and appears to be partly equivalent to both (Figs. 8 and 11). Dolomite shales and dolomite lenses in the upper part of the Attikamagen III are indicative of an increasing carbonate precipitation in the west of the Trough. The Denault Formation is only a few feet thick in the west, where it is a finely laminated slightly shaly dolomite. It thickens very rapidly basinwards. Conglomeratic dolomites occur on the slope of the basin. These show many characteristics of allodapic limestones as defined by Meischner (1964); shale interbeds increase

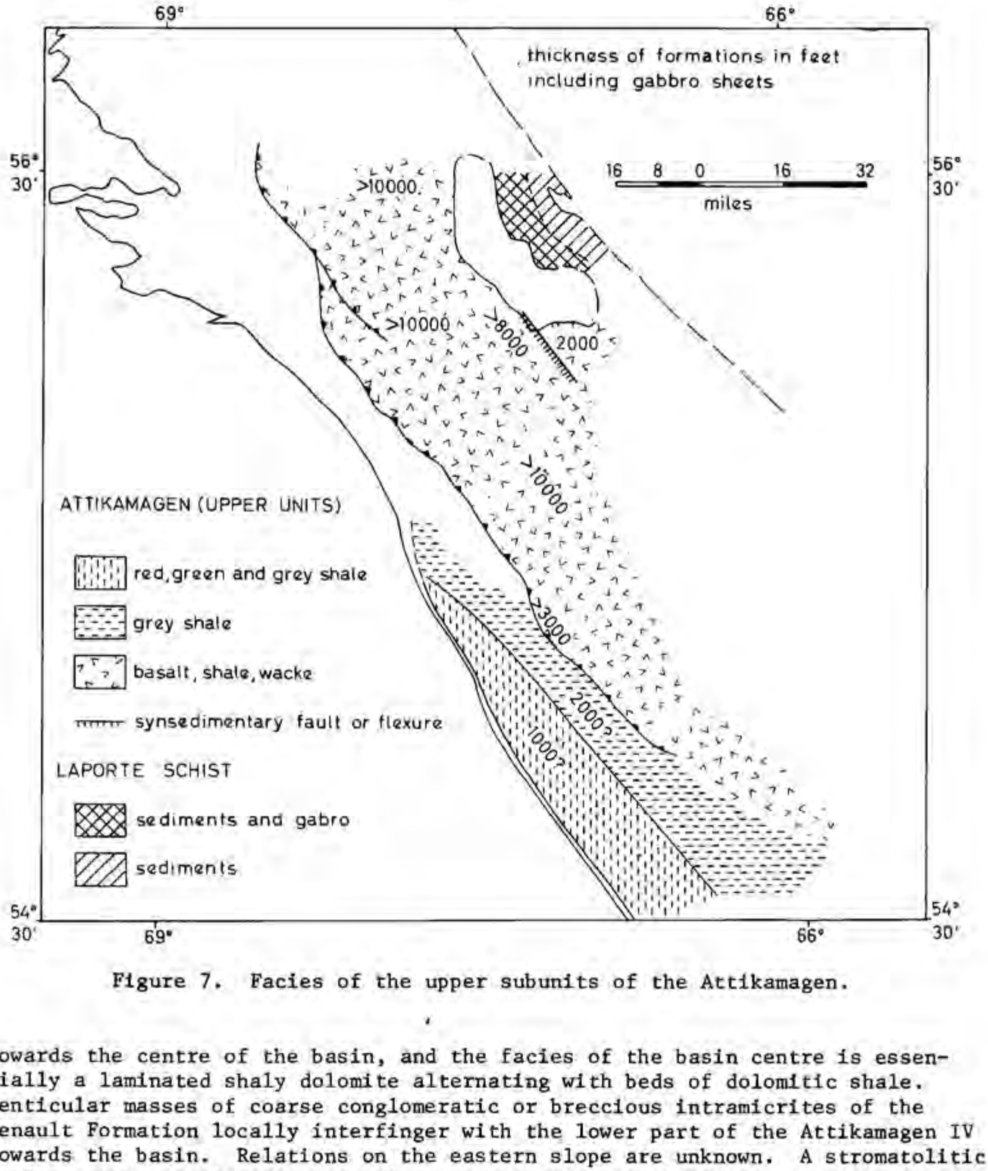


Figure 7. Facies of the upper subunits of the Attikamagen.

towards the centre of the basin, and the facies of the basin centre is essentially a laminated shaly dolomite alternating with beds of dolomitic shale. Lenticular masses of coarse conglomeratic or breccious intramicrites of the Denault Formation locally interfinger with the lower part of the Attikamagen IV towards the basin. Relations on the eastern slope are unknown. A stromatolitic facies, with interlayered beds of laminated dolomite showing features of strong synsedimentary deformation (Donaldson, 1963; Baragar, 1967) occupies a platform in the east of the Trough, and dolomitic sandstone is present in the extreme northeast of this facies zone north of 56°N. The Denault Formation apparently lenses out rapidly at the eastern margin of the Trough.

Attikamagen IV and Fleming Formations: The Attikamagen IV (Dolly Lake) Formation occurs only in the extreme south of the area considered in this part of the paper. Its lateral extent and approximate thickness are shown on Figure 8. The formation is composed of black, grey, or red and green laminated shale with local beds of fine-grained sandstone. Tongues of breccious intramicritic dolomites that are apparently continuous with the Denault Dolomite

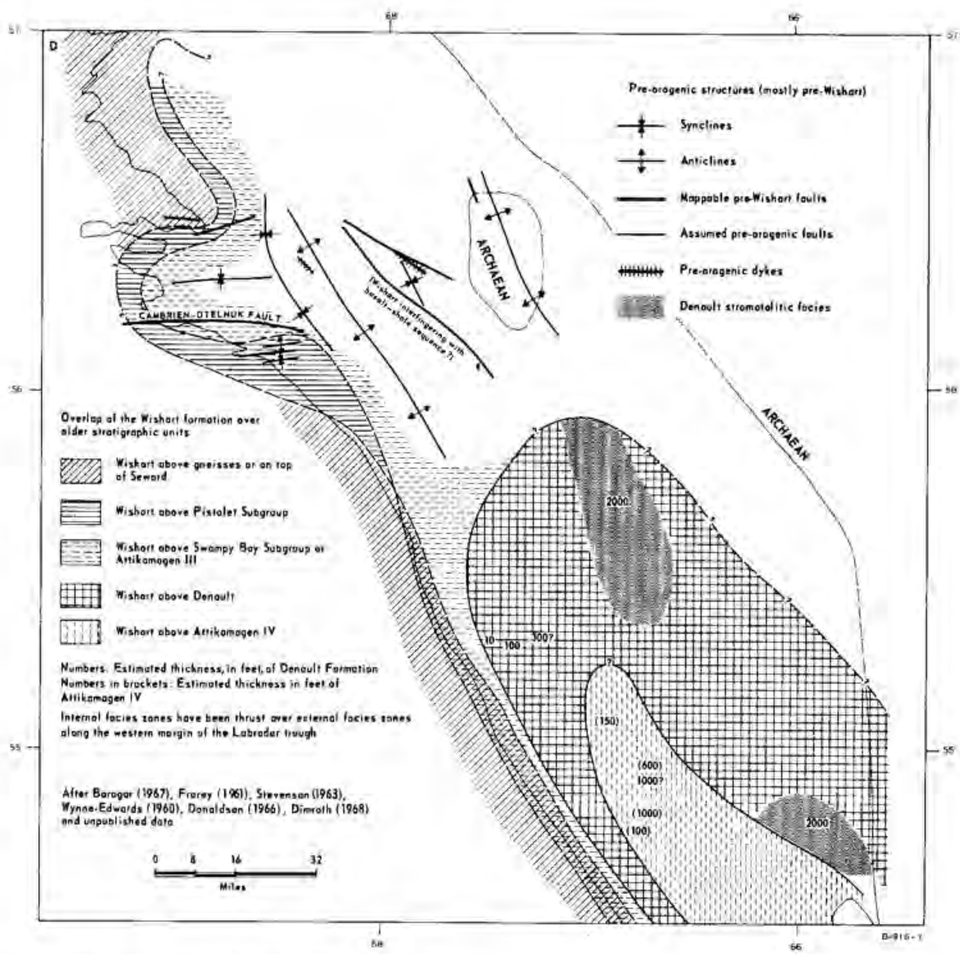


Figure 8. Overlap of the Wishart Formation over older stratigraphic units.

interfinger into shales of the lower part of the formation in the railway cut 1 mile south of Schefferville. It is therefore believed that the formation is in part a time-equivalent of the Denault.

The Fleming Formation overlies the Denault in a narrow zone (less than 10 miles wide) in the west of the Trough. The western limit of its distribution coincides closely with the western limit of the distribution of the Denault (shown on Figure 8). The formation thickens from 0 to a maximum 300 feet towards the east from where it pinches to nothing within a mile. The Fleming is likely a slumped mass derived from an originally bedded sequence consisting of alternating shale, laminated chert, sandstone, and of chert breccias, that was deposited in the hinge zone between the rising margin of the geosyncline farther west and the basin to the east. The formation is now an unstratified body of irregularly interpenetrating masses of widely contorted and brecciated laminated chert, of chert breccias cemented by chert, by cherty sandstone, and by shale, of sandstone with chert fragments and of shale.

Rocks of the Second Cycle

Introduction: Rocks of the second cycle overlie the older units unconformably at the margin of the Trough, whereas deposition was apparently continuous in the basin centre south of 55°15'. They commenced with a thin blanket of quartzites, shales, and iron formation (Ferriman Subgroup) showing minor facies differentiation and were followed by a thick succession of shales, greywackes, and basalts (Menihek Formation). Little is known of the facies differentiation of the latter, but scattered evidence suggests that deposition of the formation was accompanied by tectonic movements. The cycle ended with widespread pyroclastic eruption in the east of the Trough.

Relations between the Wishart Formation and rocks of cycle I: Partial emergence and an unconformity below the Wishart was first inferred by Bergeron (1954) and by Fahrig (1957). This unconformity is now well established in the marginal zone of the Trough, and in the area north of 56°N. Figure 8 shows the overlap of the Wishart Formation on the older formations. The relations suggest a gradual retreat of the sea to the south before deposition of the Wishart Formation. Faults and some very open folds formed before the Wishart Formation was deposited in the area north of 56°. No orogeny took place at that time, and the very open folds are probably draped over normal faults in the basement.

The relations in the basin centre are not entirely clear. A conformable contact likely exists between the Attikamagen IV or at least its higher portions and the Wishart. The Fleming Formation underlies the Wishart in a narrow zone in the west of the Trough. This information is likely developed in the hinge zone between the basin margin, where the Wishart overlies older rocks disconformably, and the basin centre, where sedimentation was more or less continuous.

Ferriman Subgroup: The Wishart, Ruth, and Sokoman Formations show little facies differentiation. The Wishart is generally composed of orthoquartzites, containing locally shale and chert fragments and beds of arkosic grits. Impure sandstones and subgreywackes occur in the centre and in the east of the Labrador Trough, and ferriferous sandstone composes part of the formation in the east. Parts of the Wishart are characterized by chlorite granules (metamorphosed glauconite?). The Wishart was likely deposited in a shallow marine environment.

The Ruth Slate and the lower part of the Sokoman Formation interfinger as shown on Table III and Figure 9. The relations suggest that the Ruth siltstone in the zone between Hematite and Purdy Lakes (between lat. 56°30' and 55°30') is derived from a source area to the west. Very little clastic material has been deposited as Ruth Slate south of 55°30'; it appears to be largely derived from tuffaceous material that had its origin in volcanic centres at Astray Lake (Zajec, personal communication). The relations are further complicated by a near-shore ridge at the western margin of the Trough west and northwest of Schefferville (lat. 55°00', long. 67°15') (Zajac, personal communication). Little or no clastic material has been deposited on top of the ridge, where the silicate-carbonate ironstone, and locally hematite ironstone were precipitated simultaneously with the deposition of the Ruth Slate farther north, west, and east. These complications are not shown on Figure 9.

Cross-lamination is common in the Ruth siltstones of Zone I (Fig. 9), whereas delicate tabular laminations characterize the Ruth shale and the equivalent portions of the Sokoman in zones 2 and 3. The graphite and pyrite content of the Ruth increases from zone 1 to zones 2 and 3. This appears to indicate that water depth generally increased towards the east.

The hematite ironstone (member 2) of the upper member of the Sokoman contains a high proportion of oolitic and intraclastic material (Dimroth, 1968a), and was deposited in very shallow water. A lenticular member of silicate-carbonate ironstone occurs in member 2 between Goethite and Lace Lakes

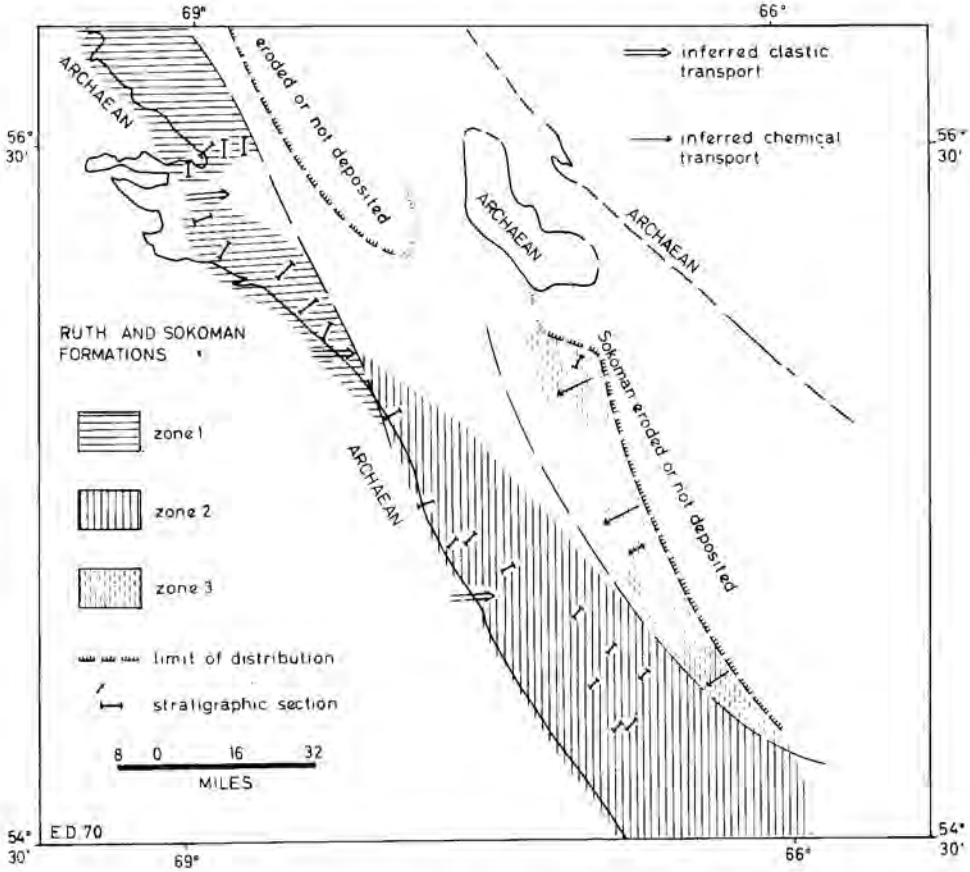


Figure 9. Facies of the Ruth and Sokoman Formations.

(lat. $56^{\circ}15'$, long. $68^{\circ}30'$). It could possibly indicate the presence of a local basin, although common intraclastic textures prove that it has been deposited above wave base.

It has previously been assumed (Kirkland, 1950; Dufresne, 1952) that the iron and silica were derived by chemical weathering from a landmass either west or east of the geosyncline. Baragar (1967) first expressed some doubts that sufficient iron could be obtained by weathering of the essentially granitic source material. Little argillaceous material was transported into the basin during Sokoman deposition. This is difficult to explain if deep chemical weathering took place at the source. Therefore the iron and silica are likely derived from a volcanic source. The very great volume of ophiolitic volcanics that show a differentiation trend towards iron enrichment should be considered a possible source of the ironstones. Some of the precipitated material could also be derived from the miogeosynclinal volcanics of the Astray Lake area.

Volcanic activity during Ferriman deposition: Volcanic rocks, mainly pyroclastic, but comprising some massive flows, interfinger into the Wishart, Ruth and Sokoman Formations in the west of the Trough south of 56°. The volcanic centres were probably located at Astray Lake ($54^{\circ}30'$N; $66^{\circ}15'$W), and may have formed subaerially exposed volcanoes (Sauvé, 1953). The volcanic rocks are associated with sandstones and chemical sediments deposited under relatively stable conditions. They are not very thick and comprise only a comparatively small part of the sedimentary thickness even close to the volcanic centres. The volcanic rocks are orthoclase bearing (Sauvé, 1953) and highly potassic (Zajac, personal communication). Their geology, petrography and chemistry is therefore in complete contrast to the thick submarine basalts of the Attikamagen, Menihek, or Willbob Lake Formations.

Menihek Formation and higher units: The facies differentiation of the Menihek Formation is not well known. A thick sequence of Menihek slate and greywacke occurs along the west side of the Trough (Fig. 10) and similarly along the east side where it interfingers upward and northward into massive and pillowed basalts. These latter are overlain by pyroclastic rocks of the Murdoch Formation (55°N-56°N). Farther north (57°N) a thin shale unit that overlies the Sokoman Formation is succeeded disconformably by a thick assemblage of lavas and, in turn, pyroclastic rocks. These were interpreted (p. 12) as Menihek shales and lavas overlain by the equivalent of Murdoch pyroclastic rocks. They are continuous with the thick volcanic assemblage that overlies the Attikamagen II (basalt shale sequence) Formation at 56°30'N.

Some tectonic movements may have taken place during deposition of the Menihek Formation. The Menihek is conformable with the underlying Sokoman Formation in the west and east of the Trough between latitudes 55°N-56°N but Menihek lavas seem to rest directly on Attikamagen II Formation at 56°30'N as noted above. The same lavas appear to overlie Menihek shales and older rocks with discordance farther north.

North of 57°N, the Chioak Formation (Bergeron, 1954; Bérard, 1966) which corresponds to the Menihek in its relation to the ironstones, contains much conglomerate and is unconformable with units below. It was deposited during a period of faulting (Bérard, 1966). Intraformational conglomerates occurring in the Menihek at Petitsikapau Lake (54°30'N, 66°30'W; Iron Ore Company of Canada, 1949) suggests movement during deposition of the formation. Thus the character of the Menihek Formation may be analoguous to that of the Attikamagen Subgroup.

The formations of the Doublet Group are limited to the eastern side of the Labrador Trough; nothing is known therefore, on the facies distribution of these units in the Central Labrador Trough.

The Castignon Lake Graben

Fahrig (1957) pointed out that the area between 56°00' and 56°30' occupies a unique position in the Labrador Trough. It was a zone of strong subsidence during Chakonipau deposition. Very likely an east trending fault graben (named Castignon Lake graben) developed during that time. East trending facies boundaries persist into the upper Seward and Pistolet rocks, and are particularly evident in the facies of the lower member of the Uvé Formation (*see* Fig. 5). The importance of this graben zone was waning with time, and no evidence of east-trending facies boundaries is apparent in the Wishart or younger formations.

Several faults and folds of pre-Wishart age have been mapped in this zone, and other faults are pre-orogenic, although not certainly of pre-Wishart age. Fahrig inferred that the zone was an erosionsl valley trending across the geosyncline but considered that a tectonic origin was possible. The evidence obtained during recent years suggests that it is a tectonic feature. It may be significant to note that this fault basin is situated near the northern limit of the occurrence of the pre-Wishart rocks.

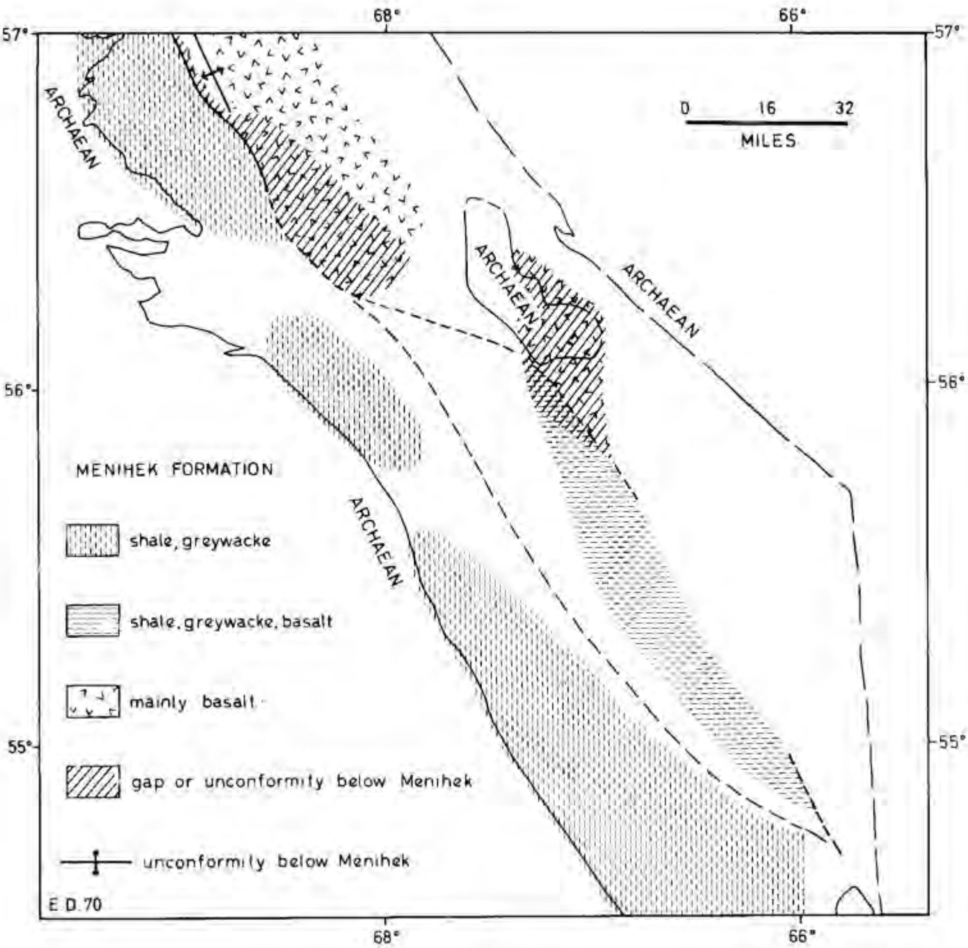


Figure 10. Facies of the Menihek Formation.

Summary and Discussion

Figure 11 schematically represents the stratigraphy after the deposition of the Sokoman Formation, and illustrates that the sedimentary evolution of this part of the Trough in the first cycle (before Wishart deposition) may be subdivided into four stages.

1. Development of a fault basin; deposition of arkoses and conglomerates; andesitic and trachybasaltic volcanic activity.
2. Marine transgression and deposition of an orthoquartzite-dolomite sequence; waning continental influence; migration of the basin centre westward; the sediments deposited at this stage lens out rapidly at the eastern margin of the Trough.
3. After a short interlude of euxinic deposition followed the emersion of a ridge in the centre of the geosyncline. Deposition of flysch-type greywackes and shales in a rapidly subsiding basin in the west, and of conglomerates and shales in a basin (or basins?) in the east; intense faulting.

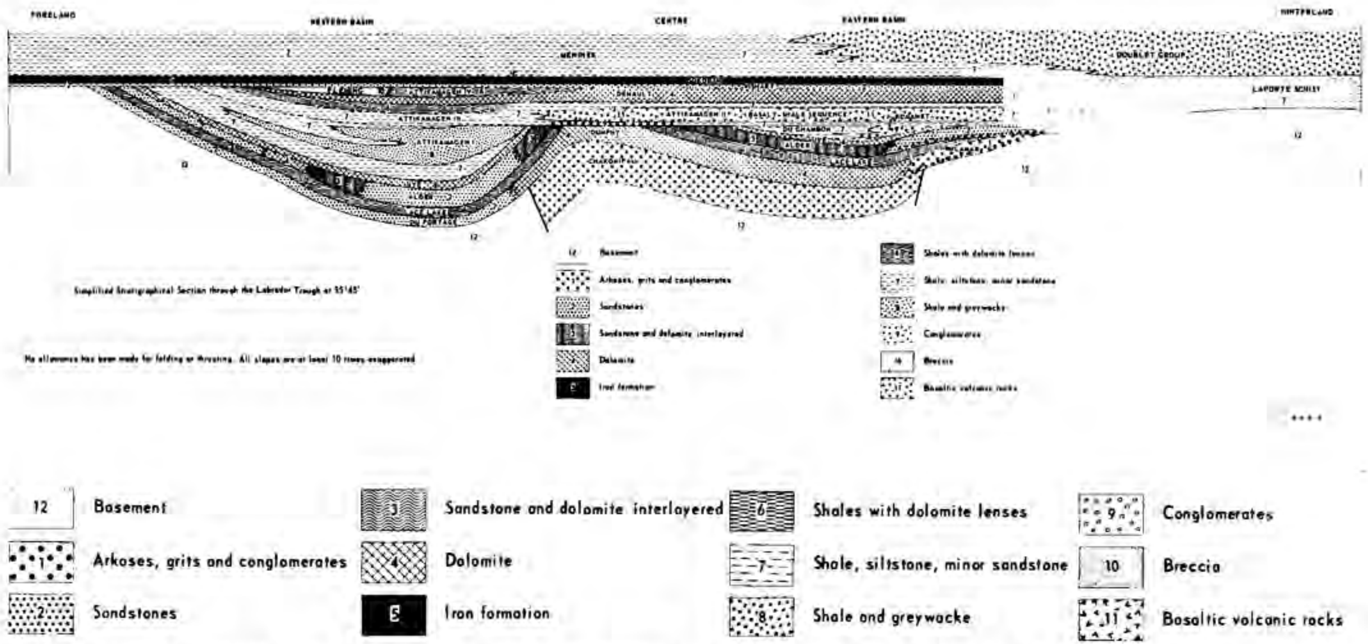
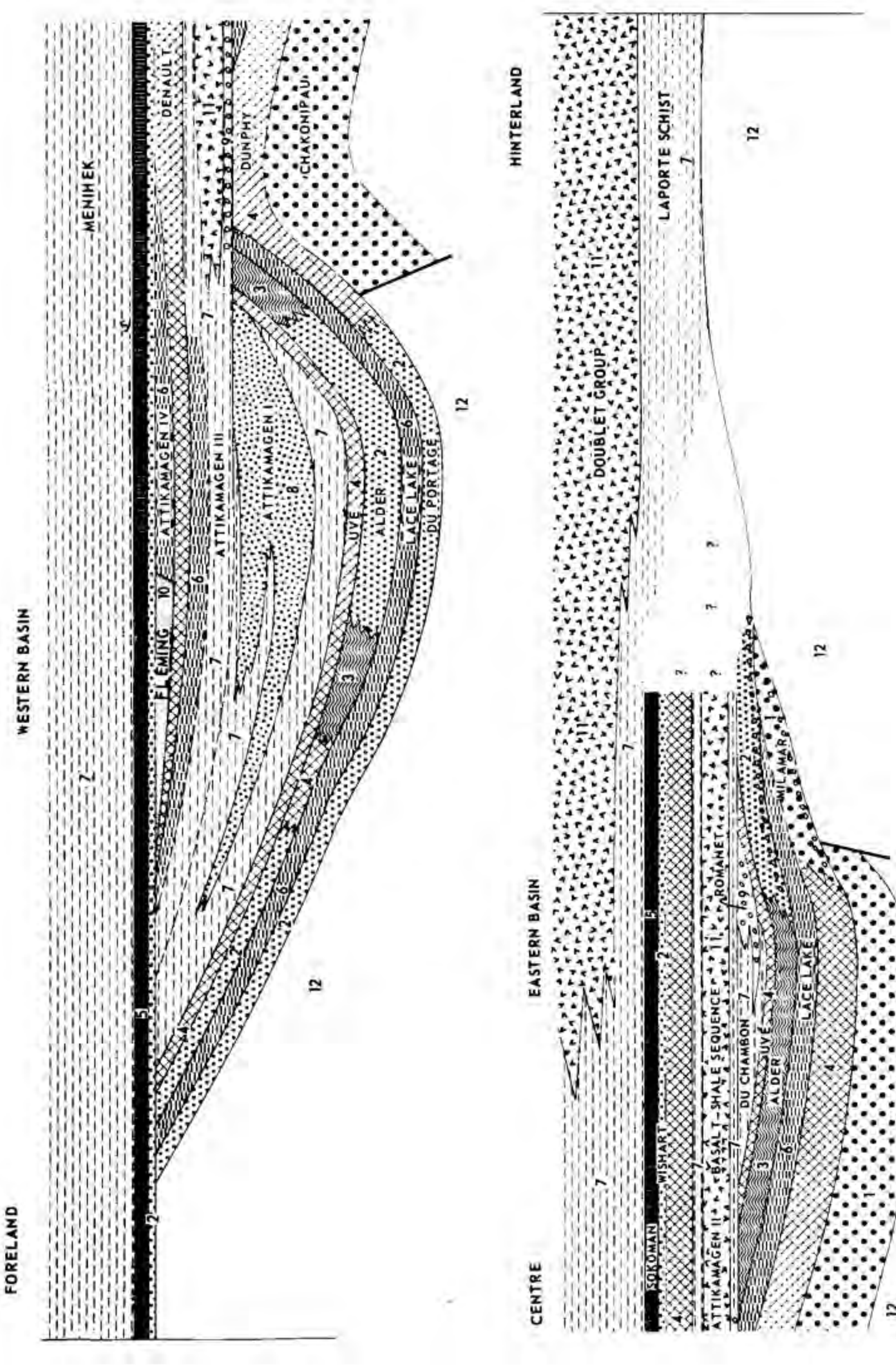


Figure 11. Simplified stratigraphical section through the Labrador Trough at 55°45'. No allowance has been made for folding or thrusting. All slopes are exaggerated at least 10 times.

Figure 11a. Western portion.
11b. Eastern portion.


FORELAND
WESTERN BASIN
MENIHEK
DENAULT
DUNPHY
CHAKONIPAU
FLEMING
ATTIKAMAGEN IV
ATTIKAMAGEN III
ATTIKAMAGEN I
UVÉ
ALDER
LACE LAKE
DU PORTAGE
CENTRE
EASTERN BASIN
HINTERLAND
DOUBLET GROUP
LAPORTE SCHIST
SOKOMAN
WISHART
ATTIKAMAGEN II
BASALT-SHALE SEQUENCE
ROMANET
DU CHAMBON
UVÉ
ALDER
LACE LAKE
MILAMAR

4. Renewed strong subsidence of the whole geosyncline, particularly in the east; continuing deposition of shales and greywackes; mafic volcanic eruptions in the east accompanied by intrusion of gabbro sheets; the subsidence rate later decreased and dolomitic shales and dolomites were deposited.

Emersion and erosion of the basin followed before the quartzites and iron formations of the second cycle were deposited. What little is known about the facies and paleogeography of the formations of cycle II suggests an evolution that on the whole, duplicates steps 2 to 4. A northward migration of the geosyncline is indicated by the fact that the rocks of the first cycle are absent north of 57°N, where the formations of the second and third (?) cycles are exceptionally thick.

The geosyncline is subdivided into two major facies domains: a western zone of sediments, and an eastern zone with a dominantly mafic magmatic filling. Sedimentary rocks (silty shales with interbeds of calcareous rocks and of quartzites) again overlie the hinterland east of the Trough. The thickness of the filling generally increases from the foreland towards the east of the Trough, and apparently decreases rapidly at the eastern boundary of the geosyncline. A ridge formed within the geosyncline during some stages of its evolution; parts of the sequence are therefore absent in the median part of the geosyncline.

THE NORTHERN LABRADOR TROUGH

The Northern Labrador Trough extends roughly from the 57th degree of latitude to the western tip of Ungava Bay, a distance of about 300 miles. At the southern limit, it is approximately 40 miles wide and the width decreases northward to about one mile south of Payne River. From there it increases to about 20 miles, before tapering off north of Robert Lake. A small basin 16 miles long appears a few miles north of the end of the Trough.

For the description of the stratigraphic sequences, Payne River will serve as a boundary between a southern and a northern sector.

Southern sector (south of Payne River)

The stratigraphic succession, outlined in Table I, can be subdivided in two cycles with orthoquartzites and chemical sediments at the base and shales and greywackes on top. The Allison Quartzite and Fenimore Ironstone represent the first orthoquartzite-precipitate phase; they are overlain by the pelitic-psammitic Chioak Formation. A second dolomite phase (Abner Dolomite) follows and is again overlain by the argillaceous Larch River Formation. The Hellancourt Basalt concludes the sequence in the east of the Trough.

Allison Quartzite: At most points, south of Larch River, the Archean rocks are overlain by a thin- to thick-bedded, pink, feldspathic quartzite, itself overlain, in many places, by alternate laminae of light grey siltstone and reddish argillite. In many localities, these rocks are overlain by laminated siltstones which, in places, appear to grade into dolomite. The uppermost unit is almost everywhere a black shale, possibly equivalent of the Ruth Lake Formation of the Central Labrador Trough. The thickness of the formation averages 500 feet just south of Larch River (58°N).

North of that river, the entire formation is represented by 40 to 150 feet of black, grey or green quartzites occurring in beds 1-3 feet thick. The colour of the quartzite depends of the nature of the cement. Most quartzites contain about 95% quartz, the important other minerals being feldspars, chlorite and biotite. In places, thin layers of shale alternate with quartzite; magnetite-bearing shales are commonly found near the top of the formation. It appears, from the description of the quartzite north of the Larch River, that the Allison Formation resulted from deposition in shallow water of sediments derived from almost complete weathering of the source material.

Fenimore Formation: The Fenimore Formation is composed almost entirely of chemical sediments made up successively upwards of sulphide, oxide, and carbonate facies. As evidenced by abundant ooids and intraformational diastems illustrated by breccias consisting of chert fragments in a matrix of oolitic siderite, the depositional environment was turbulent and shallow. The succession is quite variable in detail; the general upward sequence is from ferruginous shale, cherty metallic ironstone (magnetite-jaspery ironstone, hematite-magnetite ironstone) to carbonate ironstone, and spotted chert.

The cherty metallic ironstones are banded siliceous rocks with alternating varying amounts of magnetite or magnetite and hematite; numerous carbonate interbeds are also found. The rock is steel blue both on fresh and weathered surfaces. It is essentially made up of abundant ooids more or less completely recrystallized.

The carbonate ironstone consists of bands of carbonate, mainly ferrodolomite, normally 2 to 8 inches thick, alternating with much thinner beds or lenses of chert. Spotted chert forms the omnipresent uppermost member of the Fenimore Formation. It is a rock with a characteristic saccharoidal texture with grains ranging in diameter from 0.01 to 0.5 mm. The spots are two to four millimetres in diameter and are due to siderite concretions.

Chioak Formation: The nature and history of the Chioak Formation is important in the history of the evolution of the northern part of the Labrador Trough. As explained by Bérard (1965, p. 65), "The evidence of deformation near the base of the formation indicates the tectonic activity that began with the deposition of the Chioak sediments. Faulting isolated certain sedimentary basins, causing a considerable accumulation of detrital material on the down-side".

In a few points, the Chioak Formation rests directly on basement, but elsewhere it rests on the Allison Quartzite or on the Fenimore Formation.

The basal member of the Chioak Formation is a rhythmically deposited, light grey siltstone, from one to ten centimeters thick and black fissile shale. Bérard (1965) separates those rocks from the Chioak under the name Dragon Formation. The higher members of the Chioak consist of conglomerates, sandstones, and shales. The conglomerates are of three main types, arkosic conglomerates with granite fragments, iron-bearing conglomerates consisting of rounded fragments of basement rocks and all formations underlying the Chioak, and cherty conglomerates made up of dominant chert pebbles in an arkosic matrix. The thickness of the formation varies from 600 to 800 feet.

Abner Dolomite: The Abner Dolomite is commonly fine grained and relatively pure, although some beds with a rough weathering surface contain much sand and silt. The thickness of the formation is between 100 and 500 feet with individual beds ranging from a few inches to six feet thickness. Beds are thicker at the base and thinner near the top of the formation. Box-like patterns of quartz bands and stringers are very commonly developed through differential weathering. Stromatolites are commonly found in the Abner Dolomite.

The contact of the dolomite with the underlying Chioak Formation is either gradational or oscillatory, as sandstone alternates with rather pure dolomite.

Larch River Formation: The Larch River Formation overlies the Abner Dolomite paraconformably. It is a thick marine sequence of fine-grained detrital material, locally with bands of dolomite, red sandstones, and ironstones. It is relatively undeformed to the west but deformation and metamorphism increase eastward. The Larch River Formation is intruded by numerous gabbro sills and is overlain by the thick Hellancourt volcanic sequence.

Hellancourt Formation: Most, if not all, of the volcanic rocks of the northern part of the Labrador Trough belong to the Hellancourt Formation. The formation, between 4,000 and 5,000 feet thick, is made up almost entirely of pillowed and massive flows of metabasalt. Minor agglomerates occur and some of the coarser-grained rocks may be meta-diabase sills. Although, in general, pillow lavas predominate near the top of the formation and massive flows towards the base, there are many exceptions.

Chemical analyses indicate that the metabasalts are of normal tholeiitic type. Curiously, neither necks nor dyke swarms have been found. Sauvé and Bergeron (1965, p. 32) have shown that the extruded magma was extremely fluid.

Thévenet Formation: The Thévenet Formation overlies the Hellancourt volcanics conformably. The formation, the thickness of which exceeds 2,000 feet, consists of zones of massive argillite and quartzite or of an interlayering of both rock-types.

Intrusive Rocks: Sills are the only important intrusive rocks in the area and are largely formed of metagabbro and meta-diabase with minor amounts of ultramafic gabbro and quartz diorite. The sills range in thickness from a few feet to about 3,500 feet, but few are thicker than 1,500 feet. Features suggestive of multiple intrusions have been seen in nearly all of those sills that exceed 1,000 feet thickness.

Northern sector

North of Payne River, rock units are generally thinner and progressively more metamorphosed, so that except for the Fenimore Formation, formations described south of Payne River have not yet been distinguished.

Rocks below the Fenimore Formation are quartzite interbedded with mica schists. The thickness of the sequence ranges from 20 to 50 feet. A quartzite rests over the basement rocks in many localities, but in other places mica schists are the basal rocks.

The Fenimore Formation usually rests conformably on quartzites and mica schists, but is found in places directly on top of the gneissic basement. It consists of three members, a lower unit composed of magnetiferous mica schists and silicate ironstones, a middle member of hematite-magnetite ironstone, and an upper spotted ironstone.

Pelitic schists conformably overlie the Fenimore Formation. They are green or grey to black with increasing biotite content. All are banded, quartz-rich bands alternating with others rich in biotite or chlorite. A dolomite band outcrops within the pelitic schists near the eastern margin of the Labrador Trough. The dolomite, beige to brown on weathered surface and pale grey on fresh surface, is cut by numerous chert veins. It is possibly the equivalent of the Abner Dolomite.

The pelitic schists are overlain in the central part of the Trough by volcanic rocks up to 9,000 feet thick. The volcanic rocks generally show pillow structures but are locally massive; they have been invaded by gabbro and ultrabasic sills. In a few localities, beds of slate, two to six inches thick are interbedded with the volcanic rocks. Some of these beds have been followed for a distance of four miles (Hardy, 1969, p. 6).

Intrusive Rocks: Numerous sills of metagabbros and serpentinites occur mainly within the volcanic rocks. The thin metagabbro sills are of uniform composition while the thick ones show evidence of differentiation. Where serpentinite is associated with metagabbro, the median part of the sill consists of pyroxenite and, the top part, of serpentinite. Maximum thickness of the sills is 2,000 feet.

A 36-square-mile body of unmetamorphosed hypersthene gabbro occurs in the central part of the Labrador Trough, just north of Payne River. Two associated lenses of peridotite almost two miles long are present along the eastern border of the body of hypersthene gabbro.

Mineral assemblages observed in the rocks of the Labrador Trough north of Payne River show an increase in the degree of metamorphism from west to east, from greenschist to almandine-amphibolite facies. The basement rocks have been more deformed along the eastern contact of the Trough than along the western contact.

THE SOUTHERN LABRADOR TROUGH*

Introduction

The term Southern Labrador Trough (Gastil and Knowles, 1960) is used here to designate that part of the Grenville Province containing recognizable Kaniapiskau strata. It is continuous with the Central and Northern Trough areas but is structurally distinct from them because it is marked by complex cross-folds, by large-size recumbent folds and nappes, and by strong metamorphism that may all be products of several periods of mountain building. Stockwell (1963) introduced the name "Normanville Fold Belt" for the same region; this name is not used here because the term 'fold belt' should be restricted to regions with one single predominating fold direction. Gill (1949, 1952) used the name "Wapussakatoo Mountains" for the Mount Wright-Wabush Lake area of the Southern Labrador Trough.

A comprehensive summary of the geology of the southern part of the Labrador Trough is outside the scope of this paper, and only a few of the major features will be discussed. Since the early work of Low (1897) a large amount of geological work has been done in this region by numerous mining companies who have made their findings available to the many government geologists and students working on theses in this area. Summaries of previous work are provided by Gross (1968), Jackson (1962), and others. Several reports, in addition to those referred to in this paper, contain information on various aspects of the geology. These include reports by Chakraborty (1966), Gleeson (1956), Mloszewski (1956), Phillips (1958, 1959), and Sinclair (1960, 1961).

Strata of the Kaniapiskau Supergroup cross the Grenville Front between Sawbill and Gabbro Lakes (Dufresne and Neal, 1956; Gill, 1949; Macdonald, 1960; Stockwell, 1961), from where they have been traced for another 190 miles to the Mushalagan and Matonipi Lake areas (Bérard, 1964; Murtaugh and Currie, 1969). West of longitude 69 degrees the northern boundary of recognized Trough rocks swings southward away from the Grenville Front (Fig. 12). The width of the Southern Labrador Trough varies from as little as 10 miles in the Shabogamo Lake area to 60 miles or more in the Pekans River-Mushalagan Lake region. Many of the gneisses east of the Shabogamo Lake-Moisie River region may be Trough rocks (Fahrig, 1967; Jackson, unpubl. data). The high degree of structural complexity, the recrystallization of the Trough rocks by regional and contact metamorphism, mainly in the regional almandine amphibolite facies, and generally poor outcrops make regional stratigraphic correlations and facies studies increasingly difficult toward the southwest. Only the formations of the Gagnon Group (Table IV) can be traced and correlated with any degree of confidence.

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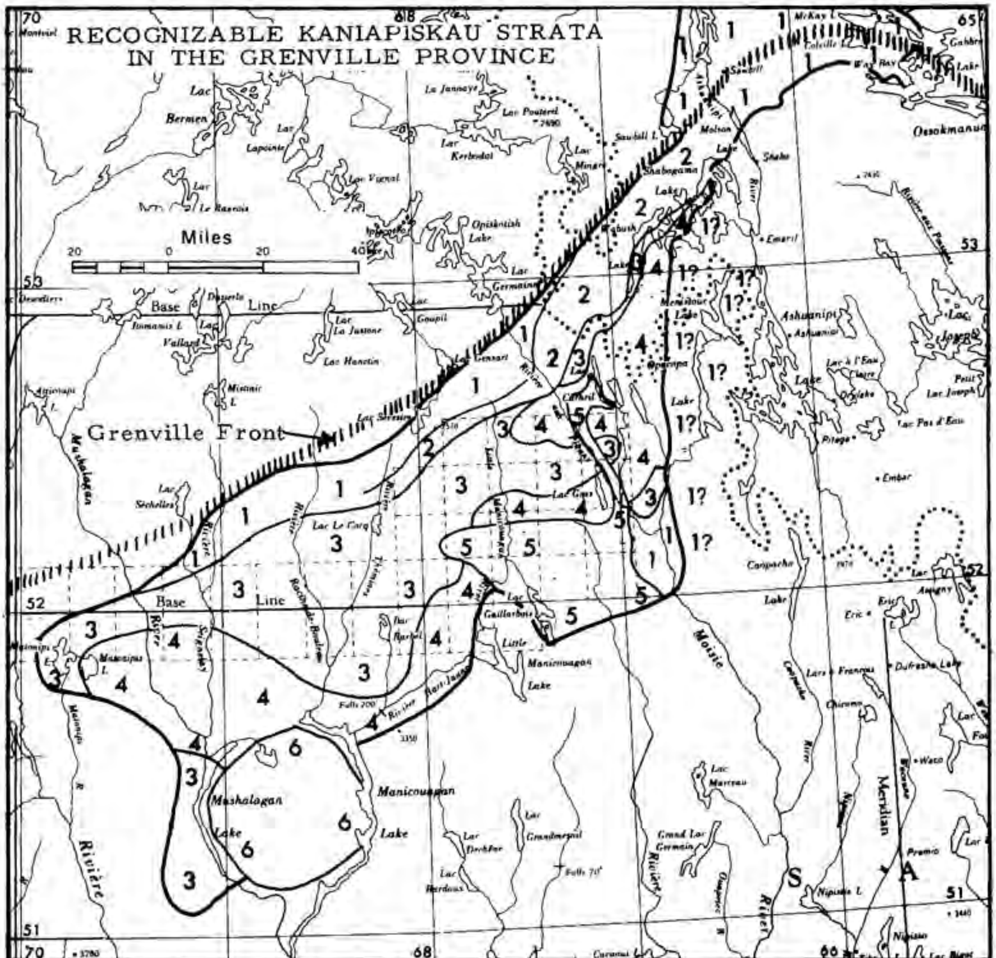
*Editor's Note: The chapter on the Southern Labrador Trough had not been written at the time of the workshop. The information it contains was therefore not presented to the participants.



1 - Undifferentiated, Kaniapiskau Supergroup.

2 - Wapussakatoo Formation present, Duley Formation absent.

3 - Wapussakatoo and Duley Formations present.

4 - Duley Formation present, Wapussakatoo Formation absent.

5 - Amphibolites possibly derived from volcanic rocks partially migmatized.

6 - Mainly Phanerozoic rocks.

Figure 12. Distribution of recognizable Kaniapiskau strata in the Grenville Province.

Structure

In general, structural complexity increases toward the south and southwest, and flow folding predominates. Two main structural trends, probably representing two major periods of deformation, have been recognized throughout much of the Southern Trough, and it is now generally recognized that a younger northwesterly trend has been superimposed on an earlier northeasterly trend (Clarke, 1967; Jackson, unpubl. data; Knowles, 1967). It should be noted, however that the axial traces of the northwesterly trending folds are approximately parallel to: (1) fold axial traces in the Central and Northern Trough, (2) metamorphic isograds in the Northern Trough, and (3) to the traces of axes of cross folds in the eastern Cape Smith Belt.

Northeasterly structural trends predominate along most of the trough Grenville Front, at least from Sawbill Lake southwest to Mistassini Lake, and large faults and shear-zones occur along parts of the Front (Chown, 1964, 1965; Fahrig, 1967; Gross, 1968; Jackson, 1962), and probably include thrust-faults. Major folds are commonly overturned to the northwest and include complex recumbent structures, some of which may actually be nappe structures. Results of age dating, discussed in the following section, lend support to the inference that some deformation probably occurred along the Grenville Front as early as Aphebian time, and perhaps even in the late Archean as suggested by Roach and Duffell (1968).

Southeast of the Grenville Front region superimposed northwest to north-south trends predominate, and folds are overturned to recumbent westward. There is some evidence, however that northeast trends again predominate along the southeast border of the area underlain by recognized Trough rocks, south of Gras and Felix Lakes (Clarke, 1967; Jackson, unpubl. data).

Post-Grenville faulting is present west of Wabush Lake (Jackson, 1962).

Metamorphism

In the southern part of the Central Trough, the grade of regional metamorphism increases southward and southeastward toward the Grenville Front (Fahrig, 1967; Goodwin, 1951; Jackson, 1952; Wynne-Edwards, 1961). From the Sawbill-Gabbro Lakes area south to Labrador City, the metamorphic grade increases both southwest along the Grenville Front and southeast away from it. Southwest of Labrador City however greenschist and lower amphibolite facies metamorphism predominates along the Front and in general, the metamorphic grade increases southeastward away from the Front up to the granulite facies.

West of Wabush Lake, Archean basement gneisses along the Grenville Front have been retrograded to the greenschist facies whereas coexisting Trough rocks have been prograded to the greenschist facies (Jackson, 1962). Southeast of the Front, basement rocks reappear as domes or as upfaulted blocks, the metamorphic grade of basement gneisses and Trough rocks increases southeastward away from the Front, and both groups of rocks reach granulite facies grade in the southern part of the Southern Trough (Clarke, 1960, 1965, 1967; Jackson, 1962, unpubl. data).

A zone of granulite facies rocks trends northeastward through the southern part of the Opocopa (23 B, E 1/2) topographic sheet (Jackson, unpubl.). This is probably an extension of the granulite belt shown by Douglas (1969) as extending from Quebec City to northeast of Manicouagan Lake. In the Opocopa sheet this zone contains both basement gneisses and Trough rocks, the metamorphic grade is again somewhat lower southeast of the belt.

Three radiometric ages by the K-Ar method on biotite were determined at the Massachusetts Institute of Technology (Hurley *et al.*, 1958, 1959) on samples collected by Jackson (1962) from the northwestern side of the Southern Trough. Dates of 1120 m.y., 1250 m.y., and 1580 m.y. were obtained, and

Jackson (1962) has suggested that the peak of metamorphism may have been reached at an earlier time along the northwestern side of the Southern Trough than farther south. Subsequent age determinations (Fahrig, 1967; Fig. 1 in Wanless *et al.*, 1968) have yielded late Aphebian and Helikian ages for several localities adjacent to the Grenville Front. These may represent a partial resetting of Archean and Hudsonian ages during a Grenville orogeny, as is customarily suggested. It seems equally possible, however, that at least some of these dates represent a complete resetting of the K-Ar clock for Archean basement gneisses and Aphebian strata during the Hudsonian orogeny (Fahrig, 1967), and during a Helikian period of metamorphism and deformation. Similar tentative conclusions have been reached by A.J. Baer (pers. comm., 1969).

Stratigraphy

Introduction

The stratigraphic relationships and lithologies of the Southern Trough strata and some associated rocks are summarized in Table IV. This succession seems to be characteristic of most of the Southern Trough. Within the Gagnon Group many apparent variations, recognized locally by early workers (e.g. Gross, 1955) have since been shown to be due to structural overturning of the normal stratigraphic succession (e.g. Clarke, 1960; Murphy, 1959, 1960). There is little doubt, however, that in places one or more of the formations is missing, or that some quartzite and/or marble overlie the iron-formation. The total thickness of Trough strata present in the Southern Trough is undoubtedly much greater than indicated by the figures in Table IV.

Ashuanipi Complex: Along the north side of the Southern Trough, Kaniapiskau strata overlie Archean gneisses and intrusions which include granulite facies rocks and charnockites (Clarke, 1960; Duffell and Roach, 1959; Eade, 1966; Jackson, 1962). The Trough rocks have been infolded and downfaulted into the basement gneisses (Clarke, 1960; Fahrig, 1967; Jackson, 1962), and basement rocks outcrop in dome-like structures in the Trough rocks. Granulite facies basement rocks reappear near the southern boundary of recognized Trough rocks (Clarke, 1965, 1967; Jackson, unpubl. data).

Katsao Formation: The actual contact between the Ashuanipi Complex and the Kaniapiskau strata is rarely exposed. The Katsao Formation, however, is believed to lie unconformably on the gneisses of the Ashuanipi Complex, and contains the oldest strata of the Kaniapiskau Supergroup recognized in the Southern Labrador Trough (Table IV). Mapping indicates that this formation is composed mainly of quartz-mica-feldspar schists and gneisses (Table IV) and that it is probably present throughout the southern part of the Labrador Trough.

Gagnon Group

The geology and distribution of the Gagnon Group (Table IV) has been compiled and summarized by Gross (1968), who refers to much of the available literature. Significant areas underlain by the Group, but not shown by Gross, occur between Greenwater Lake (52°53'N, 67°16'W) and Lac Montenon (53°7'N, 67°W) (Jackson, 1962; Neal, 1950, 1951); in the area between Wabush and De Mille Lakes on the north (53°N, 67°40'-67°52'W) and Felix Lake (52°24'N, 66°40'W) and the junction of the Pekans and Moisie Rivers (52°12'N, 67°50'W) on the south (Clarke, 1967; Jackson, unpubl.); and in the vicinity of Mushalagan Lake (51°15'N, 69°10'W) (Bérard, 1962; Murtaugh and Currie, 1969).

Duley Formation: The Duly marble lies conformably on the Katsao Formation and interfinger with it in the Wabush Lake region, where locally it may rest

Table IV. Table of Formations, Southern Labrador Trough.
(After Clarke, 1967; Duffell and Roach, 1959; Gastil and Knowles, 1960; Gastil *et al.*, 1960; Gill *et al.*, 1937; Gross, 1968; Iron Ore Co. Can., 1952; Jackson, 1962, NTS 23B East 1/2 - Unpub.; Macdonald, 1960; Neal, 1950, 1951).

		Sawbill Intrusions		Granite, quartz monzonite, granodiorite, syenodiorite, pegmatite
		Shabogamo Intrusions		Gabbro-diorite, diabase, anorthosite, amphibolite ± garnet; ultrabasic rocks
Kaniapiskau Supergroup		Nault Formation (Menihek)*	500'-1000' +	Phyllite; quartz-mica-feldspar schist and gneiss - some with sillimanite, kyanite, hornblende, garnet; hornblende-garnet amphibolite; graphitic schist; migmatite
Kaniapiskau Supergroup	Gagnon Group	Wabush Iron Formation (Sokoman)*	0'-1000'	Quartz-iron carbonate; olivine-pyroxene-carbonate; quartz-specularite; quartz-magnetite, quartz-magnetite-iron silicate; quartz-iron silicate
Kaniapiskau Supergroup	Gagnon Group	Huguette Formation (Ruth ?)*	0'-100'	Quartz-mica-feldspar schist, amphibolite, quartz-iron carbonate, chert
Kaniapiskau Supergroup	Gagnon Group	Wapussakatoo Formation (Wishart)*	0'-625'	Meta-orthoquartzite, meta-chert, ferruginous quartzite, garnetiferous quartzite, grit, conglomerate.
				D I S C O N F O R M I T Y
Kaniapiskau Supergroup	Gagnon Group	Duley Formation (Denault)*	0'-600 +	Marble, may contain quartz, phlogopite, amphiboles, pyroxenes; conglomerate.
Kaniapiskau Supergroup		Katsao Formation (Attikamagen)*	0'-1000' +	Quartz-mica-feldspar schist and gneiss - some with garnet, kyanite, hornblende; graphitic schist; migmatite.
		Ashuanipi Complex		Archean basement gneisses and intrusions, granulite, charnockite.

*Names in brackets are stratigraphic equivalents in the Central Labrador Trough unconformably on basement gneisses (Gastil and Knowles, 1960). A disconformity separates the Duley Formation from the overlying Wapussakatoo Formation in several places (Currie, 1956; Gill *et al.*, 1937; Murphy, 1959; Neal, 1951; Spat, 1959). Elsewhere the Duley Formation is gradational with the overlying Wapussakatoo and Wabush Formations. Murphy (1959, 1960) suggested that the Duley Formation is in part an offshore equivalent of the nearshore Wapussakatoo Quartzite, and that it is also isofacial with the lower part of the Wabush Iron Formation.

Wapussakatoo Formation: Gross (1955) has shown that at Mount Wright this formation is composed mainly of a white glassy meta-orthoquartzite. West of Wabush Lake, Jackson (1962) found this quartzite to be texturally different from recrystallized lean cherts of the Wabush Iron Formation. Any distinguishing sedimentary structures that may have been present have been destroyed by metamorphism and it is difficult to determine whether the formation is composed mainly of metamorphosed orthoquartzite or chert. The stratigraphically equivalent Wishart Formation of the Central Trough contains both, with orthoquartzite predominating. Mumtazuddin (1958) divided the Wapussakatoo Formation west of Wabush Lake into a lower garnetiferous muscovite schist member (30-35 ft.), a middle meta-orthoquartzite member 200 feet, and an upper calcareous quartzite member (30-50 ft.).

Huguette Formation: As much as 100 feet of quartz-mica-feldspar schist lies conformably between the Wapussakatoo and Wabush Formations in the Wabush Lake-Mount Wright region (Clarke, 1960; Gastil and Knowles, 1960; Gastil *et al.*, 1960; Jackson, 1962). The stratigraphy relationships (Table IV) and lithologies are relatively well exposed west of Huguette Lake ($52^{\circ}53'$N, $67^{\circ}9'$W).

Wabush Iron Formation: It has generally been concluded that only one major iron formation sequence is present throughout the Southern Labrador Trough (Gross, 1968), although locally a thin metamorphosed argillaceous rock separates the iron formation into two parts. The stratigraphy of the Wabush Iron Formation, its primary sedimentary facies changes and its secondary metamorphic facies changes have been studied by several workers (Clarke, 1960, 1967; Gastil *et al.*, 1960; Gross, 1955, 1958; Knowles and Gastil, 1959; Gastil and Knowles, 1960; Jackson, 1962; Klein, 1966; Kranck, S.H., 1961; Mueller, 1960, 1961). Because of intense metamorphism that stratigraphy is difficult to unravel. Recrystallized lean cherts resemble meta-orthoquartzites and decarbonation of siliceous iron carbonate rocks yields iron silicate rocks.

Oxide and silicate-carbonate-facies strata comprise the bulk of the Wabush Iron Formation. Each of these two main facies seems to predominate in different parts of the central part of the Southern Labrador Trough. Several large beneficiating orebodies occur within strata of the oxide facies at Labrador City, Wabush, Gagnon, and elsewhere. There is meagre evidence that silicate-carbonate strata may overlap oxide facies strata along the northwestern and southeastern margins of the Southern Labrador Trough.

Nault Formation: Quartz-mica-feldspar schist and gneiss form the main lithology of this formation (Table IV). It is believed to overlie the Wabush Iron Formation conformably throughout much of the Southern Labrador Trough (Fahrig, 1967; Jackson, 1962, unpubl. data; Macdonald, 1960; Neal, 1950, 1951); but because its strata cannot generally be differentiated from those of the Katsao Formation, Klein (1966) has been skeptical about the existence of the formation. West of Wabush Lake, strata believed to belong in the Nault Formation grade into (structurally) underlying Wabush Iron Formation and the section appears to be right-side up.

Nault Metavolcanics (?): A large area north and northeast of Little Manicouagan Lake is underlain by amphibolite and garnetiferous amphibolite that appear to be stratigraphically equivalent to the Nault Formation, and have been interpreted as metavolcanics (Roach and Duffell, 1968). Similar strata occur along the east side of the Pekans River between Gras and Carheil Lakes, and at many other localities throughout the Southern Trough. It would be difficult to differentiate metavolcanics from sheared metamorphosed basic intrusions in this region. It is suggested that these amphibolites may represent the eugeosynclinal zone as they appear to be in the right stratigraphic position. Volcanic-rocks outcrop west of Gabbro Lake (Jackson, 1952; Wynne-Edwards, 1961).

Shabogamo Intrusions: Gabbroic intrusions and derived amphibolites (Gross, 1955; Jackson, 1962) outcrop throughout the Southern Labrador Trough. They range in size from small sills and dykes to large stock-like bodies. South of Wabush Lake melanocratic gabbros are gradational into anorthosite (Jackson, unpubl. data), and Fahrig (1967) is probably right in relating these rocks to the anorthosite-gabbro complexes lying within the Grenville Tectonic Province, and that probably have an age of primary crystallization of about 1400 m.y. (Emslie, 1964). These rocks therefore are probably younger than and unrelated to basic intrusions in the Central Trough. The Shabogamo Intrusions have, however been deformed and recrystallized to varying degrees along with the Trough rocks, and they predate the Grenville orogeny. They may have been emplaced during a Helikian period of deformation (Jackson, 1962).

A large number of small ultrabasic bodies occur throughout the Southern Labrador Trough. Most of them have been metamorphosed, therefore predate the Grenville orogeny, and they may be part of the original Trough filling.

Sawbill Intrusions: These pegmatites, granitic dykes, sills, and stocks that intrude basement and Trough rocks near the Grenville Front and in the southern part of the Trough (Clarke, 1965, 1967; Jackson, 1962, unpubl. data; Kish, 1965; Neal, 1950, 1951) are evidently related to the Grenville orogeny (Goldich *et al.*, 1958).

Distribution of formations of the Kaniapiskau Supergroup

As in the rest of the Labrador Trough there is evidence of marine onlaps toward the northwest in the southern Labrador Trough. In the Wabush Lake region for example the Duley Formation is the first to die out toward the northwest, followed by the Katsao, Wapussakatoo and Wabush Formations until finally the Nault Formation rests directly on the Ashauanipi Complex (Jackson, 1962).

Generally speaking the three formations of the Gagnon Group (Fig. 12) form three parallel northeasterly trending zones (Clarke, 1967; Gastil and Knowles, 1960; Gastil *et al.*, 1960). The southeastern zone contains only the Duley Formation. The central zone contains both the Wapussakatoo and Duley Formations, and the northwestern zone contains only the Wapussakatoo Formation. The Wabush Iron Formation outcrops in all three zones.

These relationships are taken to indicate that the old shoreline extended south along the west side of Menihek Lake and then extended southwest into the Grenville Province region. If the Nault amphibolites are metavolcanics, then the Southern Labrador Trough may also have possessed a eugeosynclinal component which lay to the southeast of the presently recognized Southern Trough.

The Igneous Succession

Introduction

Igneous rocks that accumulated during formation of the Circum-Ungava geosyncline account for a large part of its contents. They are divided nearly equally between extrusive and intrusive components, the latter being somewhat more prominent in the Labrador Trough than in the Cape Smith-Wakeham Bay and the Belcher Islands segments. The extrusive rocks are mainly pillowed and massive basalts, with subordinate amounts of basic pyroclastic rocks, and minor quantities of acidic flows and tuffs. The intrusive rocks are sills that have preferentially invaded shaly rocks of the geosyncline in great profusion. Where they are thin and fine grained they may be virtually indistinguishable from the massive flows. There can be little question but that the sills were emplaced before folding and are therefore an integral part of the geosynclinal filling. They appear to come from the same magma as the extrusive rocks and their emplacement as sills rather than flows was probably a function of the structural environment into which they were introduced.

The distribution of the various types of igneous rocks in the Labrador Trough, Cape Smith-Wakeham Bay Belt, and Belcher Fold Belt is shown in Figures 16 and 17.

Volcanic Rocks

In the central part of the Labrador Trough volcanic rocks appear at a number of levels in the stratigraphic succession but are most voluminous in its upper part. They form part of the Seward, Attikamagen II and Sokoman Formations, but thicker successions are found in the Menihek, Murdoch and Willbob Lake Formations.

The oldest volcanic rocks in the succession are in the Seward Formation where an unknown thickness of massive flows and fragmental rocks is exposed near Lac Musset (lat. 55°30', long. 67°15'). They are highly oxidized rocks of possibly subaerial extrusion. Andesite fragments in the Chakonipau Formation noted previously may belong to the same episode.

The next known period of volcanism is represented by the Attikamagen II Formation where thickness of pillowed and massive basalts commonly range between 2,000 and 5,000 feet. The unit has fairly limited lateral extent. It is bounded by the latitudes 54°45'N and 56°30'N and is confined to the east-central part of the Trough.

Higher in the sequence, volcanic rocks (Nimish volcanics, Sauvé, 1953) are interbedded with and overlie the Sokoman Formation south of latitudes 54°45'N, near Astray and Petitsikapau Lakes. Near Astray Lake the volcanic rocks are in excess of 600 feet thick (Sauvé, 1953, p. 42). Sauvé (1953) reports that in the vicinity of Astray Lake the flows are generally massive and highly oxidized. These characteristics added to the association of the flows with conglomerates and massive tuffs led him to suggest that they were subaerial in part.

The major volcanic sequence in the central part of the Labrador Trough begins with a 3,000 to 4,000-foot thick assemblage of grey, rarely pillowed, basalts that cap the Menihek Formation just west of Murdoch Lake (55°40'N). The succession of pillow lavas that overlie the Attikamagen II Formation and extend northward to about 57°20'N is probably the equivalent unit (p. 14). This is succeeded by the Murdoch Formation of predominantly basic pyroclastic rocks that ranges from 2,000 to 6,000 feet thick and, following a thin mainly shaly succession (Thompson Lake Formation). The sequence culminates with the 15,000-foot thick assemblage of pillowed and massive basalts of the Willbob Formation.

The Murdoch Formation is unique within the Trough assemblage. It has considerable lateral extent and, although composed almost entirely of basic pyroclastic rocks, does contain or is closely associated with rare pods and lenses of rhyolitic flows and tuffs (Frarey, 1961; Baragar, 1967). It has only been recognized south of 56°N but as previously noted (p. 12) what may be in part its equivalent is a basic volcanic unit that was mapped by Fahrig (1965) and Roscoe (1957) between 56°40'N and 57°30'N and that shows a substantial pyroclastic content in its upper part. It also contains lenses and pods of rhyolitic rocks (Fahrig, 1965; Hasimoto, 1964). Apart from these two sets of occurrences, rhyolites are unknown in the Circum-Ungava geosyncline. They therefore, constitute strong evidence for the correlation of the two units.

The Willbob Lake Formation is restricted to one segment of a more or less continuous belt of pillowed and massive basalts that line the eastern side of the Labrador Trough. In all places the volcanic assemblages that compose this belt are at, or near, the top of the local geosynclinal sequence and all of them are probably parts of a single unit of widely varying thicknesses. In the Wakuach Lake area (55°-56°N) the unit is in excess of 15,000 feet thick (Frarey, 1967; Baragar, 1967) near Hellancourt Lake (58°N) it is 4,000 to 5,000 feet thick (Sauvé and Bergeron, 1965) and at the north end of the Trough, it is 9,000 feet thick (Hardy, 1968). Presence of certain distinctive glomeroporphyritic flows (blotchy flows) in both the Willbob (Fahrig, 1964; Baragar, 1967) lavas at the southern end of the Trough and the Hellancourt lavas (Sauvé and Bergeron, 1965) at its northern end is an additional element favouring correlation along the entire belt. The glomeroporphyritic flows are assumed to be the surface expressions of glomeroporphyritic gabbro sills (blotchy gabbros, leopard rock) that are characteristic of the Trough throughout its length.

Sills

Three major varieties of sills can be distinguished in the Labrador Trough; those of normal gabbro or metagabbro, of glomeroporphyritic gabbro, and of ultramafic rock. Normal gabbro sills are the most abundant and are thickly distributed in shaly members of the assemblage, along the central or east-central parts of the Trough. Glomeroporphyritic gabbros are coarse feldspathic rocks spotted with spherical feldspar aggregates up to 15 cm in diameter. The margins of these sills are formed of chilled, normal gabbro and it has been proposed that the concentration of feldspathic aggregates in the centre of the sills is the result of a type of flow differentiation called plug-flow (Baragar, 1960). Glomeroporphyritic gabbros are confined to a fairly narrow belt in the eastern part of the Trough. The ultramafic sills occur within or close to the Willbob-Hellancourt lava belt. They may be composed entirely of ultramafic rock or more commonly of an ultramafic lower and gabbroic upper part. In either case they are generally found to possess somewhat gabbroic borders. These features too are attributed to flow differentiation (Fahrig, 1953, 1962; Baragar, 1967). Ultramafic lenses (Taylor, 1969) that intrude schists and gneisses east of the Trough (Fig. 16) are probably of related origin.

In the Wakuach Lake area (55°N-56°N) the sills have been estimated to have a minimum aggregate thickness of 20,000 feet (Baragar, 1967). Individually they range in thickness from about 100 feet to more than 3,000 feet and some can be traced for many tens of miles along strike. One of the thicker of the ultramafic sills can be followed, with minor interpretation, for a strike distance of 120 miles. In contrast, ultramafic bodies that intrude the crystalline rock east of the Trough are lenses of rarely more than four or five miles in length.

The age relations of the sills to one another and to the surface deposits of the Labrador Trough cannot be determined precisely. Normal gabbros are compositionally similar to lava flows of the basalt-shale sequence (Attikamagen II), of the Menihek Formation, and of the Willbob Formation. It is possible therefore that they were intruded throughout much of the history of the geosyncline. However, since the basalt-shale sequence itself, and the much younger Menihek and Thompson Lake Formations are particularly profusely intruded by normal gabbro sills, it seems logical to relate most of them to the extrusion of the Willbob lavas. In the case of the glomeroporphyritic sills which have counterparts in the glomeroporphyritic flows of the Willbob and Hellancourt Formations this can be done with confidence. This is also the case for those normal gabbros closely enough related in time with the glomeroporphyritic gabbros to form composite sills with them. This group includes most of the swarm that invades the Menihek Formation of Wakuach Lake map-area (Baragar, 1967). Ultramafic sills which typically intrude the Willbob-Hellancourt lava belt or the sediments immediately underneath come obviously late in the geosynclinal sequence. In general, most of the igneous filling of the geosyncline was probably added late in its history.

Eugeosynclinal-Miogeosynclinal Relations

The distribution of volcanic rocks in the Circum-Ungava geosyncline, shown in Figure 16 and 17, marks roughly the extent of the eugeosynclinal part of it. It represents a more or less continuous zone on the distal side of the geosyncline away from the craton. The boundary between miogeosyncline and eugeosyncline is approximately the median line of the Labrador Trough, but because of the spread of sills away from their source and of the shift of volcanic centres with time, it is difficult to define precisely.

Eugeosynclinal deposits appear to have been thickest in the region between latitudes 55°N and 56°N and from there to have thinned both northward and southward. The thickest section may have measured in the vicinity of 45,000 to 50,000 feet, including 20,000 feet of sills; a thickness that is of the same order of magnitude as the thickest eugeosynclinal pile anywhere. In contrast, miogeosynclinal deposits of the western part of the Trough at about the same latitude may have a typical aggregate thickness of 10,000 feet.

THE CAPE SMITH-WAKEHAM BAY BELT

The Cape Smith-Wakeham Bay Belt trends in an easterly direction across the northern tip of Quebec. The belt is 235 miles long, up to 60 miles wide, and covers roughly 6,000 square miles. Aphebian sedimentary, volcanic and intrusive rocks underlie the belt; their metamorphism increases northwards from the lower greenschist facies to the amphibolite facies. The Aphebian rocks rest unconformably on a basement complex composed of quartzofeldspathic gneisses and of granite that underlie large areas north and south of the belt.

Regional Geology

Two distinct groups have been recognized in the Cape Smith Belt: the Povungnituk or lower Group and the Chukotat or upper Group (Bergeron, 1957c). They are separated by an angular unconformity; a basal conglomerate is at many places at the base of the Chukotat Group. The existence of these two groups has been questioned (Stam, 1961), but age determinations (Beal *et al.*, 1963) tend to support their presence.

The Povungnituk Group comprises mainly pelitic sediments, massive and pillowed basalts, and intrusions of gabbroic and, in a few places, noritic sills. Minor dolomite, sandstone and iron formation are present. The Chukotat Group is composed in large part of pillowed basalts with ultrabasic and gabbroic sills and some slates and tuffs. This group is much less deformed and metamorphosed than the lower group. In places erosion has exposed inliers of the Povungnituk rocks below the Chukotat, whereas small outliers of the upper group overlie Povungnituk rocks in the southern part of the belt.

A large thrust-fault, extending eastward at least 225 miles from Hudson Bay, abruptly terminates the Chukotat Group. This fault dips to the north, and the rocks north of it are considered to belong to the Povungnituk Group. They include an assemblage of sedimentary, volcanic and intrusive rocks which are, in general, very schistose and highly metamorphosed.

The following description applies to the central part of the belt which has been investigated in more detail than any other part (Bergeron, 1957c, 1958; Beall, 1959, 1967; Shepherd, 1959; Gold, 1962; De Montigny, 1962, and Gélinas, 1962).

Geology of the Central Part of the Belt

Table V gives the sequence of rocks along longitude 74°N. These rocks will be briefly described from south to north.

Archean Basement in the South: The most common rock type of the Archean basement is a foliated, grey to buff granodioritic gneiss composed mainly of equal amounts of plagioclase, usually andesine, microcline, and quartz. Biotite-hornblende gneisses and amphibolites form more or less continuous concordant bands within the granodiorite. The strike of the foliation is variable, but a general northerly to north-northeasterly trend is noticeable on aerial photographs.

Pink and green granites and granodiorites mainly composed of quartz, perthite and plagioclase (oligoclase or andesine) appear to be intrusive into the gneisses but locally also grade into them.

Povungnituk Group: The Povungnituk Group is largely composed of mica schists, chlorite-actinolite schists, metabasalt, quartzite, dolomite, calcareous arkose, and silicate ironstone intruded by gabbro sills.

Mica schists composed mainly of quartz, chlorite, muscovite, biotite and albite predominate; they underlie large areas south of Povungnituk River and locally contain graphite or dolomite. In some areas, layers of dark green schistose volcanic rocks are interbedded with the mica schist. Several lenticular layers of grey to white, sugary grained, massive or banded quartzites are present. Dolomite and calcareous arkose occur at some localities. The dolomite is fairly resistant and its characteristic orange-brown colour on weathered surfaces and its pitted surface make it readily distinguishable in the field. A few horizons of creamy-grey, medium-grained, massive calcareous arkose are also found in places.

Silicate ironstone of an unusual mineralogy is exposed along Povungnituk River. The rock is layered and composed largely of grunerite with minor calcite, stilpnomelane, magnetite and quartz. The mineralogy and phase equilibrium of this ironstone have been studied by Hashimoto (1962).

Relatively small volumes of greyish green or green, massive and pillowed basalt are present in the Povungnituk Group. Pillows have an average length of two to three feet. They are usually flattened and rarely useful for top determination. The basalts have been converted to albite-epidote amphibolites that contain generally more than ten per cent biotite and occasionally potash feldspar. Beall (1959, p. 27) used the biotite content in order to distinguish these basalts from potash poor, but otherwise similar rocks of the Chukotat Group.

Many metagabbro sills, some over 1,000 feet thick, intrude the metavolcanic rocks of the Povungnituk Group.

Chukotat Group: The Chukotat Group is mainly composed of pillowed basalts intruded by gabbroic and ultrabasic sills. The unconformity separating the Povungnituk and the Chukotat Groups can be observed at many points. Chert, impure sandstone or a sedimentary breccia grading into conglomerates is at the base of the Chukotat. The chert is banded or massive, grey to blue-black. It is rarely more than ten feet thick. The chert rests on volcanic rocks and a weathered surface is exposed at the contact. The basal sandstone is dark grey and consists of fragments of volcanic rocks, quartz, mica, and potash feldspar, in a chloritic matrix. It grades locally into slaty rocks. The sedimentary

Table V: Rock sequences northward along longitude 74°.

N	1655- 1715 m.y.	Rejuvenated Archean Basement	Quartzo-feldspathic gneiss.
	Folded Unconformity		
		Povungnituk ?	Quartz diorite, serpentinite, granite, metagabbro, amphibolite. Chlorite-actinolite schist, metabasalt. Dolomite, mica schists, banded quartzite.
	Northern Thrust Fault		
Proterozoic	1400- 1550 m.y.	Chukotat Group	Gabbro, metagabbro, peridotite, serpentinite, pyroxenite. Metabasalts, tuff, argillite, quartzite. Conglomerate, chert, sandstone.
	Angular Unconformity		
	1700- 2200 m.y.	Povungnituk Group	Gabbro, metagabbro. Basalt, metabasalt. Phyllite, quartz mica schist, chlorite schist, actinolite schist, quartzite, dolomite, silicate ironstone.
Major Unconformity			
Archean S	2300- 2800 m.y.	Mainly granodiorite gneiss, some massive granite and granodiorite, amphibole gneiss, biotite gneiss, amphibolite.	

breccia is composed of angular fragments of massive basalt, quartz, granite, jasper, and schist in an arkosic matrix and it grades into a conglomerate described by De Montigny (1959, p. 4).

Siliceous and graphitic slates are here and there interbedded with the lavas. At the contact with intrusive sills they have been recrystallized into rocks of variable mineralogy. Shepherd (1959) describes a rock composed of quartz, tremolite, clinozoisite and, occasionally, grossularite. In places the siliceous and graphitic slates grade into tuffs. Impure sandstones are here and there associated with them.

Well pillowed basaltic lavas form large cliffs. Beall (1960) distinguished two main types: olivine basalt (predominant) and locally important leucobasalt. The total thickness of the basalt in the Chukotat Group exceeds 15,000 feet.

"Mineralogically, the olivine basalt, originally composed of augite, plagioclase, olivine, and magnetite, all of which are frequently visible in thin section, is now largely altered to an actinolite-clinozoisite-serpentine-chlorite-magnetite rock. Serpentine or chlorite pseudomorphs after olivine, plumose clinozoisite-albite aggregates after plagioclase, and actinolite alteration of augite generally occur with devitrified glass, giving the rock a green to dark-green, fine-grained appearance.

The leucobasalt, of similar mineralogy, but pale green to whitish in hand specimen, differs from the olivine basalt in that the quantity of pale actinolite is greater, and that of the darker chlorite and serpentine less. The percentage of original augite was greater, and that of olivine, presumably less" (Beall, 1967, pp. 33-34).

The Chukotat sills can be divided into three types: ultrabasic, gabbro and composite. All workers in the Cape Smith-Wakeham Bay Belt have stressed the common intimate association of gabbro and ultrabasic sills.

Beall (1967, pp. 37-47) has described the three types and has proposed a hypothesis concerning the mode of sill injection. The following discussion is based on his model.

The ultrabasic sills are generally serpentinite and range in thickness from a few feet to more than 1,000 feet. They can be traced for up to ten miles or more. They generally show minor variation in composition and texture, the former ranging from serpentinite to porphyritic pyroxene-serpentinite where altered, and from peridotite to dunitic peridotite where less altered. Their contacts with slates are sharp whereas contracts against basalts are gradational, without unusual minerals in either rock.

The gabbro sills range in thickness from less than 50 to more than 1,000 feet. They also have been traced for long distances along strike. The thicker sills exhibit fractional crystallization and gravitational differentiation, grading from pyroxenite at the base through gabbro and diorite to quartz diorite at the top. The composition is generally typical of gabbro, with variations from olivine gabbro to feldspathic gabbro.

Gabbro and ultrabasic sills are commonly associated. They could appear to be composite sills, formed by two separate injections. This hypothesis has been proposed by Shepherd (1959), who suggested that the ultramafic material had been injected first. Some observations suggest however, that the twin sills are not composite, but are the result of differentiation in situ. In particular, the temperature indicated by the low amphibolite facies metamorphism in surrounding sediments appears to be too low for the double-injection mechanism. The ultramafic portion of the twin sills is at or near their base. There is evidence of differentiation by fractional crystallization within the thick ultrabasic portions of the sills. Contacts between gabbro and ultrabasic rocks show unusual features, such as interlayering of serpentinite and pyroxenite (the latter a common basal phase of the gabbro), and tongue-like masses of serpentinite protruding into overlying gabbro.

"These observations are all compatible with the following hypothesis of sill formation:

"The igneous material upon intrusion consisted of a gabbroic liquid in equilibrium with and saturated with a mass of ultrabasic crystals, mainly olivine. After intrusion, the crystals settled to the base in a uniform mass. The gabbro then proceeded to crystallize with consequent gravitational differentiation, the underlying ultramafic crystal mush not yet having consolidated. Minor convection produced irregular contact effects such as clot inclusions of ultramafic swirls, and convectional interlayering of settled olivine crystals and lighter newly formed pyroxenes crystallizing from the gabbro.

"If the injected material was largely ultrabasic and crystalline with only minor interstitial aqueous or gabbroic fluid, a simple ultrabasic sill resulted. If it was largely gabbroic liquid with few or no crystals, a simple gabbro sill resulted. If, however, the injected material was a mixture of crystals and liquid, a settled differentiated sill of ultrabasic material capped by gabbro developed" (Beall, 1967, pp. 41-42).

The Povungnituk (?) Group of the northern part of the belt: The lowest rocks in the north of the basin are laminated garnetiferous biotite-muscovite schists containing up to one inch thick laminae of quartz-feldspar schist with little muscovite and biotite. Garnet forms porphyroblasts, four to twenty millimeters across. Graphitic quartzites, in beds up to 25 feet thick, and lenses of metamorphosed ironstone, impure dolomite, and amphibolite are associated with the garnetiferous schist.

Farther south, the biotite-muscovite schists grade into or underlie chlorite-sericite schists with layers or bands of pyritic slates and, near the top of the sequence, thin bands of dolomite.

The sedimentary rocks are overlain by metavolcanic rocks of basaltic composition. Massive and pillowed lavas that are strongly sheared in many places occur in the south of the zone. Chlorite-actinolite schists probably of volcanic origin are also present, and dull-white, faintly schistose clinozisite-plagioclase rocks are intimately associated with them.

Sills of metagabbro and ultrabasic rocks have been intruded at or near the base of the metavolcanic rocks. They comprise serpentinite, tremolite serpentinite, and serpentine amphibolite. Small masses of quartz diorite and granite have been observed to cut the chlorite-actinolite schists.

Rejuvenated Northern Basement

The area north of the Cape Smith-Wakeham Bay Belt is composed of distinctly layered quartzo-feldspathic gneisses. They are composed of microcline, plagioclase, quartz, biotite, and muscovite. Garnet or hornblende is locally present; some bands of amphibolite are interlayered with the quartzo-feldspathic gneisses. The dominant trend of the complex is east-west, with open, irregular folds. The northern basement has been remobilized during the Hudsonian orogeny, and the resulting mineralogical and tectonic changes have not yet been studied.

Historical Geology of the Cape Smith-Wakeham Bay Belt

The rocks of the Povungnituk Group have been deposited on Archean rocks and moderately to highly metamorphosed. Potassium-argon dates have been determined (Beall *et al.*, 1963) on several Povungnituk schists, the resulting ages ranging from 1430 to 1650 m.y. Metamorphism in that Group could be as early as 1600 m.y. B.P. The same metamorphic event also affected granitic rocks of the basement to the south which yield potassium-argon ages of 1660 ± 30 m.y. (Beall, 1967).

The Chukotat Group was deposited after the metamorphism of the Povungnituk Group and after an important period of erosion. The Chukotat rocks are only slightly metamorphosed. "Potassium-argon dates on two slates from the Cross Lake and Laflamme Lake areas show dates of 1430 ± 30 and 1490 ± 50 m.y. This suggests that the Chukotat strata were rapidly deposited and deformed between 1400 and 1550 m.y., which seems to be the earliest date which could have allowed escape from the stronger metamorphic event which affected both the Povungnituk and basement rocks" (Beall, 1967, pp. 57-58). The thick gabbroic and ultrabasic sills of the Chukotat Group were emplaced prior to the periods of folding.

THE BELCHER FOLD BELT

General Features

The Belcher Fold Belt extends for over 500 miles from Cape Smith southward to west of James Bay (Fig. 13). The position of the eastern boundary between Cape Smith and James Bay, and of a small part of the southern boundary

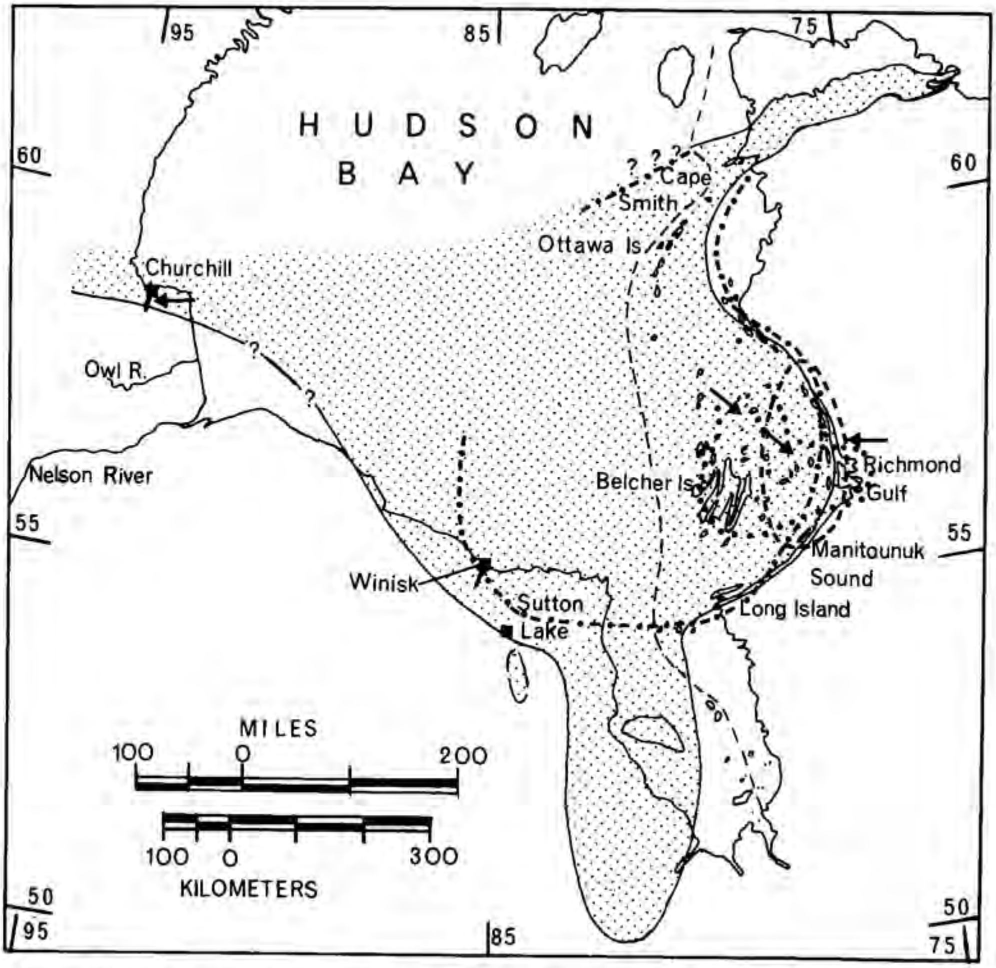


Figure 13. Inferred distribution of Aphebian strata in the Belcher Basin.

near Sutton Lake, is well documented. Along much of the eastern boundary an angular unconformity separates the strata of the Belcher Fold Belt from the underlying Archean basement gneisses of the Superior Province.

The basement gneisses have not been intensely deformed or recrystallized since the Belcher Basin strata were deposited, and give Archean K-Ar ages (Fig. 1 of Wanless *et al.*, 1968). Some block faulting, however did occur during and after deposition of the Aphebian strata, and in some places a fault contact separates the Aphebian and Archean rocks (Bell, 1879; Low, 1902; Sanford *et al.*, 1968; Bostock, in press; Woodcock, 1960). Gentle folding has affected the Aphebian and some of the adjacent basement rocks; topographic highs on the Archean basement surface, as at Castle Peninsula in Richmond Gulf, have been accentuated along the crests of anticlinal warps, so that the highs now appear as dome-like structures.

Several islands and shoals occur between Richmond Gulf and the Belcher Islands. Examination of airphotos suggests that the intensity of folding in the region probably increases gradually westward from Richmond Gulf toward the Belcher Islands where closed anticlines alternate with broad open synclines.

From a cross-section across the Belcher Islands (Jackson, 1960) it is estimated that the Aphebian strata have been shortened, mainly by folding, at least by thirty per cent in an east-west direction. The folding at the surface is cylindrical and if this persists at depth then a décollement zone may mark the contact with the underlying basement rocks.

The western extent of the Aphebian strata deposited in the Belcher Basin is largely inferred, as very little bedrock information is available. The distribution shown on Figure 13 is based partly on some drill hole information (Hogg *et al.*, 1953; B.V. Sanford, pers. comm.; Satterly, 1953), partly on the lithological similarity of the quartzites at Churchill (Unit 5 of Bostock, 1969a) to Unit 15 (pers. obs.) on the Belcher Islands, and partly on geophysical and bathymetric surveys of the Hudson Bay region (Hobson, 1969; Hood *et al.*, 1969; Innes *et al.*, 1967, 1968; Morley *et al.*, 1968; Pelletier *et al.*, 1968).

The boundary in the James Bay area is extended farther south than Bostock (1969b) shows it, mainly on the basis of two drill holes. In one (51°45'N lat., 80°40'W long.) about seven feet of gabbro-diabase and brown sandstone underlie Paleozoic strata (Hogg *et al.*, 1953). In the other (50°20'N lat., 81°50'W long.) 136 feet of amygdaloidal basalt and red argillite occur between the Paleozoic strata and Archean basement gneisses (Satterly, 1953). Bostock (in press) interprets Aphebian strata as underlying Paleozoic strata for at least another 40 miles northwest of Winisk (Fig. 13). Information from holes drilled in the northwestern part of the James Bay Lowlands, southeast of Churchill, indicates that unmetamorphosed Aphebian strata are absent in this area. Results of K-Ar age dating (Fig. 1 in Wanless *et al.*, 1968) suggest that if Aphebian strata are present under the Paleozoic strata between 56° north latitude and Churchill, they are metamorphosed.

Bathymetric and magnetic data support the interpretation that Aphebian strata extend continuously under water from Cape Smith to James Bay. Bathymetric and seismic data also indicate that the eastern limit of Phanerozoic cover rocks probably coincides with a line (Fig. 13) drawn northward through James Bay, west of the Belcher and Ottawa Islands to the northwestern tip of Ungava (Hobson, 1969; Hood *et al.*, 1969; Pelletier *et al.*, 1968). The greatest distances between islands are only 45 miles between the Marcopeet and Farmer Islands, and 60 miles between the Ottawa Islands and Cape Smith.

Hood *et al.* (1969) have noted the general agreement of gravity and magnetic trends in the Hudson Bay region. These trends probably represent at least in part, those of Aphebian strata, and indicate that the unmetamorphosed strata of the eastern Belcher Basin extend westward toward the west coast of Hudson Bay where deformed and metamorphosed equivalents may be present.

A belt of magnetic and negative gravity anomalies (Morley *et al.*, 1968; Innes *et al.*, 1967, 1968) trends southwest from Cape Smith to west of the Ottawa Islands and thence to the Sutton Lake area. From this point, the anomalies trend westward, parallel to the tectonic fabric of the region, into the Superior-Pikwitonei-Churchill Province boundary area (C.K. Bell, pers. comm.). This belt may mark a transition zone from unmetamorphosed, moderately deformed strata on the east to metamorphosed highly deformed equivalents on the west (B.V. Sanford, pers. comm.). Another interpretation, here considered to be less probable, is that this belt represents the western limit of Belcher Basin strata.

The Bouguer and free air gravity anomalies in the Hudson Bay region have been studied by Innes *et al.* (1967, 1968) who have suggested that the greater part of Hudson Bay may be underlain by considerable thicknesses of Proterozoic sedimentary and basic volcanic rocks and/or gneissic equivalents. They suggest also that the volcanics may be flood basalts.

Most of the Belcher Basin strata have undergone only a slight amount of recrystallization during diagenesis or a period of very low-grade regional metamorphism (Bostock, in press; Eade, 1966; Hofmann and Jackson, 1969; Lee, 1965; Stevenson, 1968). The metamorphic grade apparently increases slightly from south to north toward Cape Smith, and according to Lee (1965) Aphebian strata on the Hopewell Islands have been metamorphosed to the quartz-albite-epidote-biotite subfacies of the greenschist facies of regional metamorphism.

Several whole rock K-Ar age determinations have been carried out on basic extrusive and intrusive rocks of the belt, and results are somewhat contradictory. Ages for Belcher Island rocks fall into an older (1620-1693 m.y.) and younger (830-1054 m.y.) group (Jackson, 1967). A basalt sample from the Ottawa Islands gave 1595 m.y. and another from the west side of Richmond Gulf gave 1385 m.y. (Stevenson, 1966). The older ages probably represent the effects of a low-grade regional metamorphism during the Hudsonian orogeny, the younger ages may represent a thermal event and/or a mild disturbance in late Neohelikian time when a few trap dykes were emplaced (Hofmann and Jackson, 1969).

Stratigraphy

Introduction

The filling of the Belcher Basins to be cyclical (Table VII) as is that of the Labrador Trough. This feature is most obvious in the Belcher Group (Hofmann and Jackson, 1969) which provides a continuous stratigraphic section ranging in thickness from 20,000 to 30,000 feet in which no unconformity, other than very local, has yet been recognized (Figs. 14, 15). On the Belcher Islands order of deposition within an individual cycle is reversed relative to that in the Labrador Trough. Belcher Island cycles began with a pulse of volcanic activity and basin instability. Following this, conditions gradually reverted to more stable conditions with minor setbacks at various intervals. Thus in one cycle basic volcanism is followed by deposition of greywackes, tuffs and argillites. These grade upwards into interbedded and massive dolomites and orthoquartzites.

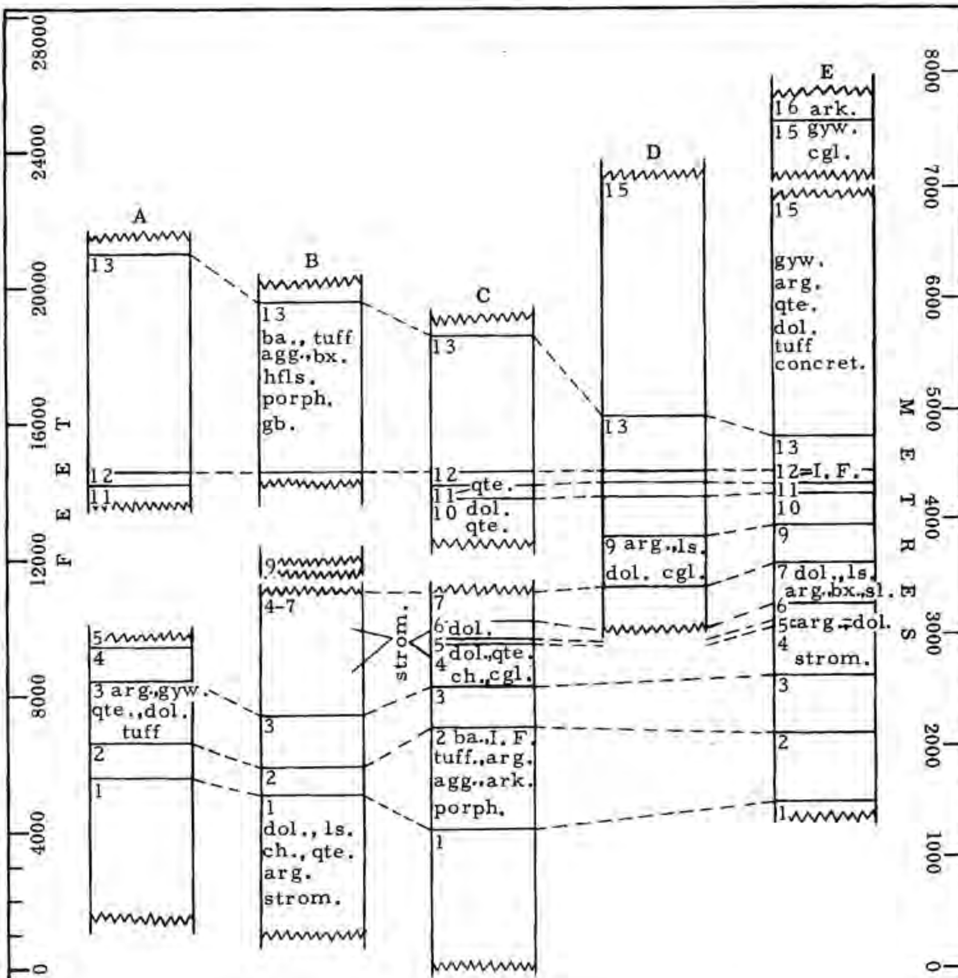
Basic igneous rocks form a larger proportion of the miogeosynclinal filling in the Belcher Basin than in the Labrador Trough. In addition ultrabasic rocks are relatively uncommon in the Belcher Basin and the basic volcanic rocks are mainly spilitic, whereas those of the Labrador Trough are tholeiitic.

Basic extrusive and/or intrusive rocks comprise about twenty per cent of the exposed rocks at Richmond Gulf and on the eastern Belcher Islands (Jackson, 1960; Low, 1902; Woodcock, 1960). They form about thirty per cent of the section on the western Belcher Islands, in the Sutton Lake area (Bostock, in press), and in the Hopewell Islands (Bell, 1879; Lee, 1965). The South Sleeper, Farmer, Twin, and Ottawa Islands are composed mainly of basic extrusive and intrusive rocks (Bell, 1879; Manning, 1947). Some iron-formation has been reported from the Sleeper Islands (J. Kilualuk, pers. comm.). Basic igneous rocks form most of the rocks in the western part of the Cape Smith Belt, and sedimentary strata probably do not predominate until some point to the southwest between the Ottawa and Sleeper Islands.

The cyclical nature of the filling in the Richmond Gulf area is similar to that found on the Belcher Islands but is complicated by the presence of unconformities (Bell, 1879; Low, 1902), and is less well documented. At least two unconformities have been recognized in the Richmond Gulf area (Bell, 1879; Stevenson, 1968; Woodcock, 1960). Low (1902), however, considered the upper unconformity (Table VII, Fig. 15) to be a thrust fault.

Three main interpretations have been proposed to explain the stratigraphic relations of rocks on the Nastapoka Islands to those on the mainland (Parks, 1949).

1. The strata on the Nastapoka Islands overlie those at Richmond Gulf as part of a continuous homoclinal sequence dipping westward (Bell, 1879; Leith, 1910; Woodcock, 1960). This interpretation is favoured by the writers who estimate that as little as 600 feet of stratigraphic section may be covered by Nastapoka Sound (Fig. 15).



A – West: Churchill Sound-Johnson Island.
B – South Central: South Kasegalik Lake.
C – North Central: Eskimo Harbour-Haig Inlet.
D – East Central: Kipalu Inlet.
E – East: Tukarak Island-Bakers Dozen Islands.

Map-unit 8 represents units 4 to 7 undivided; unit 14 represents basic sills and dykes. These units are not shown in this figure. Lithology of each unit is similar in all five sections.

Figure 14. Generalized stratigraphic sections, Belcher Islands, N.W.T.

2. The quartzites and iron-formation on the Nastapoka Islands are correlative with the quartzite and dolomite underlying the upper mainland flows ("Cap Rock"). They reappear on the mainland due to the existence of a fold in the water-covered zone between the two localities.

3. The Nastapoka Islands strata are correlated as in (2) above, but have been thrust up along a fault running between the islands and the mainland (Low, 1902; Harwood, 1949; Kranck, E.H., 1951; Parks, 1949). In support of this hypothesis Low (1903) has described reverse faults on the Nastapoka Islands that have displacements of up to 100 feet, and which have repeated the upper beds exposed on the islands. Such faulting could easily account for the 600 feet (this paper) to 1,000 feet (Bell, 1879; Woodcock, 1960) of section that is covered by Nastapoka Sound.

As pointed out by Eade (1966), considerable confusion has developed in the naming of rock units in the Richmond Gulf area. The major usages have been discussed by Eade (1966) and Stevenson (1968), and are summarized in Table VI. It is proposed here that the name Manitounuk be raised to Supergroup status and that Woodcock's (1960) nomenclature be retained for the three subdivisions on the basis that the presence of the two unconformities seems to be well established.

In the following discussion, each particular formation of the Belcher Group will be discussed, and then correlated to probably equivalent strata elsewhere in the Belcher Fold Belt.

The First Cycle

The unconformity between the Archean gneisses and the overlying Aphebian strata is exposed at several places from Richmond Gulf south, to Long Island. The lowest strata exposed along this contact outcrop at Richmond Gulf, an embayment that was already in existence at the time of deposition, and where a sequence of arkosic sediments comprises the basal Aphebian formation. Elsewhere, younger strata overlap onto the basement gneisses, indicating marine onlap. The Archean-Aphebian contact is at least in part a fault contact in the Sutton Lake and Hopewell Islands areas. The base of the Aphebian section is not exposed on the Belcher Islands, and only the strata deposited in the latter part of the first cycle are represented there by the Kasegalik Formation (Unit 1; Figs. 14, 15).

Kasegalik Formation (Unit 1), Pachi Arkoses: The Kasegalik Formation is most extensively exposed between the east and west arms of Kasegalik Lake after which it is named. One of the best exposed sections occurs along the axis of the Kasegalik Lake anticline at about 56°10' north latitude and 79°20' west longitude and is taken as the type section. The Kasegalik Formation (Unit 1) is also exposed on several islands in Churchill Sound and information from the type section is augmented by data obtained from these islands as well as elsewhere along the Kasegalik Lake anticline. The formation probably underlies a drift-covered depression in the centre of Tukarak Island on the east side of the Belcher Islands (Jackson, 1960).

About 4,000 feet of Kasegalik (Unit 1) strata are exposed along cores of the Kasegalik Lake and Churchill Sound anticlines (Jackson, 1960).

The predominant lithologies of the Kasegalik Formation are thin- to thick-bedded, aphanitic to finely crystalline, pink to grey dolomites and siliceous dolomites (Figs. 14, 15; Table VII). The lowest exposed strata are dolomite and light-grey medium-grained dolomitic quartzite. Red calcareous argillite is interbedded with pink dolomite in sequences up to 15 feet thick near the base of the formation.

A few beds of tuff, up to four feet thick, are interstratified with a few feet of thinly interbedded red shale and limestone near the top of the Kasegalik Formation which is marked in places by up to 15 feet of green chloritic argillite.

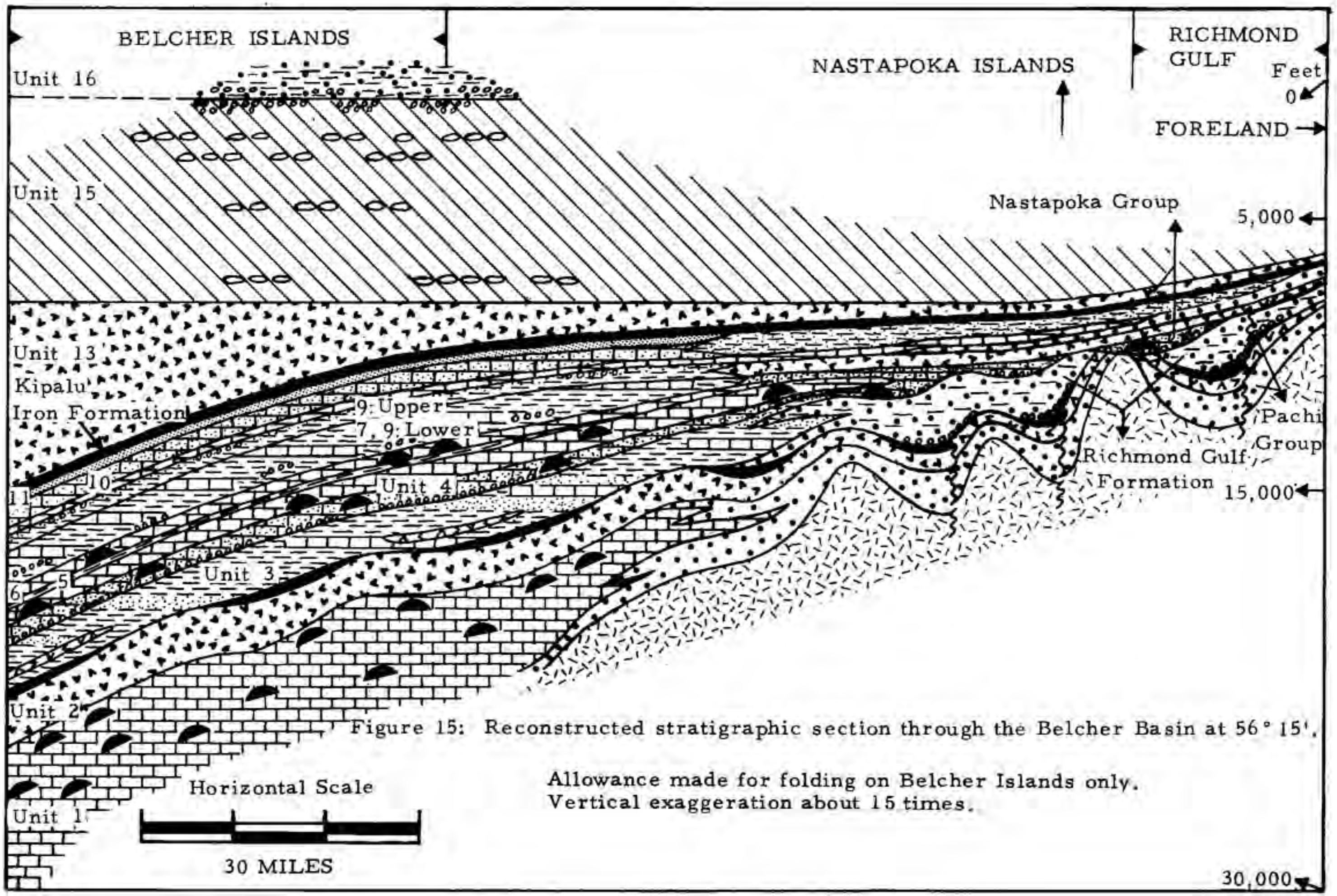


Figure 15: Reconstructed stratigraphic section through the Belcher Basin at 56° 15'.

Allowance made for folding on Belcher Islands only.
Vertical exaggeration about 15 times.

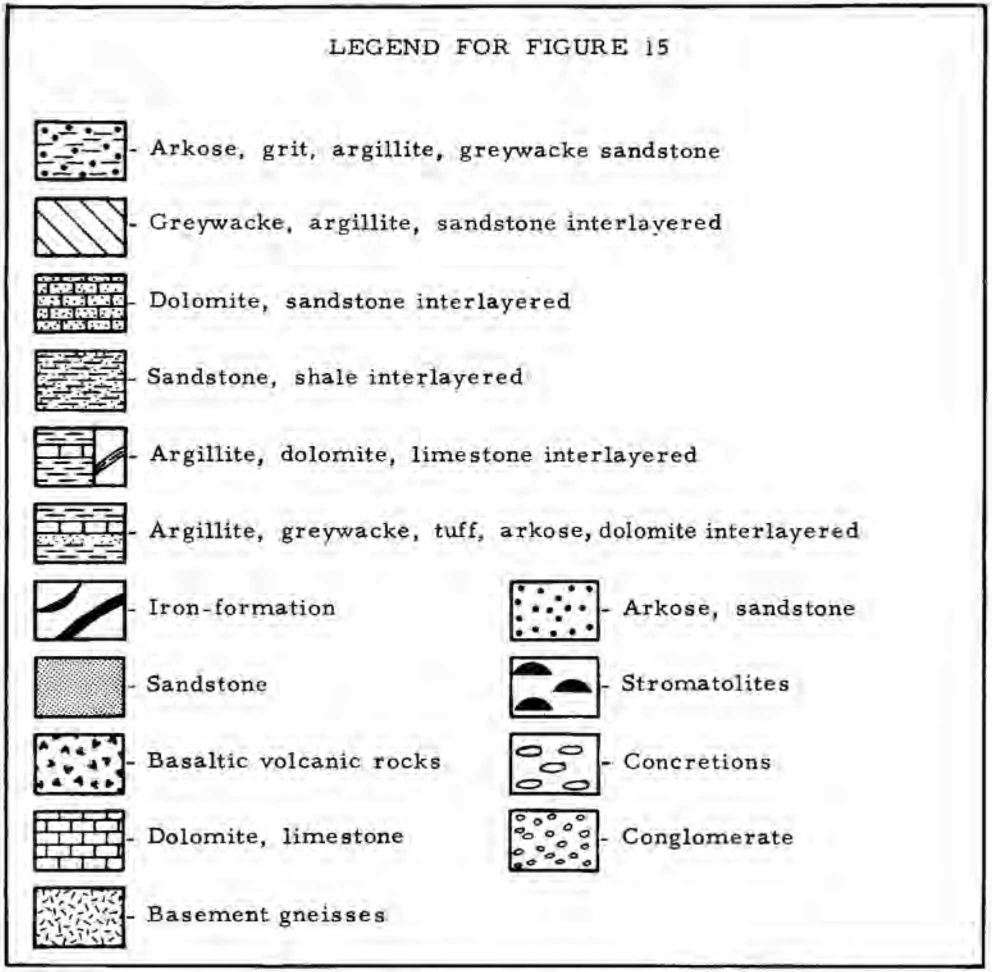


Figure 15. Reconstructed stratigraphic section through the Belcher Basin at 56°15'N.

TABLE VI: SUMMARY OF NOMENCLATURES USED IN RICHMOND GULF AREA

Bell, 1879	Low, 1902 1903	Leith, 1910 Young, 1922	Kranck, 1951	Woodcock, 1960	Eade, 1966 Stevenson, 1968	this paper	
Manitounuck Group	Nastapoka Group and Nastapoka Series	Nastapoka Group	Nastapoka Series	Nastapoka Group	Manitounuk Group	Manitounuk Supergroup	Nastapoka Group
	Manitounuk Series	Unnamed					
Unconformity	Thrust	Unconformity		Unconformity			Unconformity
	Unnamed			Richmond Gulf Formation			Richmond Gulf Formation
	Unconformity			Unconformity			Unconformity
	Unnamed	Richmond Group		Pachi Group			Pachi Group

Stromatolite-bearing zones are abundant throughout the formation and are restricted to grey coloured beds. They range in size and shape from small bun-, horn- and columnar-shaped types less than an inch long, to spherically shaped bodies a foot or more in diameter, and large dome-shaped bodies up to 12 feet across. Some of the latter types, about 3 feet across are associated with small siliceous horn-shaped stromatolities that stick out like spines perpendicularly to the domal layering. Grey to black chert is present in most stromatolite zones and microfossils have been reported from one of these zones (Hofmann and Jackson, 1969).

Cross-bedding is seen occasionally, mainly in the quartzitic beds. A few small local faults and unconformities are probably slump features. Kasegalik strata are brecciated, deformed, and locally slickensided and cross-faulted adjacent to the axial planes of the Kasegalik Lake and Churchill Sound anticlines, but there has probably been little displacement as faults have not breached the noses of the doubly plunging Kasegalik Lake anticline (Jackson, 1960).

The base of the Kasegalik Formation is not exposed, and its total thickness is therefore unknown. The top of the formation is conformable, and perhaps gradational with overlying green and red argillites at the base of the Eskimo Formation.

In the Richmond Gulf area 500 to 1,000 feet of pink to light grey arkose, quartzite, grit and pebble conglomerate unconformably overlie the Archean basement gneisses and represent the lower formation of the Pachi Group (Tables VI, VII; Low, 1902; Woodcock, 1960). These strata have been tentatively correlated with the strata of the Kasegalik Formation (Unit 1) of the Belcher Group (Fig. 15), and probable are a near-shore facies equivalent.

The Second Cycle

The second cycle is well-represented only on the Belcher Islands and in the Richmond Gulf area (Table VII). At these places the second cycle was initiated abruptly by deposition of basic volcanic strata, conformably on strata of the first cycle. On the Belcher Islands the second cycle includes all of the strata from the base of the Eskimo Formation to the top of the Kipalu Iron Formation (Units 2-12). At Richmond Gulf it includes all of the strata from the base of the Pachi volcanics to the top of the iron-formation on the Nastapoka Islands (Table VII).

Eskimo Formation (Unit 2), Pachi Volcanics: On the Belcher Islands the second cycle was initiated abruptly by the volcanism represented by the Eskimo Formation, which is 3,000 feet thick along the northern part of the Kasegalik Lake anticline where it is well exposed, from Eskimo Harbour, after which it is named, southward, to Kasegalik Lake. One of the best sections occurs along the western limb of the Kasegalik Lake anticline, just north of the west end of Windy Lake, about 4 miles south of Eskimo Harbour. Much of the information however was obtained from many localities along the Kasegalik Lake anticline and from Tukarak Island. The formation is about 2,000 feet thick in the centre of Tukarak Island to the east, about 1,000 feet thick in the northern Churchill Sound area to the west, and may be absent from the Robertson Bay and southern Churchill Sound areas to the southwest. Thus the Eskimo Formation is somewhat wedge-shaped, thinning westward and southward, and thickening north-northeastward.

The main rock types of the Eskimo Formation are aphanitic to fine-grained, massive, dark green to grey spilitic basalt, in part amygdaloidal, variolitic, and locally pillowed; and medium to coarse-grained feldspar porphyry and gabbro. At least a dozen different flows are represented in the section south of Eskimo Harbour. Megascopically the flows have an "older" or more altered appearance than those of the Kugong Formation (Unit 13). In thin-section, however, there seems to be little significant difference between them, although epidote is less common in the Eskimo flows.

TABLE VII: TENTATIVE CORRELATIONS IN THE BELCHER FOLD BELT

Sutton Lake (Bostock, 1969, in press)	Long Is.-Manitounuk Sd. (Bell, 1879; Low, 1902; Young, 1922)	Belcher Islands (Jackson, 1960; Hofmann and Jackson, 1969; Moore, 1918; Young, 1922)
		Unit 16: Red ark., qte., arg., cgl.; minor dol., dol. cgl.
		Unit 15: Gyw., sub gyw., arg.; minor tuff, qte., ark., dol., cgl; concretions mainly in upper pt.
		Disconformity?
Unit 4: Db.-gb.	Ba., db.-gb.	Unit 13: Spilitic ba. and feld. porph. flows, tuff, agg., volc. bx.; minor arg., ls., hfls., qtz. latite, gb.-db. dykes and sills.
U-Unit 3 (Iron Fm): Iron-rich chty. iron cbt., ch., arg., granular jasper, qte., gyw., slts.	Iron Fm.: Iron-rich arg., chty. iron cbt., ch.	Unit 12: (Kipalu I.F.): Vari-coloured iron-rich arg., qte., oolitic jasper, and chty. iron cbt.; sl., tuff at top.
Unit 3, Middle: Qte., sl.	Qte.	Unit 11: Qte., feld. qte., minor gyw., ch., arg., tuff., dol. cgl.
Unit 3, Lower: Ch. bx.-cgl., qte.	Dol., ls., ch., strom.?	Unit 10: U. mem.: Inbd. and cross-bd. qte., dol., qtzose. dol. L. mem.: Dol.; minor qte., dol. cgl.
? ? Unconformity	Ba., db.-gb. Gyw., sl., arg.	Unit 9: U. mem.: Varicoloured thinly inbd. arg.; minor qte., ls., dol., dol. cgl. L. Mem.: Thinly inbd. red arg., ls., dol.
Unit 2: Inbd. dol., sil. dol., arg., Some lumpy-bd.	Inbd. ls., dol., ss., qte.	Unit 7: Varicoloured thin, lumpy-bd. ls., dol., arg.; minor ch. bx.; bk. sl. at base.
Unit 2: Strom. dol., sil. dol., arg., ls., ch.	Ss., ls., minor shaly ls., some strom.	Unit 6: Strom. dol.; minor arg.
	Qte.; minor ls. cgl.	Unit 5: U. mem.: Thinly inbd. red arg., dol., ls., L. mem.: Vari-coloured inbd. arg., qte., ch., dol.
	Ls., dol., ch., some strom.	Unit 4: U. mem.: Strom. dol., minor ch. M. mem.: Sil. dol., strom. dol., qte. L. mem.: Dol., qtzose. dol., qte., cgl., few strom.
		Unit 3: Varicoloured arg.; gyw., dol.; qte. abundant toward top; tuff, ark., ba. near base.
		Disconformity?
		Unit 2: Spilitic ba. and feld. porph. flows, arg., gb-db.; minor tuff, agg., ark., qtz. latite; thin iron fm. locally at top.
		Unit 1: Strom. dol., dol., sil. dol., ch.; minor ls., sl., dolic. qte., tuff., green arg. at top, red arg. inbd. with dol. near base.
ARCHEAN BASEMENT	ARCHEAN BASEMENT	

TABLE VII: CONTINUED

Richmond Gulf (Bell,1879; Low,1898, 1902, 1903; Parks,1949; Woodcock,1960)			Hopewell Is. (Bell,1879; Low,1902, Lee,1965)	Cycle
Nastapoka Group:				
Flint, Belanger and Ross Is.	Anderson Is.-Daviean Is.	Nicholson Is.-Cotter Is.		3
		Ba., db.-gb.	Ba., db.-gb. Inbd. ss., sl.; qte., sh., ls.	
Iron Fm.: Iron-rich chty. iron cbt.,arg., sh., ch., tuff, ba.	Iron Fm.: Iron-rich chty. iron cbt., ch.,granular jasper, purple gyw.-sh. at base	Iron Fm.: Iron rich chty. iron cbt.,ch., sh.; gyw.-sh. at base	Iron Fm.: Iron-rich sh., chty. iron cbt.	2
Ark.,ba.-db., ss., qte., sl., sh. Inbd. ss., sh. Ls., dol., strom.?	Inbd. ss., sh.	Ss., gyw., sh., ls., calc. ss., feld. ss.	Qte.; inbd. ss., sl.; ch., chty. iron cbt.	
Ark., feld. qte				
Spilitic ba., volc. bx., tuff, ch., db.-gb.		Ba-db. ?		
Qte., ss., ark., arg., sh.; minor ls., dol.		Sh., ss.; minor ls., ? strom.?		
Dol., ls., ch., strom. dol., cgl.; gal., sphal. mineraliz'n.				
Ss., cal. ss., sh., ark., grit., cgl., ch. bx.				
Dol., ls.; minor sh., strom. dol.; sil. dol.; basal bx.				
Unconformity				
Richmond Gulf Fm.: Ark., grit, feld. ss.; minor arg., gyw.; pebble-boulder cgl.				
Unconformity				
Pachi Gp. Volcs.: Inbd. and - ba., ark. These overlain locally by iron fm. of iron-rich gyw., arg., s ., granular jasper				
Pachi Gp. Seds.: Ark., feld. qte., grit, pebble cgl.				1
Unconformity				
ARCHEAN BASEMENT GNEISSES			ARCHEAN BASEMENT	

A variety of rock types are interbedded with the igneous rocks, mainly in the upper part of the formation (Table VII). Interbedded green and red argillites comprise sequences up to 200 feet thick at several stratigraphic horizons within the Eskimo Formation (Unit 2), and are most abundant in the northern Kasegalik Lake-Eskimo Harbour area. A thin zone of these argillites represents the lowest strata of the formation exposed along much of the Kasegalik Lake anticline, although relations are obscured by the presence of one or more feldspar porphyry gabbro sills.

About 3 feet of lean thin-bedded granular, jasper iron-formation occurs at the top of the Eskimo Formation on a small island at the north end of Churchill Sound where it is associated with 12 feet of green and red argillites. This is the only observed occurrence of this iron-formation, but a zone of aeromagnetic anomalies is associated with the same stratigraphic level along the Churchill Sound and Kasegalik Lake anticlines (Belcher Mining Corp. Ltd., Company Rept.). The anomalies are similar to those produced by the Kipalu Iron Formation (Unit 12), but the upper contact of the Eskimo Formation is mostly drift-covered and the source of the anomalies is unknown. They may be attributable to a continuation of the iron-formation described above. Thus a lean iron-formation, possibly a ferruginous argillite, may be present at or near the top of the Eskimo Formation in this part of the Belcher Islands.

Both the upper and lower contact areas of the Eskimo Formation are poorly exposed, but information from many localities indicates that both contacts are probably conformable and gradational. The thinning of the Eskimo Formation southward and westward probably reflects decreasing volcanism in these directions, although the presence of a disconformity at the top of the formation cannot be completely discounted.

At Richmond Gulf the volcanic-bearing formation of the Pachi Group overlies the lower arkoses conformably and is about 350 feet thick (Fig. 15, Table VII). It comprises four or five dark green to grey propylitized andesite (Woodcock, 1960) or basalt flows up to 60 feet thick, interbedded with grey, pink and green beds and lenses of arkose. These may be near-shore equivalents of the argillite sequences of the Eskimo Formation on the Belcher Islands. The flows are more altered than those of the Eskimo Formation. The postulated distribution of these volcanic rocks is shown by the dotted line in Figure 13.

Lean iron-formation is reported to overlie the Pachi volcanics in two sections measured along the south side of Richmond Gulf by Low (1902). It is respectively 250 feet and possibly 550 feet thick in the two sections and is composed of ferruginous jasper, argillite and greywacke with a few interbeds of sandstone and arkose.

<u>Fairweather Formation (Unit 3), Richmond Gulf Formation</u>: The Fairweather Formation derives its name from the Fairweather Sound area. An auxiliary section is exposed on the broad point west of the north end of Mavor Island. The type section lies 3 miles west by south of Laddie Harbour. The formation is 1,650 feet thick on Tukarak Island, about 1,800 feet thick on Moore Island, and ranges from 1,200 to 2,000 feet thick along the Kasegalik Lake anticline (Fig. 14).

The Fairweather Formation consists mainly of varicoloured argillite with interbedded greywacke in the lower part and quartzite in the upper (Table VII). Red colours occur almost exclusively in the lower part and a few amygdaloidal basalt flows lie near the base of the formation. On Tukarak Island the upper part changes within 5 miles along strike from mainly argillite to mainly quartzite. Beds of dolomite as much as 130 feet thick, occur at several widely spaced stratigraphic levels. A white to grey quartzite bed is taken as the top of the formation.

Some layers contain graded bedding and cross-bedding is present in a few places. The bottom part of the formation is characteristically poorly exposed but the formation seems conformable and gradational with the underlying Eskimo Formation. The upper contact is conformable and gradational with the overlying McLeary Formation.

The Richmond Gulf Formation (Fig. 15, Table VII) is interpreted as the near-shore facies equivalent of the Fairweather Formation, although an unconformity lies between it and the underlying Pachi Group (Low, 1902; Woodcock, 1960). Its thickness is variable and ranges from 500 to 1,500 feet. It is missing in some places, either by erosion or by depositional onlap. The Richmond Gulf Formation is composed mainly of varicoloured arkoses and coarse grits. Sedimentary structures such as cross-bedding, ripple marks and swash marks are common. In some places erosion of the spherical structures from the underlying Pachi volcanics has produced a "basal boulder conglomerate with an arkose-matrix" (Woodcock, 1960).

The Richmond Gulf Formation lies unconformably between the underlying Pachi Group and the overlying Nastapoka Group. According to Woodcock (1960) the Richmond Gulf Formation was deposited in pre-existing valleys in the Pachi Group. The Richmond Gulf strata were subsequently uplifted, tilted to the southwest, cut by numerous small faults and then bevelled, prior to deposition of the Nastapoka Group (Fig. 15).

<u>McLeary Formation (Unit 4), and Equivalents</u>: The McLeary Formation is named after McLeary Point on the east side of Tukarak Island. The type section lies 3 miles west of the Point and 3,000 feet northeast of the north end of Fairweather Sound. An auxiliary section is well exposed 2 miles west of Laddie Harbour.

The formation is 1,430 feet thick west of McLeary Point and is well exposed along the Tukarak Island anticline. It is about 1,200 feet thick along the northern part of the Kasegalik Lake anticline and along the east side of Moore Island.

The type section of the McLeary Formation has been divided into three members, a division that holds generally throughout the Belcher Islands (Fig. 14, Table VII). The lower member, 680 feet thick, is composed of thin- to thick-bedded light grey and pink dolomite with a few stromatolitic zones, arenaceous dolomite, and white to light grey orthoquartzite. The quartzite occurs mainly in zones up to 30 feet thick. In the middle of the member traces of chalcopyrite are associated with dark chert lenses and blobs to 4 inches in diameter.

The middle member, 290 feet thick, is mainly a pale-pink, aphanitic, siliceous dolomite with minor interbeds of orthoquartzite and grey stromatolitic dolomite. The upper member, 460 feet thick, is mainly a laminated to thin-bedded, grey, finely crystalline dolomite with a few chert beds and disseminated quartz grains. Several beds contain stromatolites up to 3 feet in diameter.

Stromatolites are most abundant in the upper member and several types are represented in the formation. They are:

(a) Small cherty columnar stromatolites about 1 inch long.

(b) Columnar or digitate varieties up to 1 foot long.

(c) Orange-weathering bun-shaped stromatolites 1-3 inches across at the top of Unit 4 in East Belcher Islands.

(d) Spherical types about 1/2-2 feet in diameter.

(e) Broad dome-shaped stromatolites up to 3 feet across.

Cross-bedding is present in some of the quartzite horizons. Throughout most of the Belcher Islands a bed of dolomitic quartzite and dolomite-conglomerate forms the base of this formation and overlies the Fairweather Formation conformably. The top of the formation is conformable with the overlying Tukarak Formation and on Tukarak and Mavor Islands coincides with a grey, orange-weathering dolomite bed containing stromatolites one to three inches in diameter.

Up to 30 feet of dolomite-cemented arkose breccia overlies the Richmond Gulf Formation unconformably and is the basal member of the Nastapoka Group at Richmond Gulf (Woodcock, 1960). The breccia is overlain by a stromatolite-bearing dolomite limestone unit, 155-235 feet thick. In the Long Island-Manitounuk Sound area (Table VII, Fig. 13) at least 400 feet of stromatolitic dolomite and limestone comprise the basal unit. Both of these carbonate sequences are probably equivalent, in part at least, to the McLeary Formation.

Tukarak Formation (Unit 5) and Equivalents: The Tukarak Formation, a good marker horizon, is well exposed on Tukarak Island after which it is named. The type section lies about 14,000 feet west of McLeary Point where it is 230 feet thick. It is about 200 feet thick on the west side of Tukarak Island, but may be only 150 feet thick on Mavor Island, and even thinner along the northern part of the Kasegalik Lake anticline.

At the type section the lower member is 100 feet thick but is less on Mavor Island. It consists of thinly interbedded varicoloured argillite, quartzite, dolomite and minor chert. The upper member, 130 feet thick, contains interbedded red argillite, dolomite, and limestone in beds mostly two inches thick or less. Weathering of the carbonate beds produces a corrugated, lumpy-bedded, appearance.

The lower contact of the formation is conformable and gradational with the McLeary Formation whereas the upper contact holds the same relationships with the Mavor Formation.

In the Richmond Gulf area the same relative stratigraphic position (Table VII) is occupied by a sequence, 40-400 feet thick, of sandstones with minor calcareous sandstone, shale, and conglomerate (Low, 1902). At Manitounuk Sound, 443 feet of bluish grey quartzite occupies a similar stratigraphic position (Low, 1902). Both units form parts of conformable sequences.

Mavor Formation (Unit 6) and Equivalents: The Mavor Formation is named after Mavor Island where it is well exposed and where information was obtained to supplement that from the type section located about 2 miles southwest from the southwest corner of Laddie Harbour. The formation is 450 feet thick in the type section and may be over 550 feet thick on Mavor Island. It is about 300 feet thick on the west side of Tukarak Island and along the northern part of the Kasegalik Lake anticline (Fig. 14).

The formation is composed largely of thick-bedded, grey, pink and tan, aphanitic to finely crystalline dolomite, with thin argillite partings. The strata are more thinly bedded toward the top of the formation, and at the top a few interbeds of grey dolomitic argillite are present. Stromatolites are restricted to the more coarsely crystalline grey dolomite beds, and are mostly spherical types two to eight inches in diameter. The lower contact, between this and the Tukarak Formation is gradational through a thickness of about three feet. The upper contact with the Costello Formation is also gradational.

The same relative stratigraphic position (Table VII) in the Richmond Gulf area is occupied by a dolomite-limestone unit that is 110-334 feet thick within the Nastapoka Group (Bell, 1879; Low, 1902). Other lithologies include cherty vuggy limestone containing galena and sphalerite, siliceous limestone and stromatolitic dolomite. One section that includes 755 feet of carbonate strata, probably includes facies equivalents of the underlying sandstone unit in nearby sections.

At Manitounuk Sound, a 283-foot section composed of sandstones with shaly limestone partings and bluish grey quartzite may be stratigraphically equivalent to the Mavor Formation (Table VII). Three limestone sequences comprise one-third of the section; the thickest, 52 feet thick, occurs at the top of the section (Low, 1902). Southeast of Long Island a stromatolitic carbonate sequence described by Low (1902) may be equivalent to either the Mavor (Unit 6) or the McLeary (Unit 4) Formations on the Belcher Islands.

About 250 feet or more of thin- to thick-bedded stromatolitic dolomite comprises the lower part of the Nowashe Formation (Unit 2) at the base of the Aphebian section in the Sutton Lake region (Bostock, in press). Minor lithologies include chert, argillite and limestone. These strata are tentatively correlated with the Mavor Formation on the Belcher Islands.

Costello Formation (Unit 7) and Equivalents: The Costello Formation is exposed discontinuously along the south shore of Costello Lake, Tukarak Island, after which it is named. The type section lies about 9,000 feet south by west from the southwest corner of Laddie Harbour.

The formation is 1,205 feet thick in the type section; and is about 1,000 feet thick on Mavor Island, 780 feet thick on the west side of Tukarak Island and 1,200 feet thick near the north end of Kipalu Inlet (Fig. 14). It is poorly exposed along the northern part of the Kasegalik Lake anticline.

The basal member, 85-115 feet thick, is composed mainly of dark grey to black slate and argillite, which grade upward into a sequence predominantly of alternating grey, green, red, white, or tan to brown zones, each 5-200 feet thick. The zones consist of alternating beds mostly less than two inches thick of limestone, dolomite, calcareous chert and argillite. A few beds of intraformational conglomerate and chert breccia are also present. Carbonate-rich beds tend to be discontinuous and lenticular and they weather with a corrugated or knotty, lumpy-bedded appearance.

The lower and upper contacts are gradational with the Mavor and Laddie Formations respectively.

A sequence composed predominantly of interbedded sandstone, quartzite, shale and arkose in the Richmond Gulf area (Bell, 1879; Low, 1902) is tentatively correlated with the Costello Formation and possibly with the lower member of the Laddie Formaton on the Belcher Islands. The sequence is 45 to 130 feet thick, shows numerous facies changes, and different lithologies predominate at different localities (Table VII). Thirty-six feet of greywacke and slate occupy this position at Manitounuk Sound (low, 1902) and 85 feet may be present on Cotter Island, the northernmost of the Nastapoka Islands (Low, 1903).

A thin sequence of thinly interbedded dolomite, siliceous dolomite and lenticular argillite outcrop near Aquatuk River in the Sutton Lake area. These are assigned by Bostock (in press) to his Nowashe Formation (Unit 2). Possibly these strata lie near the top of the Nowashe Formation, and they are correlated here with the Costello Formation on the Belcher Islands.

Laddie Formation (Unit 9) and Equivalents: The Laddie Formation is well exposed along the western shore of Laddie Harbour after which it is named. The type section lies about 6,500 feet south of the southwestern side of Laddie Harbour.

The formation is 1,085 feet thick in the type section and is well exposed on Tukarak Island. It may be about 1,500 feet thick at the north end of Kipalu Inlet (Fig. 14) and is exposed at the north end of Innetalling Island. It is mostly drift-covered elsewhere.

The lower member, 425 feet thick in the type section, is composed of alternating beds, commonly less than two inches thick, of red argillite, limestone, and dolomite. Carbonate beds decrease in number upward and include a few white beds.

The upper member is 660 feet thick. Its lower part is predominantly thinly interbedded varicoloured argillite that grades upward into green argillite. In the upper part of this member zones of green argillite alternate with zones, up to 75 feet thick, of thick-bedded quartzite, dolomitic quartzite and minor arenaceous dolomite conglomerate.

The lower contact is gradational with the Costello Formation; the upper contact is gradational through about three feet from quartzite to sandy chert, dolomite conglomerate and dolomite.

On the mainland of the Richmond Gulf area, a basic volcanic unit ("Cap Rock") overlies the rest of the Nastapoka Group strata conformably, or with only slight disconformity. Exposed thickness ranges from about 50 to 750 feet; but is about 400 feet in most places (Woodcock, 1960). Three to five individual flows are present in most places and the unit is composed mainly of altered amygdaloidal spilitic basalt and basalt breccia with thin interbeds of pyroclastics (Parks, 1949). A porphyritic diabase sill forms the base of the unit and is 106 feet thick at one locality (Parks, 1949). A few feet of arkose and feldspathic quartzite rest on the "Cap Rock" in a few places (Parks, 1949; Woodcock, 1960).

The volcanic unit ("Cap Rock") is correlated with the lower part of the upper member of the Laddie Formation on the Belcher Islands, and may also be equivalent in part to the lower member of the Laddie Formation. The thin arkose remnants overlying the "Cap Rock" may be stratigraphically equivalent to the thick-bedded quartzite zones in the upper part of the upper member of the Laddie Formation or to the lower member of the Rowatt Formation on the Belcher Islands. The postulated distribution of the "Cap Rock" volcanics is shown by the broad dashed lines in Figure 13.

At Manitounuk Sound (Table VII) 36 feet of greywacke and slate underlie the "Cap Rock" and are probably equivalent to the lower member of the Laddie Formation (Low, 1902). Farther southwest, at Long Island, sandstones appear to occupy the approximate stratigraphic position of the "Cap Rock" (Low, 1902). Stratigraphic equivalents of the Laddie Formation have not yet been recognized in the Sutton Lake area (Bostock, in press).

Rowatt Formation (Unit 10) and Equivalents: The Rowatt Formation is well exposed at the north end of Kipalu Inlet, about 4,000 feet west of Rowatt Harbour after which it is named. It is also well exposed on Tukarak Island and the type section lies 6,500 feet south of the south side of Laddie Harbour. Good exposures are present at the north end of Innetalling Island, and in a few places along the west side of Mukpollo Peninsula.

The formation is 900 feet thick in its type section and may be 1,150 feet west of Rowatt Harbour (Fig. 14). It has been divided into two members. The lower member, 280 feet thick south of Laddie Harbour may be only 150 feet thick west of Rowatt Harbour. It is composed mainly of thick-bedded to massive, light grey, tan and pink dolomite with minor quartzite and conglomerate (Table VII). The upper member is 620 feet thick in the type section and possibly 1,000 feet thick west of Rowatt Harbour. It consists of thin- to thick-bedded, grey, pink, and tan quartzite, dolomite, dolomitic quartzite and arenaceous dolomite, that are interbedded in alternating quartzitic and dolomitic zones, 18 to 260 feet thick.

Much of the quartzite is cross-bedded. Irregular double-walled quartz veinlets in some of the dolomite beds in the upper member and at the top of the lower member may represent mud-crack fillings. The lower contact is gradational with the Laddie Formation. The upper contact is conformable and in places gradational with the Mukpollo Formation.

About 50 feet of deformed dolomite and limestone on the east side of Belanger Island (Table VII) contain stromatolites 2 to 12 inches in diameter (Low, 1903) and the strata are probably equivalent to the upper part of the Rowatt Formation. Similar strata outcrop on Gull Islet (Bell, 1879). At least 150-200 feet of dolomite and limestone outcrop on Long Island (Low, 1902; Young, 1922) and are probably equivalent to the Rowatt Formation on the Belcher Islands.

Discontinuous lenses of chert breccia-conglomerate up to 85 feet thick rest unconformably on the Nowashe Formation (Unit 2) and comprise the lowest member (1) of the Sutton Ranges Formation (Unit 3) in the Sutton Lake area (Bostock, in press). Several varieties of chert are represented and fragments up to one foot in diameter occur in a sandy chert or argillaceous matrix. Lenses or beds of white quartzite up to seven feet thick occur in the breccia and a few slate lenses are also present. The breccia is correlated with the Rowatt Formation on the Belcher Islands.

Mukpollo Formation (Unit 11) and Equivalents: This formation is exposed at several places along the west side of Mukpollo Peninsula after which it is named. The type section lies about 6,500 feet south of Laddie Harbour where about 360 feet of strata are exposed. The formation is generally poorly exposed, and may range from 130 feet in thickness north of Little Costello Lake to at least 475 feet northeast of the north end of Moore Island.

Most of the rock is a thick-bedded to massive, white to light grey and pink orthoquartzite and feldspathic quartzite (Table VII). Individual beds may be up to twelve feet thick. Cross-bedding is present in a few places. The lower contact, with the Rowatt Formation, is conformable. The upper contact, with the Kipalu Iron Formation, is rarely exposed but seems conformable.

On the Nastapoka Islands the iron-formation overlies interbedded greenish grey sandstones and shales. Quartzite, arkose, slate and basalt or diabase are also present at various localities, mainly in the upper part of the sequence. Sandstone or shale predominates as the major lithology in alternate zones up to 220 feet thick. The thickest sections occur on Belanger and Ross Islands where about 300 feet of these strata are exposed. They are tentatively correlated with the Mukpollo Formation.

The iron-formation on Long Island overlies at least 60 feet of quartzite (Young, 1922). At least eight feet of light-coloured quartzite and slate appear to lie conformably between the underlying chert breccia member and the overlying iron-formation member of the Sutton Ranges Formation (Unit 3) in the Sutton Lake area (Bostock, in press).

Kipalu Iron Formation (Unit 12) and Equivalents: Moore (1918) named this unit the Keepalloo Formation because of the excellent exposures along Keepalloo Sound or Kipalu Inlet as it is called now. The Belcher Mining Corporation has used the name Kipalu Iron Formation in its company reports and this usage was followed by Jackson (1960). The formation has been described in detail by Young (1922).

The formation, although widespread on the Belcher Islands is commonly poorly exposed. It is well exposed in the eastern part of the Belcher Islands on Tukarak and Innetalling Islands, along Kipalu Inlet, and between Haig Inlet and Eskimo Harbour. Some exposed and probable thicknesses are tabulated below.

Location	Exposed Thickness (ft.)	Probable Thickness (ft.)
6,500' South Laddie Harbour	245	350
Spracklin Pt. (Innetalling Is.)	380	400
Kipalu Inlet (W. of Walton Is.)	200	?
Kipalu Inlet (W. of Gilmour Pen.)	335	370
1.7 mi. N. of Haig Inlet	335	410
S.E. corner Eskimo Harbour	210	410
Howard Peninsula (N.E. Moore Is.)	170	400
S.W. end of Kugong Is.	180	?

The sections south of Laddie Harbour, at Spracklin Point, west of Gilmour Peninsula, and north of Haig Inlet may be considered as auxiliary type sections.

Most of the formation is dense, aphanitic to very fine grained, red to reddish brown and laminated to thin-bedded with a few beds up to five feet thick. The iron occurs mainly as hematite, although considerable magnetite is present locally. Megascopically most beds resemble ferruginous argillite and slaty argillite, they are commonly granular, siliceous, and contain small amounts of carbonate. The strata contain varying amounts of chert, jasper, hematite, magnetite, iron carbonates, chlorite, greenalite, stilpnomelane, minnesotaite, riebeckite, biotite, muscovite, graphite, and detrital quartz plagioclase, microcline, and argillaceous material. Thus the so-called argillites are probably composed of varying amounts of clastic and chemically deposited material, considerable pyroclastic material is probably present in many beds. Most of the iron-formation is rather lean, but hematite- and magnetite-rich zones north of Haig Inlet and at the north end of Innetalling Island range up to 30 feet in thickness and contain 35 to 45 per cent metallic iron (Young, 1922; Jackson, 1960). Individual beds up to two feet thick contain about 50 per cent iron.

A few beds, composed of chert with iron carbonate and chert or jasper with iron oxides, occur mainly in the lower part of the formation. At Fairweather Harbour and elsewhere, however, coarse granular jasper-magnetite-hematite rock comprises 62-128 feet of the upper part of the formation. Ferruginous feldspathic quartzite is abundant south of Laddie Harbour. The top of the Kipalu Formation is a dark coloured chert zone 25-40 feet thick containing brown iron carbonate.

The contact with the underlying Mukpollo Formation is conformable and commonly sharp, although in some places it is gradational. The top of the iron-formation is commonly in contact with a gabbro sill. In a few places, however, as on Kugong Island and north of Haig Inlet, the chert zone at the top of the formation grades upward into a slightly ferruginous dark slate-argillite-greywacke zone up to 30 feet or more thick, that contains beds of tuff, agglomerate and iron carbonate, and which in turn grades upward into pyroclastics that are in contact with a gabbro sill. The slate-argillite zone has been included in the Flaherty Formation for expediency (Jackson, 1960) but might more properly be part of the Kipalu Formation.

The Kipalu Iron Formation is probably related, in part at least to the oncoming volcanism of Unit 13 that initiated the third cycle and which was probably already underway in an adjacent area. Iron-formation occurs with the volcanic strata of Unit 13 on the Sleeper Islands 30 miles north of the North Belcher Islands (J. Kilualuk, pers. comm.).

The iron-formation on the Nastapoka Islands is part of a conformable sequence. It is very similar to the Kipalu Iron Formation with which it is correlated on the basis of the similar stratigraphic positions and of the remarkably close lithological and stratigraphic similarities of the iron-formations themselves. On Flint and Belanger Islands (Table VII) the iron-formation overlies 15-25 feet of basalt-diabase. To the north this stratigraphic position is occupied by 10-85 feet of ferruginous, purple and dark green greywacke-shale. Up to 15 feet of lean jasper is commonly present either above or below the purple shale. On the southern and northern Nastapoka Islands the middle part of the iron-formation is composed mainly of 75 to 185 feet of ferruginous siliceous shale and argillite. From Anderson to Davieau Island this shale-argillite sequence is replaced largely by 100 to 220 feet of ferruginous chert and granular jasper. The jasper zones attain their greatest thickness of about 110 feet between Clarke and Taylor Islands. The top unit of the iron-formation is 20 to 60 feet of ankerite-bearing chert.

The iron-formation and the Hopewell Islands (Table VII) is composed of about 10 feet of lean chert and jasper including an iron-carbonate-rich bed 1/2 to 2 feet thick (Lee, 1965). The iron-formation overlies 80 to 135 feet of interbedded sandstone, quartzite and slate, and it is overlain by 50 to 90 feet of similar strata (Table VII). A maximum of 240 feet of conformable sedimentary strata are exposed, and are probably equivalent to both the Mukpollo and Kipalu Formations on the Belcher Islands.

At Long Island (Table VII) iron-formation 340 feet thick is very similar to the Kipalu Iron Formation and is probably stratigraphically equivalent to it. At least 20 feet of iron carbonate-bearing chert outcrops at the top of the Long Island iron-formation.

An iron-formation 110 feet or more thick (Table VII) comprises the upper three members of the Sutton Ranges Formation (Unit 3) in the Sutton Lakes area (Hawley, 1926; Bostock, in press). The lower member has a maximum exposed thickness of 230 feet and is composed of slightly ferruginous greywacke, siltstone and argillite. About 35 feet of jaspillitic, cherty, and slaty iron-formation comprise the middle member which contains iron silicate and iron carbonate minerals. Jaspillitic and slaty iron-formation predominate in the upper member which is 60 feet thick. This unit is probably equivalent to the Kipalu Iron Formation. The lower member at Sutton Lake bears a closer resemblance to the lower iron-formation of the Nastapoka Islands than to the lower Kipalu Iron Formation.

The Third Cycle

Like the second, the third cycle on the Belcher Islands (Units 13-16) was initiated with little warning by a period of volcanism which seems to have terminated more abruptly than the first volcanism. Also, whereas the Eskimo Formation thickens toward the north-northeast, the Flaherty Formation thickens westward.

Flaherty Formation (Unit 13) and Equivalents: Although the Flaherty Formation is well exposed throughout the Belcher Islands the relationships between it and the stratigraphically adjacent formations (Fig. 14, Table VII) are best exposed on Flaherty Island after which the formation is named. No single type section is proposed because the stratigraphy within the formation is variable. The section 6,500 feet south of Laddie Harbour is representative of the formation's stratigraphy on Tukarak and Innetalling Islands. The sections west of Gilmour Peninsula, south of Kasegalik Lake, west and south of Mani Lake, east and west of the north end of Moore Island and on Johnson Island are similar in general make-up.

On the east side of Tukarak Island south of Laddie Harbour the formation is about 960 feet thick (Fig. 14, Table VII) and contains at least twelve flows, each up to 60 feet thick, and separated from one another by sedimentary-pyroclastic sequences up to 55 feet thick. On the west side of the Belcher Islands the Flaherty Formation is about 6,400 feet thick. There are at least twice as many flows in the western part of the Islands and individual flows may attain 400 feet in thickness. Intervening sedimentary-pyroclastic sequences are up to 200 feet thick and have been traced for well over 16 miles along strike. In addition they contain a higher proportion of pyroclastic material to sedimentary material than in the eastern part of the Islands. Some gabbro-diabase sills occur mainly in the lower part of the formation and may attain thicknesses of over 400 feet. Elsewhere the formation is 1,600 feet thick west of Gilmour Peninsula, at least 3,000 feet thick east of Kasegalik Lake and at least 5,500 feet thick south of the lake (Fig. 14).

The Flaherty Formation consists mainly of aphanitic to fine-grained amygdaloidal massive and pillowed spilitic basalt (Schenk, 1959). Volcanic breccia, ropy and blocky basalt, and a few medium- to coarse-grained massive to pillowed feldspar porphyry flows are also present. The main constituents are plagioclase, diopside, chlorite, epidote, actinolite, muscovite, calcite, and minor iron oxides and olivine. Quartz and calcite fill most of the voids between the pillows and occupy the cores of a few pillows, a feature observed also on the Sleeper Islands (Manning, 1947). Anthraxolite is abundant locally as a filling between pillows. Argillite, black slate, greywacke, limestone, chert and hornfels occur interbedded with the pyroclastic strata (Fig. 14, Table VII).

In most places the basal volcanic strata are tuffs which are gradational with the underlying Kipalu Iron Formation. South of Kasegalik Lake, in the Rowatt Harbour area, and west of Mavi Lake, the top of the formation is in sharp, conformable contact with the overlying Omarolluk Formation.

Spilitic basalt and diabase cap the stratigraphic sections exposed on the four northernmost Nastapoka Islands and on the Hopewell Islands (Bell, 1879; Low, 1903; Lee, 1965). These are correlated with the Flaherty Formation

(Table VII). A maximum thickness of 100 feet is exposed and includes three flows, the thickest of which is 50 feet. Basalt and diabase also overlie the iron-formation on Long Island (Bell, 1879; Low, 1902; Young, 1922), and one or more diabase sills are associated with the iron-formation of Sutton Lake (Bostock, in press; Dowling, 1905; Hawley, 1926). These are also correlated with the Flaherty Formation. The stratigraphic position of the basalt and diabase, and the associated sediments, present in the two drill holes west of James Bay (*see* General Features, Belcher Fold Belt in this paper) is uncertain.

Pillowed basalts are exposed on practically all of a chain of islands that extends from the Belcher Islands north to the Ottawa Islands (Bell, 1885; Manning, 1947). The islands include the Sleeper, Marcopeet, Farmer, and Two Brothers Islands, and it seems likely that most if not all of the volcanic strata are stratigraphically equivalent to the Flaherty Formation on the Belcher Islands. A few pebbles and cobbles of a thinly bedded red argillite have been found on the Ottawa Islands (Andrews, J.T. and Falconer, G., 1969) and are identical in appearance to some of the Belcher Group strata. Serpentine is associated with the basalt on House Island (Manning, 1947), one of the southern Ottawa Islands. Patches of rhyolite have been reported to occur in basalts of the Sleeper Islands.

The Flaherty Formation on the Belcher Islands and the stratigraphically equivalent volcanic strata in the rest of the Belcher Fold Belt therefore, represent a major volcanic episode, and the related rocks now occupy a long, wide belt that extends from the Sutton Lake area and Long Island in the south to the Ottawa Islands and Cape Smith in the north (Fig. 13).

Haig Intrusions (Unit 14) and Equivalents: The intrusive relationships of the gabbro-diabase sills and dykes within the formations of the Belcher Group are well exposed in the general area around Haig Inlet after which these intrusions are named. Gabbro-diabase sills and dykes intrude all earlier formations on the Belcher Islands except, possibly, the upper part of the Flaherty Formation. Most of these basic intrusions are probably related to the volcanic activity represented by the Flaherty Formation, and to-date no basic intrusions, or extrusions have been identified in the Omarolluk (Unit 15) and Loaf Formations (Unit 16).

Some of the sills are differentiated and individual and composite sills range up to 900 feet in thickness. One remarkable sill is present at or adjacent to the contact between the Kipalu Iron Formation and the Flaherty Formation throughout the Belcher Islands, and is commonly less than 100 feet thick. Most of the sills are massive, aphanitic to medium grained diabase with a subophitic texture. There are some coarse-grained feldspar porphyry intrusions, and a few granitic and ultrabasic pockets are present in the basic intrusions of the eastern part of the Belcher Islands. Thin-section examination, however, shows that one of these "granitic" pockets is a thermally metamorphosed, partially ingested block of arkose that has probably been brought up, possibly from an arkose bed near the base of the Belcher Group.

Diabase sills are present between the iron-formation and overlying basalt on the Hopewell Islands (Lee, 1965) and at Long Island (Low, 1902). These sills and the diabase intrusions in the Sutton Lake area (Bostock, in press; Hawley, 1925) are probably related to the volcanic episode represented by the Flaherty Formation on the Belcher Islands. Most of the basic intrusions at Richmond Gulf, however (Woodcock, 1960) including the Wiachuan Sill between the Pachi volcanics and the overlying Richmond Gulf Formation may be related to the volcanic unit that in this area, is correlated here with the upper member of the Laddie Formation (Unit 9) on the Belcher Islands.

Late aphanitic trap dykes intrude Haig diabase at the south end of Churchill Sound on the Belcher Islands (Jackson, 1960). These have been dated at 881 m.y. (Jackson, 1967), and have recently been named as Unit 17 on the Belcher Islands (Hofmann and Jackson, 1969).

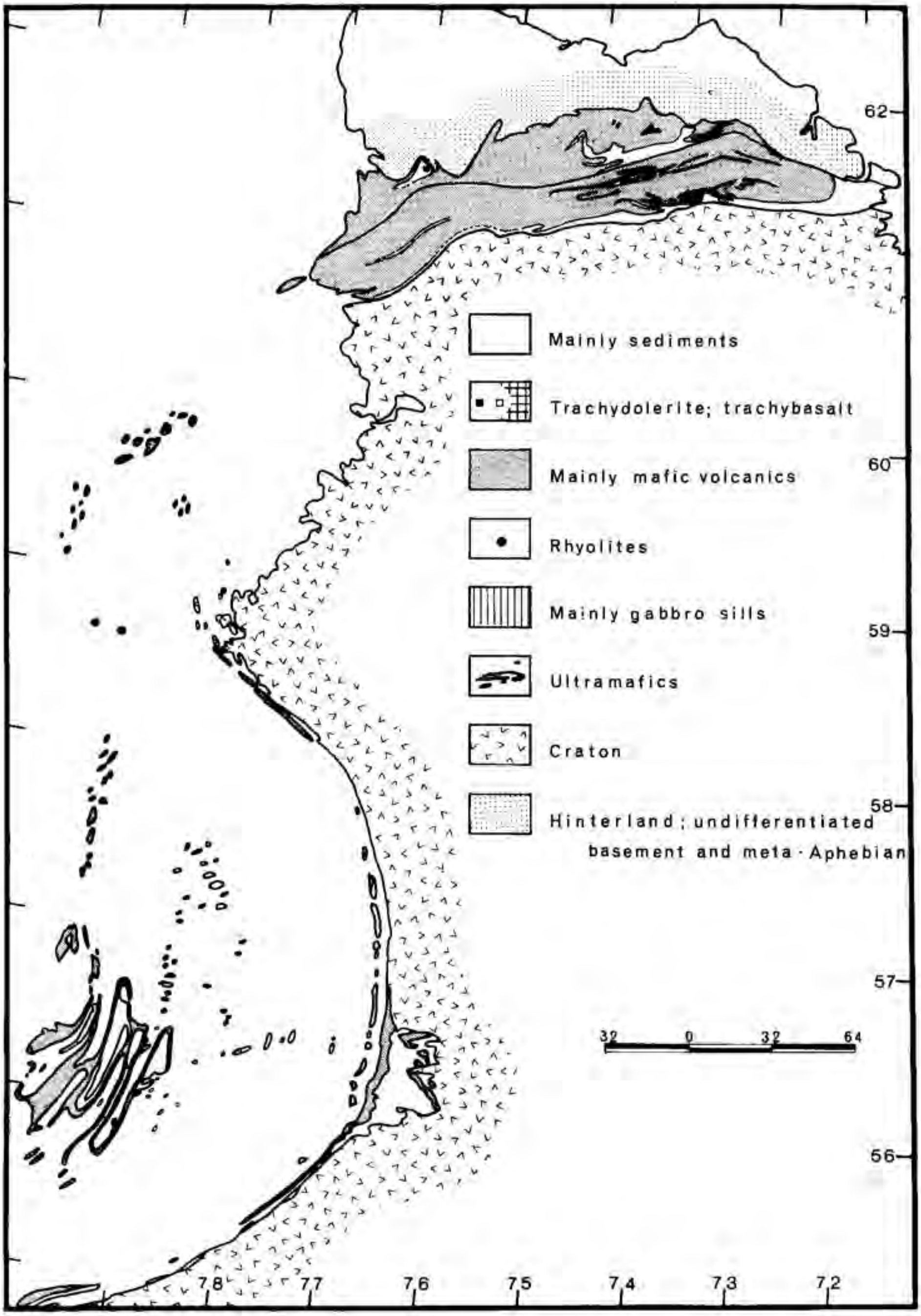


Figure 16. Distribution of magmatic rocks in the Cape Smith Belt and Belcher Fold Belt.

Omarolluk Formation (Unit 15) and Equivalents: Within the Belcher Islands the Omarolluk Formation is poorly exposed, and the best exposures are in Omarolluk Sound after which the formation is named. The type section for the lower member, which probably represents most of the formation, is on Gilmour Peninsula where about 7,000 feet of strata are well exposed (Fig. 14). About 4,300 feet of strata may be exposed on the smaller peninsula north of Gilmour Peninsula (Belcher Mining Corp., company rept.). The upper member is exposed on the Bakers Dozen and King George Islands lying north of the Belcher Islands (Fig. 16). Another 2,000 feet of strata (Fig. 14) are estimated to lie between the uppermost strata exposed on Gilmour Peninsula and the top of the formation as exposed on Loaf Island, the largest of the Bakers Dozen Islands that lies about 10 miles north of Cape Bartlett. The Omarolluk Formation also outcrops on several islands lying between the Belcher Islands and Richmond Gulf (Fig. 16).

The strata of the Omarolluk Formation are evenly layered and thinly laminated to thickly bedded. Although most beds are less than four feet thick a few are twelve feet or more thick. The lower part of the formation is composed mainly of dark green to grey greywacke and subgreywacke with thinner slate and argillite interbeds and a few tuff beds (Fig. 14, Table VII). A few intervals up to about 800 feet thick are composed almost entirely of sandstones. Toward the top of the formation concretionary structures become increasingly abundant and the strata become progressively lighter coloured and cleaner. The upper part of the formation is predominantly light greyish green, particularly on the Bakers Dozen and King George Islands. In addition the matrix component in the sandstones decreases upward in the section and the microcline content increases, so that quartzites and lithic and arkosic sandstones predominate in the upper member of the formation along with slate and argillite. This feature is not as obvious on the islands lying a few miles east of the Belcher Islands. This may indicate that only the lower strata of the Omarolluk Formation are exposed on these islands.

Several additional characteristics of the strata exposed on the Bakers Dozen and King George Islands tend to separate these upper strata from those lower in the section as exposed on Gilmour Peninsula. Whereas the sandstones in the lower part of the formation are predominantly very fine grained, those in the upper part are predominantly fine-grained but contain a considerable proportion of very fine, and medium-grained sandstones. Quartz and plagioclase grains are predominantly angular to subangular throughout. Clay balls, now altered to argillite and dolomite are locally abundant in the upper strata, and lenses of argillite-quartzite- and dolomite-clast conglomerate up to 15 feet long are not uncommon and contain clasts up to eight inches in diameter. Beds of conglomerate up to six feet thick occur near the top of the formation and contain quartzite, greywacke, and argillite clasts up to 2 1/2 feet in diameter in an argillite matrix. Conglomerate beds at the contact with the Loaf Formation (Unit 16) contain red quartzite boulders in a green argillite matrix.

Oscillation ripple-marks were seen in several places and channel-fill structures occur in some of the upper strata. Some of the latter are highly deformed, probably by slumping. Cross-bedding is rare or absent on Gilmour Peninsula and elsewhere in Omarolluk Sound, but low-angle cross-bedding is common on the Bakers Dozen and King George Islands, and torrential cross-bedding is present near the top of the formation. Several cross-beds were measured in subhorizontal strata on an island (Beach Island) at $78^{\circ}50'$ west longitude and $57^{\circ}20'$ north latitude, and on another island 40 miles to the southeast at $78^{\circ}02'$ west longitude and 57° north latitude. Both sets of measurements indicated transport in a southeast direction.

Small, flattened, brachiopod-like structures, about one inch long occur in the strata on the island at $78^{\circ}02'$ west longitude and 57° north latitude. They are composed of a rim, some made of calcite and others of quartz, up to 2 mm thick in which the quartz or calcite has crystallized perpendicular to the rim, and an anthraxolite core. A.W. Norris (pers. comm.) considers these structures to be inorganic. Similarly, structures closely resembling sponge spicules and occurring in a calcite vein in basalt near the top of the Flaherty Formation were considered by G.W. Sinclair (pers. comm.) to be inorganic.

South of Kasegalik Lake uneroded pillows at the top of the Flaherty Formation are overlain conformably by dark greenish grey coloured tuff, agglomerate, greywacke, argillite and conglomerate containing sedimentary and tuff cobbles up to six inches in diameter. In most places on the Bakers Dozen Islands the upper contact with the Loaf Formation is gradational and red-coloured sediments are interbedded with greyish green strata. A few small unconformities, however, occur locally at and near the top of the Omarolluk Formation.

At least 2,400 feet of Proterozoic subgreywacke, quartzite, slate, siltstone and conglomerate outcrop in the Churchill area (Bostock, 1969b). Cross-bedding is abundant in certain units and indicates a westward current flow. Feldspars seem to be much less abundant than in the Omarolluk Formation of the Belcher Islands with which the "Churchill Quartzites" are tentatively correlated.

Loaf Formation (Unit 16) and Equivalents: The Loaf Formation outcrops on only three of the Bakers Dozen Islands (Fig. 16). The formation is named after Loaf Island, the largest of the Bakers Dozen Islands, on which about 700 feet of these strata are exposed.

Nearly all of the rocks are thin-bedded to massive arkoses and subarkoses in beds up to eight feet thick that are stained pink to red by interstitial hematite. A few beds up to 1 1/2 feet thick, of ferruginous carbonate, argillite and conglomerates with cobbles up to eight inches in diameter are also present (Table VII). The strata are commonly fine- to medium-grained. Detrital quartz and feldspar grains are angular to subangular and a few per cent of microcline are present in most rocks.

The strata are commonly cross-bedded and slump structures and a few minor local unconformities have been observed. The lower contact with the Omarolluk Formation is conformable and gradational. A few red beds occur locally in the Omarolluk Formation to the north and south of the Loaf Formation. Some of the uppermost beds of the Omarolluk Formation in these areas may, therefore, be stratigraphically equivalent to the lower strata of the Loaf Formation.

The stratigraphic position of red argillite found in a drill hole west of James Bay (*see* General Features, Belcher Fold Belt, this paper) is uncertain, but the rock may be equivalent to the Loaf Formation.

Depositional Environment

First Cycle

The depositional environment varied considerably during the cyclical filling of the Belcher Basin. In the Belcher Islands area, deposition of the Kasegalik Formation (Unit 1) was in a shallow-water stable shelf to miogeosynclinal environment. The Kasegalik Formation may be underlain by arkose and is interpreted as grading laterally eastward into the basal arkoses of the Pachi Group at Richmond Gulf. These arkoses are probably largely terrestrial and were deposited on the basement gneisses, partly in basins and small grabens.

Second Cycle

The second cycle began abruptly in the Belcher Islands-Richmond Gulf area with basic volcanism that was centred along an axis trending north-northeast through the central Belcher Islands and which may have extended northward to the Cape Smith Belt (Fig. 13). As volcanism waned the basin deepened somewhat and shales, tuffs and greywackes were deposited. As conditions became more stable, the basin became shallow again and progressively greater amounts of more mature marine sediments, mainly carbonates and pure quartz sands, were deposited. Intermittently unstable conditions, probably related to volcanic activity and deformation in nearby areas, gave rise to rhythmically interlayered argillaceous, siliceous and carbonate sediments. Deposition of the second cycle sediments of the Belcher Group occurred mainly in a fairly stable, shallow-water, marine environment, during which time the shore line migrated back and forth many times. Several relatively thin red bed sequences at different stratigraphic horizons in the second cycle strata may have been deposited in a terrestrial environment, possible on a broad fluvial plain. Black shales are rare and euxinic environments existed at rare intervals, mainly during volcanism.

The sedimentary strata of the second cycle are coarser grained in the Richmond Gulf area than on the Belcher Islands. The arkoses of the Richmond Gulf Formation rest unconformably on the Pachi volcanics. They are probably mostly terrestrial and were deposited mainly in valleys and small grabens (Woodcock, 1960). Interbedded sequences composed predominantly of dolomite and of sandstone make up most of the Nastapoka Group and overlie the Richmond Gulf unconformably. The depositional environment fluctuated between shallow-water marine and terrestrial during deposition of this sequence, which in several places begins with a stromatolite-bearing dolomite sequence that rests directly on the gneissic basement. This sequence was interrupted by a brief, localized episode of deposition of basic volcanic rocks. These rocks are stratigraphically correlated with the upper member of the Laddie Formation on the Belcher Islands.

A few cross-bedding determinations indicated westward transport (Fig. 13) of detrital sediments in the Richmond Gulf area (H. Hofmann, pers. comm.). This observation is in agreement with the correlations made in Figure 15 and Table VII which show that relatively coarse, mostly terrestrial, sandstones grade westward into mainly well-sorted sandstones and finer-grained argillites and some chemical sediments. Thus the source for most of the clastic material in the strata of cycles 1 and 2 was probably the cratonic area to the east.

Third Cycle

The volcanism that initiated the third cycle was centred west of the Belcher Islands, and was the most extensive volcanism of all (Fig. 13). It seems to have marked the end of Aphebian igneous activity in the Belcher Basin. The precise mode of extrusion of these upper volcanic rocks represented by the Flaherty Formation on the Belcher Islands is not known. They are possibly related to a large fault or rift-zone that has been postulated as extending from James Bay northward to west of the Ottawa Islands (Innes *et al.*, 1967). The relationship of these upper basic volcanics to associated strata is similar to conditions present in the Cape Smith Belt and in the Labrador Trough, where culminating basic volcanism occurs late in the geosynclinal sequence. In the Belcher Basin, this last period of volcanism marked the final encroachment of orogenic activity from a eugeosynclinal area to the west into the eastern miogeosynclinal basin, and the end of stable shelf-miogeosynclinal deposition.

Following the volcanism a thick sequence of alternately thick-bedded greywackes and thin bedded argillites representing most of the Omarolluk Formation on the Belcher Islands, was deposited in relatively deep quiet waters, and may represent turbidite deposition of a flysch-like sequence. These greywackes and argillites grade upward into slightly coarser, cleaner, and locally conglomeratic cross-bedded, lithic and arkosic sandstones in the upper part of the Omarolluk Formation. These rocks may represent a shallow-water marine molasse sequence. These strata in turn grade upward into the terrestrial conglomeratic, arkosic sandstones of the Loaf Formation that may represent a terrestrial molasse sequence.

A few cross-bedding measurements made at two localities in upper strata of the Omarolluk Formation indicate southeastward transport (Fig. 13). Westward transport is indicated for lithologically similar strata at Churchill (Bostock, 1969b). It is concluded therefore that much if not most of the detrital material deposited in the Omarolluk Formation was derived by erosion in an orogenic area to the west or northwest, in the central part of Hudson Bay. The possibility that such an orogenic area existed has previously been postulated by J.A. Donaldson (pers. comm.) as a result of work done west of Hudson Bay.

Varying proportions of quartz and plagioclase occur in all of the detrital sediments of the Belcher Group, and a few grains of microcline are commonly scattered through them also. Rare quartz and some plagioclase grains are also common in the carbonate strata and in the Kipalu Iron Formation. Microcline is increasingly abundant toward the top of the Omarolluk Formation and forms a few per cent of perhaps most sandstone beds in the upper part of the Omarolluk Formation and in all of the Loaf Formation. It is concluded therefore that cratonic rocks provided a continual source of detritus for the Belcher Group, and that following the last volcanism the cratonic detritus was diluted by detritus derived from an orogenic area to the northwest. This also occurred on a much smaller scale following the initial and perhaps the second volcanism. In any event, as the supply of orogenic, basaltic detritus decreased following the last volcanism, cratonic rocks again became the dominant source of detritus. The increased amount of microcline and the southeasterly transport may indicate that a major cratonic source had recently been uncovered to the northwest and was supplying material in addition to the cratonic area to the east.

Conclusion

The presently exposed strata of the Belcher Fold Belt were deposited in an embayed miogeosynclinal basin that may at times have been more like an exogeosyncline. An eugeosynclinal area lay somewhere to the west or northwest of the Belcher Islands. Deformation of Belcher Basin strata had already begun during deposition of the Loaf Formation. The major deformation and mild regional metamorphism, whether it occurred shortly after deposition of the Loaf Formation or somewhat later, probably took place 1,600-1,700 m.y. ago near the end of Aphebian time. Because of the general lack of flow-folding in the Belcher Island strata a decollement zone may possibly occur between the basement gneisses and the Aphebian strata. Whether or not this is expressed as a thrust fault in the Richmond Gulf-Nastapoka Island area is not yet known.

The area was probably affected by a thermal event and/or a mild disturbance of late Neohelikian age (830-1,050 m.y. ago) when pillow rims were affected and a few basic trap dykes were emplaced on the Belcher Islands.

While it is possible that the Circum-Ungava Geosyncline terminates in the Belcher Basin, the writers do not consider it as likely. The belt of magnetic and negative gravity anomalies, previously described (Hood *et al.*, 1969; Innes *et al.*, 1967, 1968; Morley *et al.*, 1968) that extends from Cape Smith to west of the Ottawa Islands, thence south, west of the Belcher Islands to the Sutton Lake area and west to the Churchill-Superior Province boundary may not only mark a transition westward into metamorphosed strata, but may also mark the approximate position of the eugeosynclinal area. The marked increase in thickness of the Flaherty Formation in the western Belcher Islands supports this hypothesis.

A second anomalous belt, perhaps more obvious from aeromagnetic than from gravity data (Hood *et al.*, 1969; Morley *et al.*, 1968) extends from Cape Smith west-southwest almost straight across Hudson Bay to the Owl River area south of Churchill and thence south-southwest to the Nelson River area (Fig. 13). Ruffman (1969) has suggested that this belt may mark the position of the Churchill Province-Superior Province boundary at depth. It might also mark the approximate position of the eugeosyncline.

RELATIONS BETWEEN THE LABRADOR TROUGH, THE CAPE SMITH BELT, AND THE BELCHER FOLD BELT

The three units described above have identical geotectonic positions at the margin of the Churchill Province. Their structures are similar. They are composed of the same rock types and those follow in somewhat similar sequence. A formational correlation between them is however not possible, and age determinations only imply that most of their filling has been folded and metamorphosed between 1,600 and 1,650 m.y. ago. It is possible that one of the stratigraphic units in the Cape Smith Belt (the Chukotat Group) is somewhat younger than most of the other stratigraphic units and that it has been deposited between 1,550 and 1,450 m.y. (Beall *et al.*, 1963).

It has been shown above that two major facies zones are present in the Labrador Trough: a zone composed mainly of sediments in the west, and a dominantly volcanic zone in the east. The western zone has been entirely eroded north of Leaf Bay (lat. 59°). Stratigraphy and facies of the part of the Labrador Trough north of Leaf Bay correspond to those of the eastern zone farther south. The Archean basement of this eastern, eugeosynclinal, zone of the Labrador Trough has been folded jointly with its cover and underwent some Hudsonian metamorphism. They therefore give Hudsonian ages by the K-Ar method.

The correlation between the Cape Smith Belt and the Labrador Trough is based on a general similarity in lithology and sequence, on age determinations, and on geophysical relationships. At least the Povungnituk Group in the Cape Smith Belt has been deposited before 1,600 m.y. (Beall *et al.*, 1963), and its general sequence is similar to that of one of the upper two cycles in the Labrador Trough. The Chukotat Group may be somewhat younger but its similarity in facies to the older Povungnituk Group is so great that it undoubtedly forms part of the same geosynclinal filling. The Cape Smith Belt is underlain by an Archean basement that suffered some Hudsonian deformation and that gives Hudsonian ages by the K-Ar method. The boundary between the age province has not been determined precisely but its approximate location is shown on Figure 1, where it has been used to separate the eugeosynclinal and miogeosynclinal domains. A negative gravity anomaly finally follows the southern boundary of the Cape Smith Belt and can be traced towards the area west of Payne Bay. In analogy to the Labrador Trough, it is believed probable that the Cape Smith Belt was once accompanied by a sedimentary belt at the frontal side (i.e. to the south); this belt has been eroded since.

The stratigraphy of the Belcher Fold Belt has properties in common with the two belts mentioned above. It shows a marked cyclicity of alternating periods of deposition of sandstone and precipitate sediments on the one hand, and of volcanics on the other. However within the Belcher Fold Belt the cycle is the inverse of that in the Labrador Trough. The thickness of the sequence increases from the frontal (eastern) towards the distal (western) portion of the belt. The belt appears to continue into the Ottawa Islands and the Cape Smith Belt.

The Belcher Fold Belt is much wider than the best preserved segment of the Labrador Trough. The subdivision into a frontal zone of mainly sedimentary rocks and a distal zone of volcanic rocks is poorly defined, the proportion of volcanic appears to increase westward and northward. It is therefore not feasible to subdivide the Belcher Fold Belt into eugeosynclinal and miogeosynclinal zones, as in the Labrador Trough; neither is it possible to interpret it as a eugeosyncline whose frontal, sedimentary zone has been eroded, as in the case of the Cape Smith Belt. The total thickness of the filling of the Belcher Fold Belt, particularly of its frontal part (Richmond Gulf area), is thin compared to the total thickness of the Labrador Trough or of the Cape Smith Belt. The Belcher Fold Belt possibly represents a miogeosynclinal embayment in the foreland of a eugeosyncline that used to lie to the west or to the northwest.

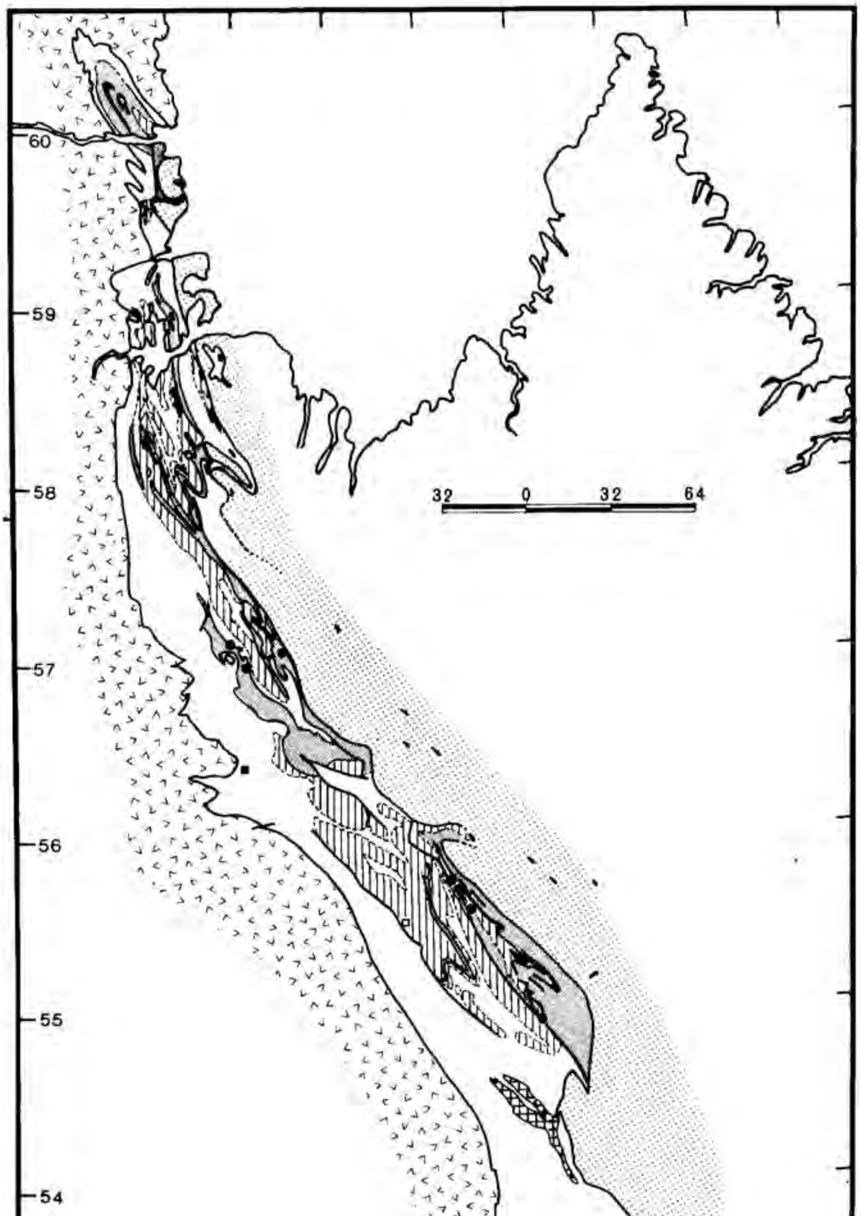


Figure 17. Distribution of magmatic rocks in the Labrador Trough.

CHEMICAL CHARACTERISTICS OF THE IGNEOUS ROCKS

Most studies of igneous rocks of the Circum-Ungava geosyncline have been done in the Labrador Trough and the following account reflects this limitation. One exception is the work of Beall (1962) on flows and sills in part of the Cape Smith-Wakeham Bay Belt. From this work, one may judge that the petrology of Cape Smith-Wakeham Bay igneous rocks is generally similar to that of the Labrador Trough igneous rocks. The principal studies upon which the present account is based are those of Sauvé and Bergeron (1965) and Hardy (1968) in the north and of Girard (1965) and Baragar (1960, 1967) in the central and southern parts of the Trough. Unpublished analyses of the Quebec Department of Natural Resources have been included in the compilations (*see* Tables XI and XII).

Volcanic Rocks

The compositions of 25 lavas from various parts of the Labrador Trough are given in Table XI. Thirteen are new analyses and the remainder are taken from the literature. They are shown on an MgO-$FeO_{(tot)}$-Alkalies, CaO-K_2O-Na_2O triangular plot in Figure 18. It is immediately evident that they fall into three categories: 1) the normal lavas that comprise most of the analyses, 2) lavas of the Seward Subgroup, and 3) an acidic lava associated with the Murdoch Formation.

The normal lavas are mainly tholeiitic basalts from the Willbob-Hellancourt, Menihek, and Attikamagen II (basalt-shale sequence) units. They spread moderately along the magnesia-iron and lime-soda sides of the triangle, variations that are characteristic of fractional crystallization of tholeiitic magmas in high-level magma chambers. The outstanding feature of the diagram is the excessively low potash content of the lavas, regardless of their degree of differentiation. Most of the lavas are metabasalts of greenschist facies. On the basis of their normative feldspar content, fourteen of them are basalts (An_{50}), five are andesites (An_{30}-An_{50}), and two are dacites (An_{30}-$An_{12\frac{1}{2}}$). However, a few appear to be spilitic, which accounts for their relatively sodic plagioclases. On the basis of colour and differentiation indexes (Fig. 19) and lack of, or minor, normative quartz, nearly all could be classed as basalts.

The main chemical characteristics of the normal lavas are summarized in the series of frequency distribution diagrams that are given in Figures 19 and 20 together with comparable diagrams for oceanic basalts and two Archean volcanic belts. A point of major interest is that the normal lavas of the Labrador Trough are more basic, as measured by colour and differentiation indexes, than both the oceanic tholeiites and the Archean lavas. The mode of their potash content (0.25%) is only slightly greater than that of the oceanic tholeiites (0.15%) and the soda and titanium contents are similar.

In Table VIII the average composition of the 22 normal lavas is compared with that of 25 oceanic tholeiites and with the weighted average of the mafic fraction of Archean volcanic belts (Baragar and Goodwin, 1969). The similarity between the averages of the Labrador, and oceanic analyses is striking, whereas the Archean volcanics are a little less mafic. Volcanic rocks of the Labrador Trough, therefore, appear to be at least as, "primitive" as oceanic tholeiites and somewhat more so than volcanic rocks of the Archean belts. This contradicts the view of Engel *et al.* (1965, p. 730) that the potash content of lavas derived from the mantle beneath continents has diminished with time.

Seward Lavas: Lavas of the Seward Subgroup have compositions that are quite unlike the tholeiitic lavas of the major part of the Labrador Trough. They are rich in alkalis and iron and the iron is highly oxidized. They are composed of a fine-grained assemblage of feldspar and subordinate chlorite clouded throughout with finely divided iron oxide. Feldspars are probably alkalic but because of their small grain-size and clouding by iron oxides and chlorite,

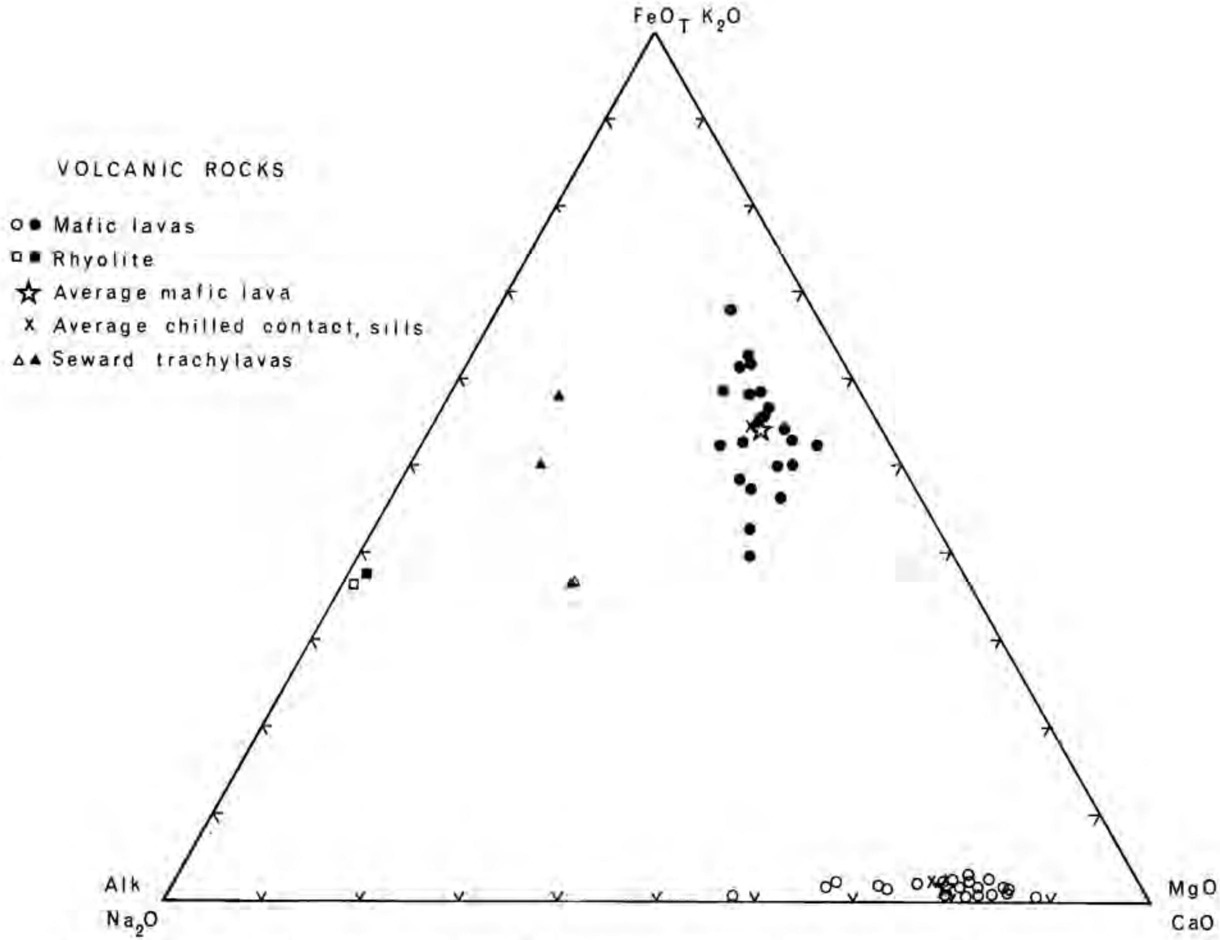


Figure 18. MgO-FeO$_{tot}$-Alkalies and CaO-Na_2O-K_2O plots of Labrador Trough volcanics.

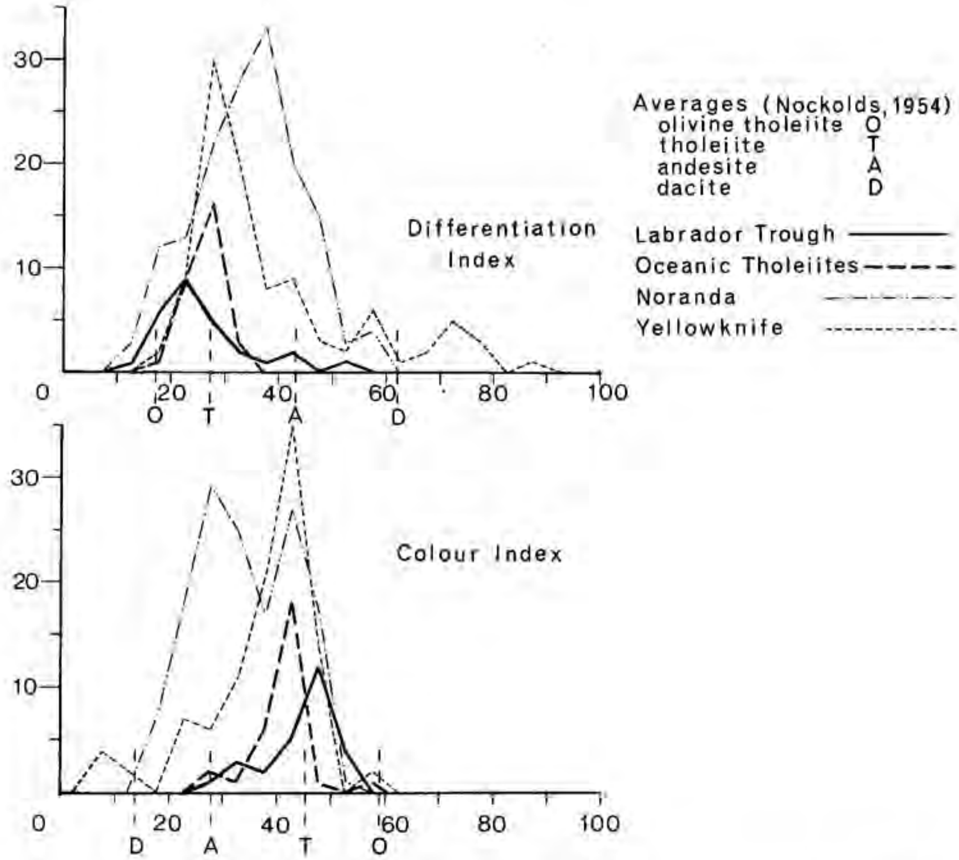


Figure 19. Frequency distribution of colour and differentiation indexes of Labrador Trough lavas and other volcanic suites.

they are difficult to determine. The rocks are altered but original igneous textures are recognizable and the composition may not differ significantly from that of the original lava.

These lavas might be trachyandesites or trachybasalts. In Table IX their analyses are compared with those of a trachyandesite and trachybasalt from Gough Island (Le Maitre, 1962). Except for its high-alumina content the trachybasalt of Gough Island is very similar to the Seward lavas. If the latter are trachybasalts or trachyandesites their presence in the Labrador Trough is noteworthy because alkaline lavas are rarely a component of geosynclinal volcanic assemblages. It may be significant that the environment into which these lavas extruded shows evidence of being subaerial, in contrast to the marine, geosynclinal environment of the tholeiitic lavas. It is also interesting to note that the Nimish lavas which Sauvé (1953) believed to be subaerial in part were described by him as being orthoclase-bearing and strongly magnetic. S. Zajac (pers. comm.) confirms that these volcanics are highly potassic. Both characterized by a high content of alkalis - both soda and potash - a low content of lime and magnesia, and by highly oxidized iron. Accordingly, the Hematite Lake sill might be classed as a trachydolerite and assigned to the same volcanic province as the Seward lavas.

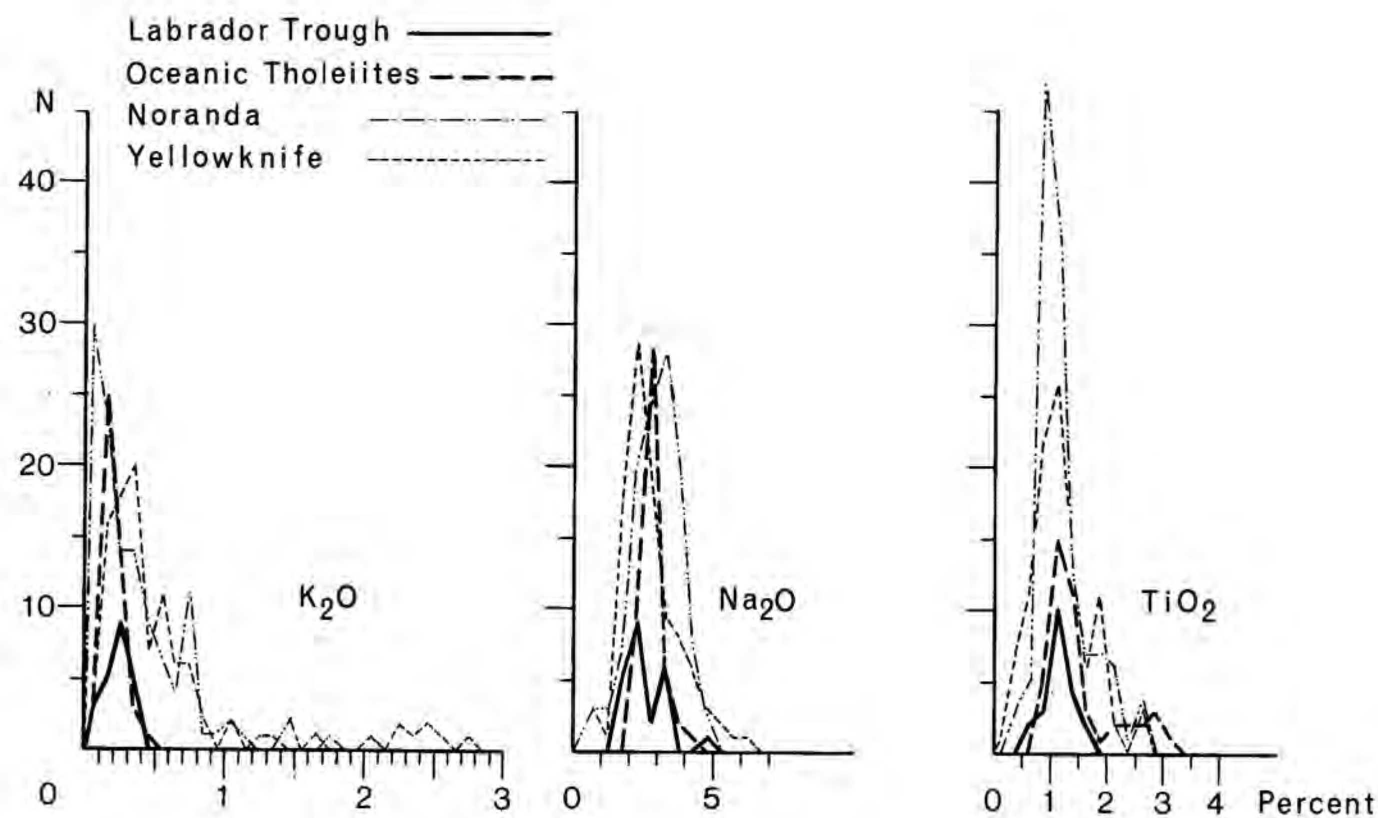


Figure 20. Frequency distribution of K_2O, Na_2O, and TiO_2 in Labrador Trough lavas and other volcanic suites.

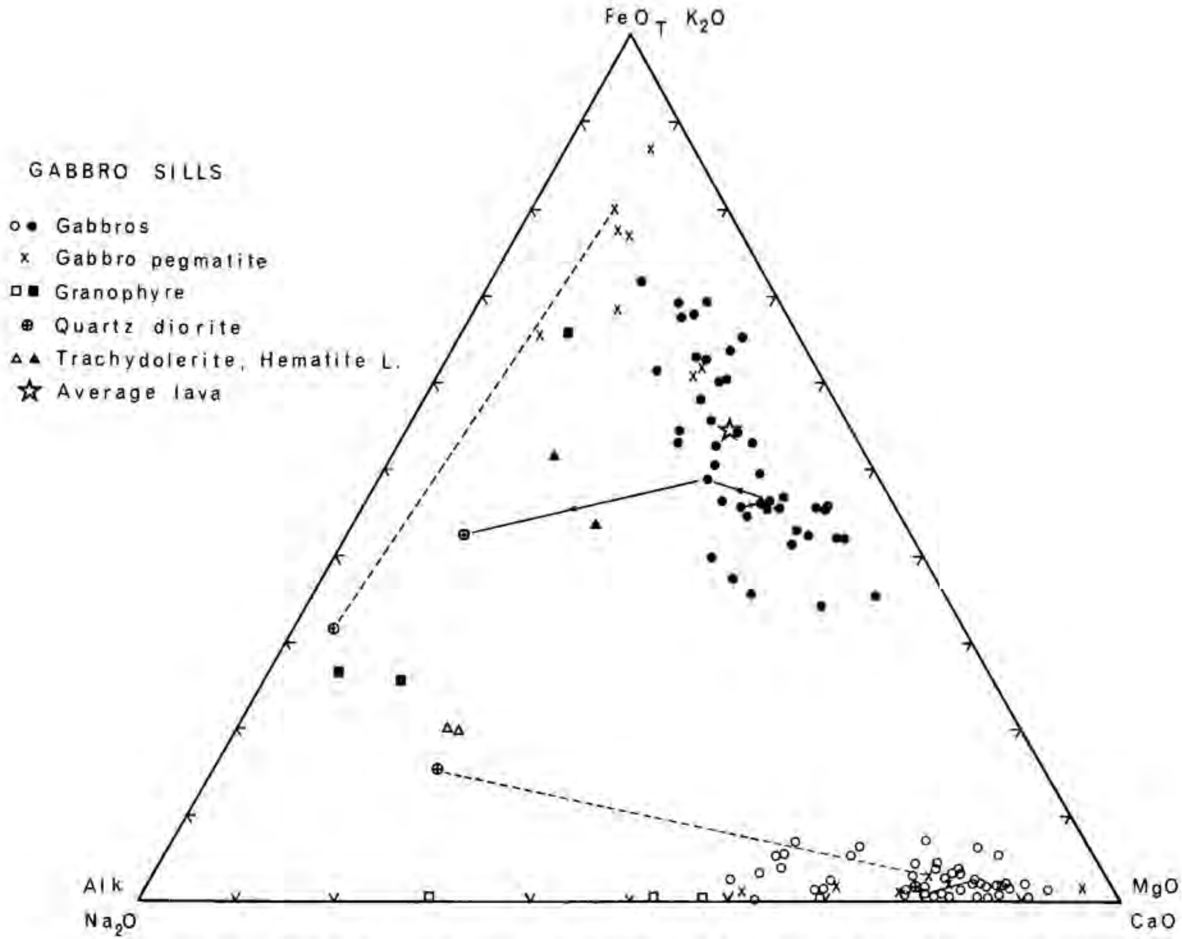


Figure 21. MgO-FeO$_{tot}$-Alkalies and CaO-Na$_2$O-K$_2$O plots of Labrador Trough gabbro.

sets of lavas are possibly part of a separate volcanic province associated with the more stable facies of the geosyncline. In Figure 16 they can be seen to occur along the eastern fringe of the miogeosynclinal part of the Labrador Trough.

Acidic Lava: The only acidic volcanic rock analysed is from a small mass adjoining the Murdoch Formation in the Menihek Lakes map-area (Frarey, 1961) at about $54^{\circ}50$'N[1]. It is a reddish, fine-grained, feldspar phyric rock containing much finely disseminated hematite. The analyses is given in Table X together with average analyses of calc-alkali rhyolite (Nockolds, 1954) and of Archean salic rocks of the Canadian Shield (Baragar and Goodwin, 1969). The Labrador Trough rhyolite is similar to Nockolds' average calc-alkaline rhyolite and quite unlike the normal, low-potash acidic lavas characteristic of Archean belts.

Gabbroic Sills

Gabbro sills of more than a few hundred feet in thickness are almost invariably differentiated to some degree. The differentiation products range from broad areas of subpegmatitic gabbro to pegmatitic patches, granophyric veins (Baragar, 1967) and in more extreme cases to layers of quartz diorite at the top of sills (Sauvé and Bergeron, 1965; Hardy, 1968). The basal zones of differentiated sills are enriched in olivine and pyroxenes.

The character of differentiation is shown on the $MgO-FeO_{(tot)}$-Alkalis $CaO-K_2O-Na_2O$ triangular diagram of Figure 21 where 65 analyses of gabbros from the Labrador Trough are plotted. Thirty-three of these are new analyses and are presented in Table XII; the remainder are from the literature (Sauvé and Bergeron, 1965; Baragar, 1960, 1967; Hardy, 1968).

Throughout the Labrador Trough, the main trend of differentiation is one of extreme iron enrichment for the major part of crystallization, followed by enrichment in alkalis. This has been noted by most previous investigators. The Labrador trend is similar to that of the Skaergaard Intrusion except in one important respect; potash generally becomes impoverished rather than enriched with differentiation. This can be clearly seen in Figure 21 where most of the acid differentiates have slightly lower potash to lime-soda ratios than the more basic rocks of the series. Such potash impoverishment is presumably a reflection of the low initial potash content of the Labrador magma which may have been lower than that in co-existing plagioclase and was therefore, further diminished by fractionation. In one of the sills with a quartz diorite layer at its top (Sauvé and Bergeron, 1965) the rock is enriched, rather than impoverished in potash, in the quartz diorite member. A slightly higher than normal potash content in the initial magma, possibly by contamination with wallrock, could account for this difference.

A contrasting trend of differentiation was found by Hardy (1968) in one sill near the north tip of the Labrador Trough (Fig. 21). Silica and soda increase upward in the sill, with little or no increase in iron. This is the typical calc-alkaline differentiation trend except that in this case potash is low throughout, in keeping with its general tenor in the Labrador Trough magmatic province. Hardy (1968, pp. 156-159) attributes this trend to a greater retention of water by the sill, resulting in early crystallization of iron in the manner suggested by Roeder and Osborne (1966).

Two analyses that appear in Table XII and are plotted in Figure 21 are in complete contrast with the main body of analyses from the Labrador Trough. Both are samples taken from the same sill in the western part of the Central Labrador Trough near Hematite Lake ($56^{\circ}26$'N, $68^{\circ}54$'W) (Dimroth, 1969, p. 32). They are remarkably similar to one another and to the Seward lavas discussed previously (Table IX). Both the Seward lavas and the Hematite Lake sill are

[1]This sample was kindly loaned for analysis by Dr. M.J. Frarey.

Relationship of Volcanic Rocks and Gabbroic Sills

The average compositions of 22 normal lavas and of five chilled contacts of gabbroic sills are compared in Table VIII and are shown on the $MgO-FeO_{(tot)}$-Alkalis, $CaO-K_2O-Na_2O$ diagram of Figure 18. The compositions are very similar and their position in Figure 18 is consistent with that of a parent magma of the differentiation sequence. Lavas and gabbros are both characterized by very low potash content and both have an iron-enrichment differentiation trend; the lavas a subdued version of that of the gabbros. It is logical, therefore, to assume that normal lavas and normal gabbros are simply surface and subsurface manifestations of the same parent magma. Variations in the composition of the lavas appear to have resulted from fractionation of the type observed in the sills.

The rhyolite is in a class by itself. Because of its extraordinarily high potash content it is inconceivable that this rhyolite could have differentiated from the normal magma sequence yet it is now intimately associated with volcanic rocks of the eugeosynclinal belt. The rhyolites are present in very small quantities and a plausible alternate explanation might be that they formed by partial melting of the underlying sialic crust.

DISCUSSION

Comparison with Alpine Geosynclines

Similarities between the Labrador Trough and Alpine geosynclines have long been recognized (de Roemer, 1956; Bergeron, 1957b; Fahrig, 1957; Gastil *et al.*, 1960; Bergeron, 1965; Baragar, 1967). A more precise comparison is now possible, and suggests that the Labrador Trough has particularly close analogies to geosynclines of the Mediterranean System described by Aubouin (1965). The analogies concern the scale of the basin, the subdivision in paleogeographic zones, the stratigraphy and lithology of the basin filling, and the final deformation, metamorphism, and post-tectonic uplift.

The Labrador Trough is at present 125 km wide at latitude 56°. A conservative estimate of post-tectonic erosion (five km at the western contact, assuming a basement gradient of five degrees) brings its former width to 175 km. The Labrador Trough, from Payne Bay to the Grenville Front, is 950 km long. Alpine chains are between 150 and 200 km wide, and between 1,000 and 2,000 km long.

The typical paleogeographic zones of Mediterranean geosynclines (Aubouin, 1965) have been recognized in the Labrador Trough: the sedimentary belt in the west of the Trough represents the miogeosynclinal basin, and the volcanic belt in the east is the eugeosynclinal basin. Traces of a ridge separating both (named miogeanticlinal ridge by Aubouin (1965)) have been recognized at some stages of the evolution of the Trough (*see* Fig. 11). This ridge is covered by a thick pile of mafic extrusives and intrusives. It is therefore part of the eugeosynclinal domain in the Labrador Trough. A source area east of the Trough corresponding to Aubouin's eugeanticlinal ridge has also been recognized. This ridge is overlain by a predominantly sedimentary sequence that is remarkably thin compared to the thickness of the deposits of the eugeosynclinal basin.

Aubouin (1965) subdivided the filling stage of the Alpine chains into four stages: (1) A pre-ophiolitic stage of limestone deposition, (2) an ophiolitic stage, (3) a synorogenic flysch stage and, (4) a post-orogenic molasse stage. Deposits of the flysch and molasse stages have not been observed in the Labrador Trough. The sandstone-precipitate stage, and the shale-greywacke-basalt stage of the Labrador Trough, on the other hand, are analogous to the pre-ophiolitic and ophiolitic stages, although with a different lithological content. The ophiolitic stage of the Labrador Trough, at least of the first

TABLE VIII

AVERAGE COMPOSITION OF NORMAL LAVAS COMPARED WITH

AVERAGES OF ARCHEAN MAFIC LAVAS AND OCEANIC THOLEIITES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O^+	H_2O^-	TiO_2	MnO	P_2O_5	CO_2	S
1	48.80	14.05	2.36	9.78	7.28	10.07	2.52	0.22	3.00	0.14	1.08	0.22[1]	0.10[1]	0.22[2]	0.13[3]
2	49.25	14.46	2.24	10.09	7.31	9.53	2.56	0.27	3.61	0.09	1.09	0.22	0.08	1.39	0.22
3	49.29	16.89	2.06	6.79	7.46	11.55	2.81	0.18	0.63[4]	0.49[4]	1.37	0.16	0.12	-	-
4	52.6	15.1	2.38	7.80	5.23	7.94	2.80	0.55	2.5	-	1.05	0.18	0.20	1.20	-

1 Average of 22 normal lavas, Labrador Trough.

2 Average of 5 chilled margins of gabbro sills, Labrador Trough.

3 Average of 25 oceanic tholeiites (Nicholls, 1961; Engel and Engel, 1963; Engel, Engel and Havens, 1965; Engel, Fisher, and Engel, 1965; Cann and Vine, 1966; Engel and Fisher, 1969).

4 Weighted average of mafic fraction of Archean volcanic belts, Canadian Shield (Baragar and Goodwin, 1969)

[1] 20 analyses, [2] 19 analyses, [3] 15 analyses, [4] $H_2O + CO_2$

cycle, may furthermore be subdivided into two substages: (a) an early substage of rapid subsidence of separated basins filled by the debris derived from a central geanticline and from the hinterland and (b) the ophiolitic stage proper characterized by wholesale eruption of volcanic material in the east of the Trough. No such subdivision has been recognized in Alpine chains. The Labrador Trough did not mature to orogenesis after the deposition of the first ophiolite suite, as did the geosyncline of the Alpine system. At least one, and possibly two more stratigraphic cycles, each with its own pre-ophiolitic and ophiolitic phase, accumulated before folding.

The lithology of the deposits of the Labrador Trough is in marked contrast to the lithology of the Alpine geosynclinal filling. This difference appears to reflect the entirely different tectonic settings of both geosynclinal types.

The Mediterranean geosynclines developed as "starved" basins embedded in a wide marine domain. Not much terrigenous material was deposited before the flysch stage. Thick reef limestones were therefore deposited on some ridges (i.e. on ridges of the "Gavrovo type", Aubouin, 1965) during the pre-ophiolitic phase whereas thin pelites and calc-pelites normally represent the basin facies. Relatively thin pelagic pelites and bedded cherts interfinger with the ophiolites, whereas calcareous rocks were commonly deposited at the same time in the external zones.

The Circum-Ungava geosyncline, on the other hand, was apparently embedded in a continental domain at all times. Thick clastic deposits therefore form the filling of basins, whereas ridges were eroded in some cases.

Shales, greywackes and open fabric grits and conglomerates are the typical clastics of the ophiolite phase in the Labrador Trough. These rocks show many of the sedimentological characteristics of flysch (Dimroth, 1968b), provided subsidence was rapid enough to sustain a relatively deep basin. They are, however, not a flysch facies in the sense of the Alpine terminology used by Aubouin (1965), and are more closely related to the slate-greywacke suites that occur in the ophiolitic stage of Paleozoic mountain chains, e.g. in the Hercynian chains of Central Europe (Plessmann, 1964; Krebs, 1968). Their evolution is closely linked to syn-sedimentary normal faulting; they are not related to thrusting as is the case in many true flysch suites.

Pre-orogenic normal faulting has been described from the Alps (Günzler-Seifert, 1952; Baer, 1959; Trümpy, 1960), and from the Hercynian chains of Central Europe (Krebs, 1968). These relations are duplicated in the Labrador Trough, where syn-sedimentary faults formed normal and parallel to the geosynclinal trend (Dimroth, 1968b), during the deposition of the Seward, Pistolet and particularly the Swampy Bay Subgroup. It is not impossible that faulting continued during the volcanic stage. The aggregate thickness of gabbro sills in the Attikamagen II Formation decreases abruptly as the eastern margin of the Trough is approached. This could be attributed to the presence of a developing marginal fault or flexure. In this respect therefore the Labrador Trough may also be similar to the European geosynclines, where the ophiolitic volcanism is closely related to faulting at basin margins (Aubouin, 1965).

The Labrador Trough was subsequently shortened to less than 1/2 of its original width, and a syn-kinematic and post-kinematic metamorphism of intermediate pressure (Gélinas, 1965; Baragar, 1967, p. 145) affected its distal zones, and its hinterland. The metamorphism was followed by post-tectonic uplift, particularly in the distal zone of the geosyncline.

All criteria discussed above suggest that the Labrador Trough represents a complete geosyncline and not only its foothill zone, as had been postulated (e.g. de Roemer, 1956; Bergeron, 1957b). In size, stratigraphic sequence and paleogeographic evolution the Labrador geosyncline closely resembles the geosynclines of the Mediterranean system, but its tectonic setting is different. The Labrador Trough is furthermore characterized by an extremely thick ophiolitic suite, unequalled in any Phanerozoic geosyncline of the Alpine type. In this respect the Labrador Trough shows closer relations to Archean volcanic belts than to the Mesozoic geosynclines of the Alpine system.

TABLE IX

SEWARD LAVAS AND HEMATITE LAKE GABBROS COMPARED WITH TRACHYANDESITE AND TRACHYBASALT FROM GOUGH ISLAND.

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O^+	H_2O^-	TiO_2	MnO	P_2O_5	CO_2	S
1	49.66	13.95	11.16	3.08	3.98	4.21	4.90	2.76	2.40	0.06	1.27	0.21	0.13	1.95	0.09
2	49.92	15.47	7.87	5.33	3.42	2.79	4.80	4.30	1.96	0.06	1.38	0.22	0.14	2.12	0.09
3	53.58	14.30	9.33	4.27	4.17	2.32	5.90	2.00	2.38	0.13	1.64	0.15	0.13	0.03	-
4	54.26	14.69	5.14	5.26	5.74	1.99	5.40	1.84	3.07	0.09	1.72	0.17	0.16	0.04	-
5	55.89	16.90	1.41	6.71	2.26	4.48	4.44	4.58	0.99	0.26	1.60	0.13	0.34	-	-
6	49.21	18.26	6.82	2.32	4.24	5.37	3.75	3.88	1.66	1.00	2.62	0.13	0.54	-	-

Analyses 1-2 Seward lavas, Lac Musset, Central Labrador Trough (Table IV).
3-4 Hematite Lake gabbros, Central Labrador Trough (Table V).
5 Trachyandesite, Gough Island (Le Maitre, 1962).
6 Trachybasalt, Gough Island (Le Maitre, 1962).

TABLE X

RHYOLITE OF THE LABRADOR TROUGH, COMPARED WITH AVERAGE CALC-ALKALINE AND ARCHEAN RHYOLITES

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O^+	H_2O^-	TiO_2	MnO	P_2O_5	CO_2	S
1	72.3	12.6	4.7	0.6	<0.5	0.1	2.9	5.0	0.8		0.50	<0.02	0.09	<0.1	
2	73.66	13.45	1.25	0.75	0.32	1.13	2.99	5.35	0.78		0.22	0.03	0.07		
3	66.8	14.8	1.09	3.01	1.78	2.55	4.06	1.78	2.87		0.50	0.10	-	-	-

1 Rhyolite, south-central Labrador Trough (Table IV).
2 Average calc-alkali rhyolite (Nockolds, 1954).
3 Weighted average salic fraction of Archean volcanic belts, Canadian Shield (Baragar and Goodwin, 1969).

TABLE XI

IGNEOUS ROCKS OF THE LABRADOR TROUGH VOLCANIC ROCKS

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O^+	H_2O^-	TiO_2	MnO	P_2O_5	CO_2	S
							NORMAL LAVAS								
1	46.36	15.30	0.81	8.65	9.38	6.55	4.80	0.06	4.36	0.07	1.16	0.19	0.06	2.35	0.15
2	46.50	14.11	2.18	9.48	8.83	12.29	1.55	0.09	3.48	0.06	0.91	0.19	0.06	0.00	0.03
3	49.23	14.44	1.85	8.70	7.74	11.21	2.52	0.13	2.82	0.07	0.96	0.16	0.07	0.00	0.07
4	53.94	11.03	1.37	8.42	8.74	8.66	4.12	0.24	2.24	0.08	1.06	0.22	0.05	0.04	-
5	49.84	13.35	2.76	9.47	7.17	10.41	2.38	0.22	3.10	0.05	0.98	0.23	0.10	0.06	-
6	50.06	13.65	1.66	10.42	7.34	6.86	3.10	0.24	3.84	0.06	1.33	0.22	0.10	0.37	-
7	49.05	13.05	1.90	9.40	9.55	8.49	3.14	0.22	3.69	0.12	0.92	0.22	0.06	0.06	-
8	48.28	14.03	2.17	9.54	7.67	11.60	1.86	0.22	3.07	0.12	1.10	0.23	0.08	0.02	-
9	48.42	14.30	2.54	8.94	7.89	11.30	1.85	0.17	3.25	0.04	0.91	0.23	0.06	0.03	-
10	49.00	14.01	3.64	6.84	7.25	11.02	3.30	0.30	2.66	0.10	0.98	0.19	0.08	0.25	-
11	48.52	14.36	1.40	11.34	7.33	10.21	2.05	0.24	2.88	0.21	1.28	0.22	0.09	0.00	0.06
12	49.89	15.07	2.39	9.63	6.34	9.64	1.76	0.43	3.18	0.34	0.53	0.37	0.20	-	0:15
13	47.60	13.30	2.62	12.58	7.04	9.58	2.01	0.27	2.88	0.14	1.12	0.22	0.11	0.22	0.11
14	48.11	13.75	0.90	13.50	6.53	9.83	2.05	0.33	2.84	0.06	1.15	0.22	0.12	0.18	0.11
15	45.68	13.70	6.50	11.30	6.02	10.00	1.80	0.39	2.80	0.05	1.33	0.22	0.14	0.08	0.17
16	48.07	13.70	3.17	10.73	7.14	9.44	2.39	0.26	3.03	0.10	1.16	0.22	0.11	0.09	0.11
17	47.80	15.38	1.62	8.40	5.82	12.09	3.15	0.10	2.62	0.14	0.66	-	-	-	0.20
18	49.02	17.01	1.46	7.55	6.75	11.88	2.80	0.35	2.24	0.05	0.57	-	-	-	0.56
19	49.86	13.62	2.25	11.25	6.20	8.42	2.99	0.18	3.00	0.08	1.58	0.17	0.14	0.00	0.08
20	48.60	13.74	2.66	10.13	6.75	10.97	2.02	0.12	2.99	0.08	1.28	0.22	0.12	0.20	0.02
21	49.44	13.89	3.47	9.88	5.83	10.38	1.70	0.20	2.98	0.97	1.55	0.20	0.14	0.00	0.09
22	50.38	14.22	2.56	9.02	6.75	10.70	2.18	0.09	2.07	0.11	1.17	0.23	0.10	0.21	0.04
							SEWARD LAVAS								
23	49.66	13.95	11.16	3.08	3.98	4.21	4.90	2.76	2.40	0.06	1.27	0.21	0.13	1.95	0.09
24	49.92	15.47	7.87	5.33	3.42	2.79	4.80	4.30	1.96	0.06	1.38	0.22	0.14	2.12	0.09
							RHYOLITE								
25	72.3	12.6	4.7	0.6	<0.5	0.1	2.9	5.0	0.8		0.50	<0.02	0.09	<0.1	-

TABLE XI (continued).

Analyses 1-10 Central Labrador Trough. New analyses. Analyst: Quebec Dept. of Nat. Resources.

11 Menihek basalt, Central Labrador Trough (Baragar, 1967, p. 128).

12 Composite sample Willbob basalt. Analysed for W.F. Fahrig (Baragar, 1967, p. 128).

13-16 Composite samples of massive, pillowed, glomeroporphyritic, and granular basalts respectively. South central Labrador Trough (Girard, 1965).

17-18 North tip of Labrador Trough (Hardy, 1968).

19-22 Hellancourt lavas, northern Labrador Trough (Sauvé and Bergeron, 1965).

23-24 Seward lavas, southeast of Lac Musset, Central Labrador Trough. New analysis. Analyst: H. Boileau, Quebec Dept. Natural Resources.

25 Rhyolite, Menihek Lakes map-area, south-central Labrador Trough, new analysis. Geological Survey of Canada, Rapid Methods Group.

TABLE XII

IGNEOUS ROCKS OF THE LABRADOR TROUGH GABBROIC ROCKS

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O^+	H_2O^-	TiO_2	MnO	P_2O_5	CO_2	S
1	45.22	11.70	7.15	12.02	5.95	9.95	2.40	0.57	2.62	0.11	1.84	0.22	0.22	0.06	0.14
2	49.09	14.04	3.33	8.57	7.10	10.62	2.21	1.00	2.46	0.07	0.96	0.24	0.35	0.12	0.15
3	49.88	14.32	1.78	9.89	7.86	11.72	2.19	0.18	0.50	0.10	1.09	0.22	0.03	0.60	0.11
4	48.48	13.18	1.93	7.63	11.35	13.33	1.26	0.30	1.40	0.04	0.45	0.23	0.10	0.04	-
5	49.44	12.92	2.75	10.32	6.27	11.61	2.36	0.70	1.79	0.06	1.17	0.26	0.07	0.04	-
6	49.12	15.55	3.95	6.38	6.59	12.20	2.56	0.59	1.85	0.07	0.77	0.21	0.06	0.05	-
7	46.80	11.03	4.99	16.65	3.55	8.10	2.86	0.62	1.78	0.09	2.90	0.37	0.12	-	-
8	49.12	16.12	2.15	6.21	7.23	12.60	2.34	0.52	2.57	0.06	0.58	0.21	0.03	-	-
9	56.29	10.88	7.68	5.95	1.32	13.30	0.42	0.22	1.64	0.05	0.49	-	0.64	0.15	-
10	62.83	10.65	6.30	4.89	1.34	6.58	4.11	0.11	1.07	0.10	1.30	0.15	0.20	0.18	0.11
11	49.60	15.87	0.52	7.08	8.42	9.34	4.11	0.04	3.76	0.09	0.72	0.15	0.01	0.04	0.02
12	48.34	13.80	4.36	6.25	8.03	5.73	2.28	1.44	4.71	0.05	1.12	0.22	0.22	3.67	0.04
13	51.04	14.96	0.76	5.70	8.03	12.05	3.03	0.46	2.85	0.07	0.57	0.15	0.01	0.03	0.04
14	49.92	15.09	6.77	6.93	4.74	8.29	2.76	0.74	3.18	0.08	1.31	0.19	0.09	0.03	0.07
15	48.10	9.20	12.84	8.66	4.33	7.87	3.49	0.13	1.42	0.09	3.22	0.23	0.07	0.00	0.22
16	49.26	13.29	3.26	9.98	6.95	7.66	3.58	0.85	3.47	0.13	1.09	0.32	0.07	0.04	-
17	49.90	15.60	1.28	7.44	8.34	8.12	4.15	0.52	3.59	0.16	0.68	-	0.04	0.05	-
18	49.52	16.60	1.65	10.66	12.91	1.12	2.94	0.24	7.55	0.29	0.98	0.28	0.07	0.04	-
19	53.58	14.30	9.33	4.27	4.17	2.32	5.90	2.00	2.38	0.13	1.64	0.15	0.13	0.03	-
20	54.26	14.69	5.14	5.26	5.74	1.99	5.40	1.84	3.07	0.09	1.72	0.17	0.16	0.04	-
21	48.63	14.28	2.21	8.14	8.30	12.04	1.47	0.28	3.09	0.06	1.05	0.19	0.01	0.02	0.10
22	46.62	14.41	0.43	8.37	6.61	8.49	3.48	0.39	4.39	0.07	1.09	0.18	0.04	6.47	0.99
23	51.21	14.33	1.90	8.59	7.83	5.33	5.63	0.40	3.28	0.14	0.89	0.23	0.06	0.12	-
24	48.72	14.17	2.28	8.49	7.91	11.51	1.41	0.18	3.47	0.10	1.02	0.22	0.14	0.32	-
25	46.44	17.15	1.15	7.80	9.74	12.41	1.39	0.82	2.58	0.10	0.58	0.15	0.05	0.00	0.03
26	48.06	15.00	1.54	8.26	9.84	12.17	1.54	0.16	2.22	0.06	0.66	0.17	0.01	0.06	0.06
27	49.60	12.39	3.05	9.93	6.94	11.00	3.02	0.12	2.44	0.08	1.34	0.20	0.03	0.00	0.09
28	49.47	12.74	2.67	10.65	6.36	11.07	2.22	0.11	2.68	0.09	1.24	0.26	0.09	0.00	0.04

TABLE XII (continued)

29	44.36	15.80	2.35	10.21	10.81	5.65	3.23	0.31	5.14	0.21	0.76	0.31	0.04	0.23	-
30	44.63	13.85	7.36	11.16	7.17	5.46	2.84	0.47	4.83	0.14	1.85	0.32	0.14	0.08	-
31	51.28	13.39	1.75	11.20	6.50	6.92	3.41	0.65	3.24	0.04	1.35	0.24	0.14	0.04	-
32	49.82	12.72	2.97	10.96	5.92	9.89	2.80	0.12	2.84	0.07	1.48	0.26	0.11	0.00	0.03
33	66.49	9.53	2.92	8.08	1.81	4.15	3.75	0.05	1.40	0.07	0.57	0.12	0.09	0.99	0.04

Analyses 1-8 Gabbros, Otelnuk Lake, Central Labrador Trough.
9-10 Coarse-grained and pegmatitic gabbros, Otelnuk Lake.
11-18 Gabbros, Minowean Lake, Central Labrador Trough.
19-20 Gabbros, Hematite Lake, Central Labrador Trough.
21-24 Chilled contacts gabbro sills, Mistamisk Lake, Central Labrador Trough.
25-31 Gabbros, Mistamisk Lake, Central Labrador Trough.
32 Pegmatitic gabbro, Mistamisk Lake.
33 Granophyre, Otelnuk Lake, Central Labrador Trough.

Analyst: H. Boileau, Quebec Dept. Natural Resources.

Comparison with Archean Geosynclinal Belts

The Circum-Ungava and Archean geosynclinal belts of the Canadian Shield share some features that are common to many orogenic geosynclines; particularly the great thickness of predominantly magmatic in-filling and the related belts of ultramafic rocks. They also tend to occur in the form of elongated troughs although this is poorly defined in the case of the Yellowknife Belt. However, a more detailed comparison reveals glaring dissimilarities.

The sedimentary rocks of the two types of geosyncline show some outstanding contrasts. Extensive, cherty iron formation of the type found in the Circum-Ungava belt is unknown in Archean belts and dolomites and mature quartzites are rare. Great thicknesses of rhythmically graded beds are characteristic of Archean belts and uncommon in the Circum-Ungava geosyncline. The relative time sequence of the main sedimentary and volcanic successions is generally reversed between Archean and Circum-Ungava belts. In most of the Archean geosynclines, volcanism was most voluminous in the early stage of geosynclinal development whereas in the Circum-Ungava belt it culminated late in the geosynclinal history. This has resulted in further contrasts in the character of the respective geosynclinal filling: in the Circum-Ungava belt, a profusion of sills is found in the sedimentary part of the sequence, whereas in the Archean belts, the sediments contain an important volcanogenic component and have few sills.

Contrasts are even more marked between igneous rock suites of the Archean and Circum-Ungava geosynclines. Volcanic rocks of the Archean geosynclines typically evolve towards a more acidic magma, producing a spectrum of rock types that occur in about the following proportions: basalts - 60%, andesites - 28%, salic rocks - 12% (Baragar and Goodwin, 1969). In the Circum-Ungava belt the intermediate and acidic fractions are essentially missing. The very small rhyolitic pods associated with the Murdoch Formation are a negligible proportion of the total volume of volcanic rock. True andesites and dacites appear to be rare. The frequency distribution curves of colour and differentiation indexes for the Labrador Trough lavas (Fig. 19) has a spread that is closer to that of the oceanic tholeiites than to those of the two Archean volcanic belts. In the $MgO-FeO_{(tot)}$-Alkalis triangular diagram of the Labrador Trough lavas, the center of the diagram, where andesites and dacites of Archean belts normally plot is empty except for the extraordinary Seward lavas. Thus the volcanic filling of the Circum-Ungava belt is clearly distinct from that of the Archean geosynclines.

The Circum-Ungava geosyncline, in common with Alpine geosynclines shows a marked tectonic assymetry. A miogeosyncline, adjoining a well-defined craton, passes outward into a eugeosyncline accompanied by increasing intensity of metamorphism. Overturned folds and thrust faults are indicative of crustal movement towards the craton. On the other hand, Archean belts do not appear to have such systematic assymetry in either their structures, their contents, or their metamorphism. This contrast suggests that the two types of geosynclines are products of fundamentally different tectonic actions. In the Circum-Ungava geosyncline we can recognize the Alpine pattern of mountain building; the Archean belts may not have a recognizable Phanerozoic counterpart.

One further distinction should be noted in the case of the Circum-Ungava geosyncline although it does not fall strictly within the province of this paper. Syn- and post-orogenic granitic intrusions so characteristic of most orogenic belts, particularly those of the Archean, appear to be essentially lacking in the Circum-Ungava belt. Small pegmatitic and granitic intrusions ranging from a few tens of feet to 3 or 4 miles long are commonly found in the metamorphosed distal margins of the geosyncline - the east side of the Labrador Trough and the north side of the Cape Smith Belt (Bergeron, 1959; Fahrig, 1965; Gélinas, 1960, 1962; Gold, 1962; Sauvé, 1957) - but granitic intrusions of stock or batholithic proportions are lacking. Is it possible that there is a relationship between the paucity of intermediate and acidic volcanic rocks within the Circum-Ungava geosyncline and the lack of significant granitic intrusions later in the orogenic cycle? These two features are the outstanding abnormalities of the Circum-Ungava belt and it seems logical to assume that they might be related.

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Comments by: A. Ruffman, Dalhousie University, present address: Bedford Institution

The Labrador geosyncline appears to be developed on top of the buried junction between the Superior and Churchill crystalline basements (based on K-Ar dates). The same statement applies to the Cape Smith-Wakeham Bay Belt. Could you please comment on the nature of the tectonic event that imparted the 1.7 b.y. age on the crystalline basement to the east and north of the above geosynclines.

Response by: the authors

The regions presently underlain by the Labrador Trough and Cape Smith Belt were not, of course, the junction between tectonic provinces when these rocks were deposited, but were uniformly Archean terrains. However, we consider the Circum-Ungava geosyncline to have been an incipient orogenic zone from the start, just as in the case of the Alpine geosyncline. Its position adjoining the foreland of the present Hudsonian orogenic belt would not, in this view, be regarded as accidental, but would be the surface manifestation of an early phase of Hudsonian orogenesis. Possibly the same is true of the Superior-Churchill boundary in Manitoba, but, unlike Labrador, whatever deposits may have formed there have since been removed.

Comments by: A. Ruffman

Why have you indicated that the rocks of the Belcher Arc Homocline are found beneath the Paleozoic of western James Bay and the eastern Moose River basin?

Response by: G.D. Jackson

This interpretation is postulated on the basis partly of Bostock's work as reported in GSC Paper 68-53 and largely on the tenuous evidence provided by two drill holes (Satterly, 1953, 50°20'N, 81°50'W; Hogg *et al.*, 1953, 51°45'N, 80°40'W).

The term homocline is a structural term that, as defined in the AGI glossary probably fits only a narrow band of strata outcropping along the eastern and southern margins of the Belcher Basin.

THE HURONIAN SUPERGROUP NORTH OF LAKE HURON

M.J. Frarey
Geological Survey of Canada, Ottawa

and

S.M. Roscoe
consulting geologist, Ottawa

Abstract

The Huronian Supergroup of earliest Aphebian age occupies a remnant fold belt about 200 miles long by up to 35 miles wide north of Lake Huron. The supergroup, which differs in several respects from other Aphebian successions of the Canadian Shield, comprises up to 40,000 feet of sediments and volcanics dominated by coarse clastic debris deposited in a high-energy environment, considered for the most part to be alluvial. The Huronian succession is especially noteworthy for its pyritic, uraniferous clastic beds and for its three till-like conglomeratic formations.

The Huronian deposits fall into four stratigraphic groups which in order upwards are the Elliot Lake, Hough Lake, Quirke Lake, and Cobalt Groups. The latter three contain major depositional cycles beginning with paraconglomerate beds, largely marine, followed by fine-grained marine strata and then by cross-bedded coarse arenites. These groups and cycles are separated by breaks in sedimentation and local unconformities produced by positive movements of the trough floor. Prior to Cobalt deposition, a major transgression produced an important regional extension of the sedimentation area northeastward into Quebec and a broad stratigraphic overlap of the Gowganda Formation onto the Archean basement along the north edge of the fold belt.

Paraconglomerates of the Ramsay Lake, Bruce, and Gowganda Formations are considered to be of glacial origin because of their extent, thickness, and associated deposits bearing rafted clasts. They are probably in large part glaciomarine. Tillitic deposits in the Chibougamau district of Quebec and in the District of Keewatin may also be products of this early Aphebian glaciation, suggesting that it was of continental proportions.

Because of their drab colour, their detrital pyrite and uraninite content, the nature of their underlying paleosol, and their lack of iron oxide minerals, the three lower Huronian groups are interpreted as having been deposited under anoxygenic conditions, prior to the accumulation of free oxygen in the earth's atmosphere. The Cobalt Group contains iron oxides and a few red beds suggesting that free oxygen had begun to accumulate at that time.

Huronian sedimentation was dominated by the deposition of coarse, cross-bedded, fluvial sands to form very extensive southward-thickening wedges that must have extended to the south well beyond the present limit of exposure. Probably 20,000 cubic miles of quartzite was deposited just in the Cobalt Group of Ontario, indicating the vast amount of uplift and erosion that occurred in the provenance area of Superior province to the north during early Aphebian time. If the area of Huronian deposition continued northeastward to the Chibougamau and Mistassini districts of Quebec, a major crustal depression some 750 miles long existed in earliest Aphebian time approximately along the present southeastern edge of the Superior province. Alternatively, there may have been a number of separate though similar marginal basins and successions near the edge of the Archean protocontinent, of which the Quebec and District of Keewatin rocks are remnants.

INTRODUCTION

The Huronian Supergroup, characterized by thick formations of coarse, cross-bedded, quartzose clastic sediments, was deposited in non-marine to marginal marine environments and differs in this respect from other Aphebian successions in the Canadian Shield which were deposited in marine to marginal marine environments. Other distinctive features of the Huronian sequence include: the presence of pyrite in coarse clastic sediments and the absence of iron oxides and red coloration except in the upper part of the succession; the existence of thick, extensive lenses of extremely coarse sand or fine gravel which contain radioactive quartz pebble conglomerate beds including the pyritic, uraninite-bearing ores of the Elliot Lake area; the occurrence of three widely-separated formations comprising or containing till-like conglomerate, - the uppermost of which - the Gowganda Formation - is no less than 4,500 feet thick in some places.

Huronian strata were deposited about 2.3×10^9 years ago and are older than other Aphebian strata whose age can only be bracketed between the wider limits of 1.7 and 2.5×10^9 years. Some of the peculiar features of the Huronian Supergroup may be related to factors attendant upon its oldest Proterozoic age, including: the evolution of the atmosphere (at about that time) from one lacking oxygen to one containing free oxygen; periodically frigid climatic conditions that did not recur later in Aphebian time; and, perhaps, unique tectonic developments that immediately followed the widespread, continent-building Kenoran orogeny. Clearly, it is important to speculate upon these matters and continue studies of the Huronian rocks in search of links between Archean history and the patterns of tectonic activity and sedimentation that have prevailed on the shield since mid-Proterozoic times.

In first examining the rocks of the Southern Province, Alexander Murray and William Logan recognized along the north shore of Lake Huron an assemblage of rocks consisting of sandstones, conglomerates, slates and limestones, and numerous intercalations of "trap" and greenstone (Logan, 1849). They also observed that this assemblage rested with great unconformity on older granite and other crystalline rocks, and that it was overlain above a second great unconformity by Paleozoic fossiliferous strata which Murray had traced from southern Ontario. In succeeding years this stratified clastic assemblage was further described and mapped, and appeared in the 1863 "Geology of Canada" as the "Huronian series", in keeping with the nomenclature style of the day. Thus the Huronian has great historical significance as the first valid stratigraphic subdivision within the Canadian Shield, or probably anywhere in Precambrian terrains for that matter. The great impact of this resulted in early workers in the Precambrian far and wide applying the name "Huronian" to various little-metamorphosed stratified sequences that were believed to overlie crystalline basements.

Geological study of the Huronian has taken place in three general stages: the era of Murray and Logan, that of W.H. Collins and associates, between 1908 and 1925, and the present stage beginning about 1953. All have had strong economic overtones. The first originated, at least in part, because of known copper deposits at Bruce Mines; in the second, Collins' overall objective was to correlate the rocks between the Cobalt silver camp, the Sudbury nickel-copper field and the copper area at Bruce Mines; and the more recent studies were closely linked to the occurrence of and exploration for uranium orebodies at Elliot Lake and elsewhere. While some of the mineral deposits are partly in younger intrusions, the mineral wealth of the Huronian belt probably exceeds that of any other belt of equivalent size.

This paper reviews some aspects of the Huronian basin, its contained strata and their special features, interprets environments of deposition of the supergroup, and speculates on the extent of the original area of deposition. Post-depositional history is not included. The paper is based mainly on work in what is called herein the main fold belt, that is, the area adjacent to Lake Huron, where most of the investigations have taken place since the Collins era.

DISTRIBUTION

The Huronian Supergroup occurs in a northerly-concave arcuate fold belt, 200 miles long and about 40 miles wide, in the Southern province, along the north shore of Lake Huron. This fold belt is truncated to the east by the Grenville province, and its extension to the west and south is concealed by Paleozoic cover rocks and waters of Lakes Superior and Huron. An irregular area that extends some 100 miles in the Superior province beyond the northeastern part of the main fold belt contains gently folded strata of the Cobalt Group which forms the upper part of the Supergroup.

The Supergroup, as much as 40,000 feet thick in the area between Sudbury and Lake Huron, thins northward within the main fold belt. Much of this thinning occurs in clastic units that also coarsen northward; some is due to northerly wedge-outs of basal units and some to unconformities within the succession.

The Chibougamau Group and the Otish Mountain Group 250 to 500 miles northeast of the Huronian belt comprise thick formations that are remarkably similar to those of the Huronian Cobalt Group. If these rocks are correlative, Cobalt sediments including many thousands of feet of quartzite may have been deposited through an area of hundreds of thousands of square miles, extending more than 750 miles in a northeasterly direction. There is no evidence to suggest that the lower Huronian Groups, were ever so extensive but neither is there any reason to believe that they were deposited only within the main fold belt where they are found today.

Rocks resembling the Huronian tillites (or supposed tillites), quartzites, and radioactive pyritic quartz pebble conglomerate are found underlying the Hurwitz Group in a small area in the southern District of Keewatin.

AGE

On purely stratigraphic grounds the age of the Huronian can only be bracketed between the Archean and the intrusion of diabase bodies that until recent years were widely but erroneously interpreted as Keweenawan. Isotope dating has provided a much closer time-reference that is also much more meaningful in terms of interpreting Precambrian history. The dating of post-Huronian diabase intrusions at approximately 2,190 m.y. by the Geological Survey of Canada (Lowden *et al.*, 1962), by W.R. Van Schmus (1965), and more recently by the Massachusetts Institute of Technology (Fairbairn *et al.*, 1969) has been one of the most significant advances emerging from modern studies in the Huronian belt, and recent dating of Huronian (Gowganda) sediments by the M.I.T. group (Fairbairn *et al.*, 1969) at 2,300 m.y. evidently represents further refinement and confirmation. The allocation of Huronian deposition to earliest Aphebian time, which deposition we believe may well reflect special events and special conditions, illustrates the great value of isotopic dating, notwithstanding its imperfections, in deciphering Precambrian history, and points up the need for similar work in other belts of the Shield.

STRATIGRAPHY

Huronian Groups

The Huronian succession - first divided into lower and upper Huronian series, then Bruce and Cobalt series - is now considered to contain four groups. In ascending order these are: the Elliot Lake Group, the Hough Lake Group, the Quirke Lake Group, and the Cobalt Group (Fig. 1).

Elliot Lake Group

The Elliot Lake Group contains volcanic rocks - the Thessalon, Stobie, and Copper Cliff Formations, feldspathic quartzite - the Matinenda and Livingstone Creek Formations, and argillaceous sediments - the McKim Formation. Successions within the group differ in various parts of the belt. Volcanics are widespread areally[1], but are most abundant at the western and eastern ends. In the central part, near Elliot Lake and Blind River, the Matinenda Formation constitutes all or almost all of the group. It nonconformably overlies an ancient weathered surface developed atop Archean granitic, metavolcanic, and metasedimentary rocks and is more than 1,000 feet thick near Blind River, 700 feet near Elliot Lake, and 300 feet at Quirke Lake, where it contains the important uranium-bearing, ore-bearing, pyritic, sericitic quartz-pebble conglomerate beds. In some places, the subarkose formation can be subdivided into coarse, poorly-sorted, conglomeratic members and finer-grained, better-sorted members. Lesser subdivisions separated by distinctive rocks such as conglomeratic zones can be recognized locally.

In the Sudbury area, three thousand or more feet of argillaceous quartzite, siltstone and argillite of the McKim Formation overlie concordantly several thousand feet of volcanic rocks. These latter include the Stobie Formation, formed mainly of basalt, and the thinner, less extensive, overlying Copper Cliff Formation composed of rhyolite. The question of whether the volcanic rocks and concordantly overlying sediments are Proterozoic (Huronian) or Archean has long been debated because the basal contact of the volcanics is obscured by intrusives and no unconformity has been found between them and undoubted Archean rocks. In this paper they are considered to be Huronian. The change in sedimentary stratigraphy of the Elliot Lake Group between the Sudbury and Elliot Lake areas can be summarized as follows: a wedge of Matinenda subarkose extends eastward toward Sudbury between the volcanics and the McKim Formation, and a tongue of McKim argillite extends westward to Elliot Lake and Blind River, where it locally overlies the Matinenda Formation.

In the western part of the belt, between Thessalon and Sault Ste. Marie, the Elliot Lake Group comprises as much as 3,500 feet of basaltic flows - the Thessalon Formation - overlying as much as 1,500 feet of subarkose - the Livingstone Creek Formation. In places the latter formation is thin or absent. Radioactive, pyritic, quartz-pebble conglomerate beds are interlayered with some of the basal flows of the Thessalon Formation and also within the Livingstone Creek Formation, much of which is not unlike the Matinenda Formation. Precise correlations or age relationships cannot be determined at present between the Livingstone Creek and Matinenda Formations, or between the Thessalon Formation and the Stobie Formation, but all appear to belong within the Elliot Lake Group (*see* Fig. 1).

Hough Lake Group

The Hough Lake Group includes the Ramsay Lake Formation - conglomeratic greywacke up to 600 feet thick in the eastern part of the belt but only a few tens of feet thick near Elliot Lake, the Pecors Formation consisting of siltstone, argillite, and fine to medium-grained quartzite, together up to 3,500 feet thick southwest of Sudbury, and the Mississagi Formation, 1,500 to 4,000 feet thick through most of the belt. The Mississagi Formation[2] consists of coarse-grained, feldspathic, sericitic quartzite. It is thickly cross-bedded, much less commonly ripple-marked. It is particularly coarse-grained in its

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[1]Salmay Lake, Baldwin, Agnew Lake, and Pater are other informal names applied locally to volcanics of the Elliot Lake Group between Blind River and Sudbury.

[2]Includes former Wanapitei Formation of the Sudbury area.

most northerly outcrops where it is very similar to the Matinenda Formation. This similarity includes the presence of radioactive, pyritic quartz-pebble conglomerate. Layers of small quartz, chert, and jasper pebbles are common elsewhere in the formation.

In the northerly parts of the belt, the Ramsay Lake Formation rests directly on the Archean rocks, and in such areas, the Hough Lake Group oversteps the northern margin of the Elliot Lake Group. This northerly margin was probably near the northern depositional limit of the Elliot Lake Group, but it also represents an erosional margin. In the Quirke Lake area, there is clear evidence of an important disconformity, possibly even a low-angle unconformity, between the Elliot Lake Group and the Ramsay Lake conglomerate[1]. This conglomerate contains unsorted clasts, up to boulder size, of grey granite, greenstone, and other rocks found in the Archean terrain of the region. The matrix varies from quartzose greywacke to siltstone. The conglomerate is not unlike some found in the Gowganda Formation, and like the latter may be glacial in origin as discussed later.

Quirke Lake Group

The Quirke Lake Group comprises, in ascending order, the Bruce Formation consisting of conglomeratic greywacke, the Espanola Formation containing limestone, siltstone, argillaceous quartzite and dolomite, and the Serpent Formation made up of arkose and feldspathic quartzite. Near the east end of the fold belt, the Serpent also contains thin limy layers at the top. At Quirke Lake, these formations are 250, 600, and 700 feet thick respectively, and the total thickness of the group is about 1,500 feet. In contrast, it is as much as 5,500 feet thick south of Sudbury. In that area, the Serpent Formation thickens more markedly than other formations of the group. The Bruce Formation is similar in its till-like aspect to other conglomeratic greywackes in the Ramsay Lake and Gowganda Formations, but has a somewhat more quartzose and pyritic matrix. It is relatively uniform throughout the belt. In places there is evidence that the Mississagi Formation was weathered and eroded prior to deposition of the conglomerate. The Espanola comprises a dominant middle siltstone member, which exceeds a thousand feet in thickness in places, underlain and overlain by much thinner limestone and dolomite layers. These basin-wide carbonate members, the only ones in the entire Huronian sequence, contain no stromatolites. South of Espanola, a fourth member made up of 500 feet or more of sandstone, in part calcareous, occurs above the siltstone member. Like the Hough Lake Group, the Quirke Lake Group overlaps onto the Archean basement in places along the northern edge of the belt.

Cobalt Group

The Cobalt Group is much thicker and more extensive than the other groups. It comprises in order upwards the Gowganda, Lorrain, Gordon Lake, and Bar River Formations. The Gowganda Formation is a heterogeneous assemblage of greywacke, conglomerate, argillite, impure quartzite and arkose. The Lorrain consists of coarse-grained, cross-bedded arkose grading up to orthoquartzite, with fine-grained orthoquartzite occurring higher in the formation. The Gordon Lake Formation consists mainly of varicoloured laminated siltstone, and the Bar River Formation comprises orthoquartzite and siltstone. Like units lower in the succession, these formations thicken along the fold belt from west to east and from north to south. Thus the Gowganda is less than a thousand feet thick

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[1]According to J.A. Robertson (personal communication) local radioactivity in the Ramsay Lake Formation is spatially related to, and derived from, bevelled ore horizons of the underlying Matinenda Formation, indicating the early origin of the ore.

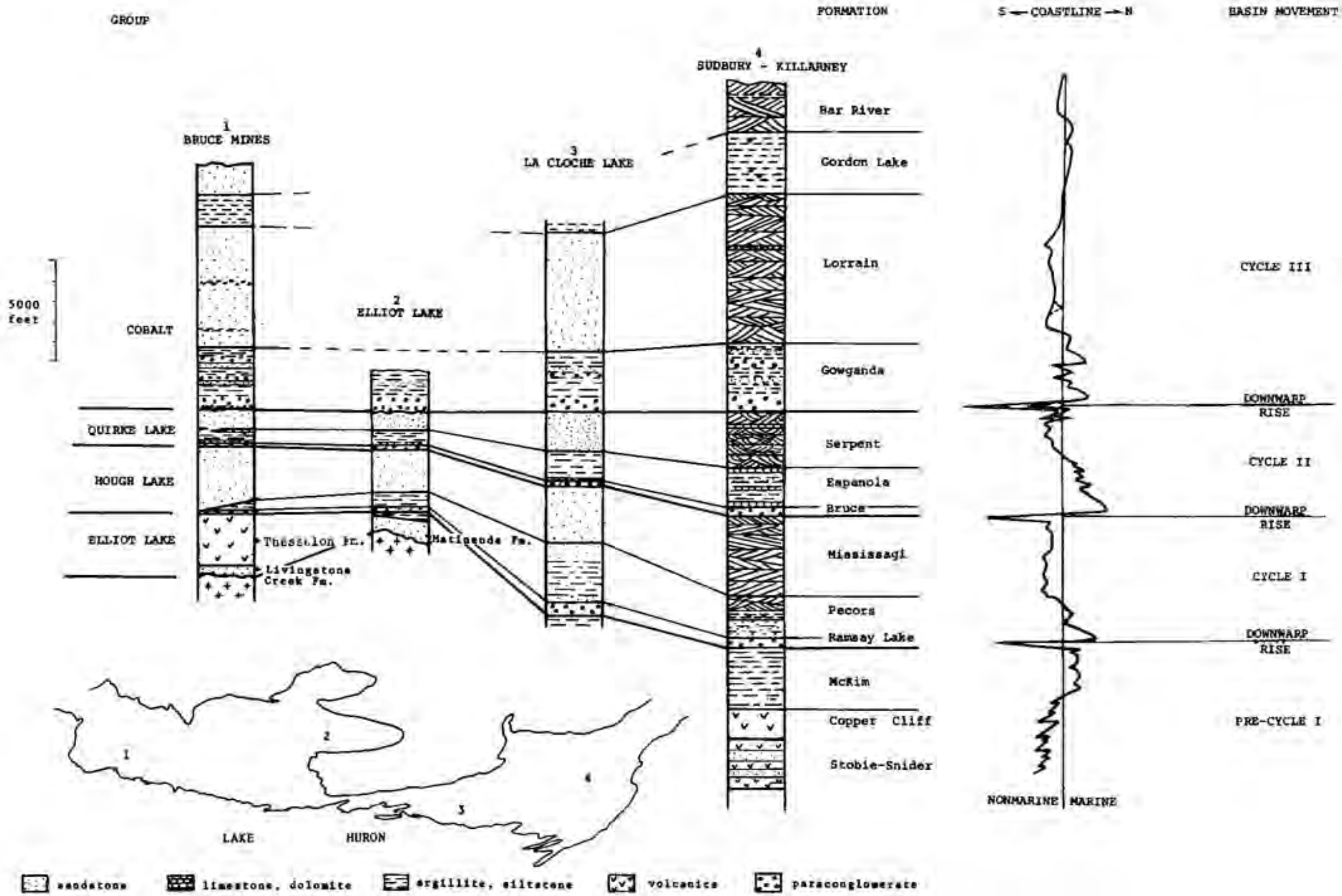


Figure 1. Selected stratigraphic sections between Sault Ste. Marie and Sudbury, Huronian cyclic deposition, and interpreted environments and movements (Section 3 courtesy F.W. Chandler and J.A. Robertson, Ontario Dept. of Mines).

at the north edge of the belt near Flack Lake, about 3,000 feet thick around Bruce Mines near the west end, and as much as 4,000 feet thick near the east end. The Lorrain is perhaps 3,000 feet thick at the first locality, 6,000 feet near Bruce Mines, and over 7,000 feet northeast of Killarney. The Gordon Lake is 1,000, 1,500 and 3,000 feet thick at these respective localities, and the Bar River, which nowhere presents a complete section, has about the same thickness. The group thickness is about 11,000 feet at Bruce Mines and 16,000 feet near Lake Panache south of Sudbury. Northeast of Sudbury in the type Cobalt district, erosion has removed most of the upper two formations, but in places imposing thicknesses of Gowganda, Lorrain and Gordon Lake have been recorded, for example in an area some 50 miles northeast of Sudbury where they attain 3,000 feet, 5,000 feet, and 2,000 feet respectively (K.D. Card, 1970, pers. comm.). Evidently deposition of the Cobalt Group was virtually continuous, with abrupt transitions from one formation to the next. The group everywhere overstepped the older groups onto the basement, the preserved overlap being about 15 miles wide north of Blind River and 80 miles northeast of Sudbury.

Cyclic deposition and relationships between groups

The Hough Lake, Quirke Lake and Cobalt Groups each show a similar upward trend from paraconglomerate to fine-grained sediments to cross-bedded arenites, as can be seen in Figure 1. The Ramsay Lake conglomerate is overlain by the argillaceous rocks of the Pecors Formation, grading up through the Mississagi quartzites. The Bruce Formation, in turn, is abruptly overlain by the thin-bedded limestone and siltstone of the Espanola Formation and then by the cross-bedded quartzite of the Serpent Formation. The cycle is repeated on a much larger scale in the Cobalt Group where the paraconglomerates and arkosic beds are succeeded by abundant argillite in the upper Gowganda, followed by coarse cross-bedded arenites of the Lorrain Formation.

Actual breaks in sedimentation appear to have occurred in places between these cycles and there is local evidence of angular unconformity developing prior to deposition of the succeeding cycle. A local low angular unconformity appears below the Ramsay Lake at Quirke Lake, where the conglomerate truncates the ore-bearing reefs of the Matinenda Formation. Between the Mississagi and Bruce Formations, an erosional break has been reported at a number of places by J.A. Robertson of the Ontario Department of Mines and others; locally, suggestions of slight angular discordance between them have also been observed. Post-quartzite uplift and erosion is, however, most conspicuous at the Quirke Lake-Cobalt boundary. This boundary separated the original upper and lower Huronian series of Murray, and the Cobalt and Bruce series of Collins. Collins (1925) considered that there was an erosional break of major significance throughout the belt and also suggested that angular unconformity existed in places in the Blind River area. He also stressed however, that in a great many places there was concordance, gradation or interbedding at this contact, suggesting conformity. From subsequent work this latter situation appears to apply mostly to areas south of the Murray Fault, whereas north of the fault, between Elliot Lake and Sault Ste. Marie, there is conspicuous disconformity. Over much of this distance the Cobalt Group overlies and includes fragments of the lower limestone member of the Espanola Formation, but in places it lies on the Bruce Formation, and north of Thessalon on the Mississagi Formation of the Hough Lake Group. In this region a well-sorted, polymict, roundstone conglomerate a few tens of feet thick commonly forms the base of the Cobalt Group.

Cyclic deposition and other variations within the Huronian Supergroup are interpreted at the extreme right in Figure 1. Abrupt rises terminate cycles and are immediately followed by sharp downwarps and deposition of conglomeratic greywacke, and subsequent deposition of fine-grained sediments in the next cycle. This in turn is followed by sudden uplift throughout the source area to produce cross-bedded fluvial quartzites.

EARLY APHEBIAN GLACIATION

Since the turn of the century, paraconglomerates and other rocks of the Huronian basin, particularly those of the Gowganda Formation, have been attributed a glacial origin. If glaciation was widespread in earliest Aphebian time, the recognition of glacial sediments assumes importance in making valid correlations between widely separated parts of the Canadian Shield or possibly between shields. Continued study of the Huronian deposits should contribute substantially toward this objective. Huronian glaciation has had both its proponents and its sceptics however, and a few workers have proposed other origins for paraconglomerates of the supergroup. Discussion here is mainly confined to a brief review of observed features of these rocks and their significance.

Table 1 lists a number of megascopic features and stratigraphic relationships of the three formations best showing glacial affinities, i.e. the Ramsay Lake, Bruce, and Gowganda Formations.

Table 1: Field characteristics of Huronian paraconglomerates.

	RAMSAY LAKE	BRUCE	GOWGANDA
MATRIX	Quartzose greywacke, pyritic	Coarse quartzose pyritic greywacke; calcareous at the top in eastern zone	Silty to very coarse greywacke, locally pyritic or quartzose
CLASTS	Dispersed grey granite, quartz and greenstone; mostly pebble-size; may be very sparse	Dispersed grey granite and quartz; 20 per cent of the rock, mostly pebbles	Dispersed, mainly pink granite; percentage highly variable; size from granules to boulders
FORM AND THICKNESS	Basin-wide sheet, 0 to 200 feet in west, up to 600 feet or more in east	Basin-wide sheet; may have two divisions in places; 50 to 600 feet thick	Individual sheets of limited lateral extent up to hundreds of feet thick, intercalated with arkose, quartzite and argillite
INTERNAL STRUCTURE	Massive with rare sorted interbeds	Massive; local interbeds at base	Massive to bedded; thin arkose and argillite interbeds common in west; slump structure common
THICKNESS VARIATION	With exceptions, varies sympathetically with other formations	As in other formations	As a formation, corresponds with other units
CONTACT RELATIONS	Base, conformable to disconformable, top conformable. Paraconglomerate in places overlain by argillite with dropstones	Base conformable to disconformable, top conformable. Paraconglomerate locally overlain by beds with dropstones	Base conformable to disconformable, top conformable. Paraconglomerate commonly intercalated with argillite carrying dropstones
SOURCE	North	North	North

The three paraconglomerates are somewhat similar in the field, but the more chloritic matrix and characteristic pink granite clasts usually enable one to identify Gowganda outcrops with confidence. As basin deposits, these formations were each laid down co-extensively on the subjacent, largely unconsolidated formations, with gradational contacts in places. Occurring consistently in the cyclic pattern and in thickness trend, they should be viewed as integral if abnormal components of the basin-filling process, and not as extraneous deposits of unrelated, independent origin.

Most recent workers appear to favor a glacial rather than a mudflow origin for these deposits. Casshyap (1969) has interpreted the Bruce and Gowganda paraconglomerates as terrigenous moraine, but from their continuity and thickness a glaciomarine origin appears preferable at least for the Bruce, and this applies also to the Ramsay Lake Formation. The carbonate content in the upper part of the Bruce, and possibly its pyrite content, support this idea. In the case of the Gowganda, evidence for glacial origin from the paraconglomerates themselves is less compelling, and at least one traditional glacial criteria, i.e., striated clasts, has been found to be equivocal. The most convincing evidence in this formation for extensive glacial activity consists of the abundant exotic clasts up to boulder size found in laminated to varved pelitic beds. These of course are very widely and strikingly displayed in the Gowganda, so that a glacial, or at least frigid regime remains established. A glaciomarine origin has been proposed for a large area southwest of Sudbury, and this may well be true of at least parts of the formation throughout the fold belt. Traditionally, the entire Gowganda Formation was assigned a glacial origin and the literature is replete with references to the "Gowganda tillite", a somewhat misleading expression. Actually, aside from the argillites with dropstones, the material in the formation could be accounted for by non-glacial sedimentary processes.

If the glacial origin of the Ramsay Lake, Bruce, and Gowganda beds is accepted, extensive ice sheets must have covered the craton north of the Huronian belt, at least intermittently, from near the onset of Huronian deposition to the time of Gowganda deposition. The thickness of the deposits between episodes of glacial deposition in the basin indicates long periods of ice retreat from the source area.

If the tillitic beds of the Chibougamau district and of southern Keewatin are time equivalents of the Huronian rocks, the possible extent of the ice sheet is considerably increased and approaches that of Pleistocene glaciations. One would expect that a glacial event of this magnitude would appear in the record of other Precambrian terrains of the northern hemisphere.

EVIDENCE IN HURONIAN ROCKS FOR A CHANGE FROM A REDUCING TO AN OXIDIZING ATMOSPHERE

It is generally believed that the Earth's atmosphere evolved from an anoxygenic (or reducing) state consisting of gases such as CO, CO_2, N_2, H_2, H_2O, but no free oxygen (*see* Rubey, 1955). There is a reason to suppose moreover, that life initially developed at a time when the atmosphere lacked free oxygen (*see* Cloud, 1968). Oxygen is produced by photo-dissociation of water vapour in the upper atmosphere and by photo-synthetic dissociation of water and carbon dioxide by plants. Reduced material on the Earth's surface and in continually emitted volcanic gases would have consumed oxygen produced by the former process during early Earth history. Build-up of free oxygen in the atmosphere would have had to await the evolution and proliferation of blue-green algae. This transition from a reducing to an oxidizing atmosphere must have been accompanied by major changes in chemistry of surficial waters, in weathering processes, in dissolved material in waters, in geologically important organic processes, and in sedimentation. Differences in character and relative abundances of various sediments in Archean rocks, Aphebian rocks, and post-Aphebian rocks are particularly striking and some of these are likely due to such changes. The relatively well-dated Huronian rocks, which differ from other Aphebian sequences and probably antedate most of them, may provide a particularly important clue to the timing and character of these atmospheric changes.

An ancient weathered zone is found on the Archean surface beneath fluvial Huronian clastic sediments and volcanics. This zone resembles younger soil profiles texturally, mineralogically, and chemically, in many respects. Pink granite, for example, grades upwards - through progressive solution and alteration of mafic minerals, plagioclase, and potash feldspar to argillaceous material - into a greenish rock containing grains of quartz and microcline and then into one containing only scattered grains of quartz floating in a clay matrix (now fine-grained sericite). The ancient weathering resulted in leaching of CaO, SrO, MnO_2, MgO, Na_2O, and enrichment in water, Rb_2O_3, and K_2O, which could be considered normal. Iron however is lost rather than accumulated, and the ferric:ferrous ratio decreases rather than increases. A number of geologists (Pienaar, 1963; Robertson, 1963; and Roscoe, 1969, p. 72) have considered that these changes indicate that ground water and atmosphere lacked free oxygen at that time. Other explanations, which may apply in cases where red ferruginous zones are lacking at unconformities beneath marine sediments or beneath sediments containing abundant organic material, do not appear to be relevant here. It is interesting that enrichments of ferric iron are found at unconformities beneath some sequences of younger non-marine Proterozoic rocks (e.g. Donaldson, 1969).

Most of the Huronian rocks are drab-coloured. This is anomalous as the sediments were derived from weathered terrains, extensively sorted in shallow waters, deposited in large part under fluvial conditions without any associated carbonaceous material, and periodically exposed to the atmosphere as shown by dessication cracks in some formations. Such sediments would normally include conspicuous red beds. Reddish colouration is lacking in the lower three groups, apart from occasional red rock clasts, and is not characteristic of the uppermost Cobalt Group. Pyrite rather than hematite occurs in these rocks but the total iron content is low. Hematite and some reddish units are present in the Cobalt Group, notably in the Gordon Lake Formation which contains variegated siltstone and in the Lorrain Formation which is hematitic. The reduced character of the Huronian sediments cannot be attributed to rapid deposition, to carbonaceous material, or to metamorphism and it has therefore been considered as evidence for an anoxygenic atmosphere (Roscoe, 1969, p. 80). The appearance of hematite in the Cobalt Group suggests that the Huronian succession records a change 2.3×10^9 years ago from an atmosphere lacking oxygen to one containing a trace of free oxygen.

Evidence that detrital uraninite and pyrite occur in Huronian quartz-pebble conglomerate beds has been presented by Ramdohr (1958a) and by Roscoe (1969, pp. 128-136). Although these minerals have been found as detrital grains in recent stream deposits, they are unstable under oxidizing conditions. Their abundance in widespread conglomerate beds in the Huronian belt indicates that the atmosphere must have lacked free oxygen at the time the beds were deposited and buried. The occurrence of more "normal", hematite-bearing, black sand placers in the Lorrain Formation of the Cobalt Group is consistent with the conclusion, previously stated, that free oxygen first began to accumulate in the atmosphere at the time the Cobalt Group was deposited, 2.3×10^9 years ago. There is no evidence that the several known other occurrences of extensive pyritic conglomerates such as the Witwatersrand, Serra de Jocabina, or Padlei N.W.T., are younger than 2.3×10^9 years.

The evolution of the Earth's atmosphere and the time at which free oxygen began to accumulate are extremely important considerations not only with respect to non-reversible changes in patterns of sedimentation but also with respect to formation of some important types of ore deposits.

An anoxygenic atmosphere would permit sulphide minerals to form either biogenically or abiotically in subaerial or near-surface environments where they could not have formed during subsequent geological history. Perhaps this is one reason why vulcanogenic base metal sulphide deposits are relatively abundant in Archean rocks. Uraniferous conglomerates, like those in the Huronian rocks, could not have been deposited prior to the unroofing of an extensive, uraniferous, granitic source terrain at the end of Archean time and, furthermore, they could not have been formed after the atmosphere became oxidizing and after uranium-rich minerals became unstable in soils and surficial waters.

The early anoxygenic atmosphere permitted large amounts of iron to be supplied to the seas in the ferrous state - so that it was available to form concentrated chemical sediments. In later times, iron has been carried largely in suspension in the ferric state and generally became dispersed in fine-grained sediments. The unique abundance, thickness and extent of hematite iron-formations in Aphebian rocks which are probably younger than the Huronian Supergroup, may be related to the initial development of free oxygen throughout the atmosphere. This oxygen was produced by proliferation of blue-green algae which may have entrapped carbonates, thus explaining the sudden appearance of thick formations of stromatolitic dolomites associated with jasper hematite iron-formations in many Aphebian successions.

PATTERN AND EXTENT OF HURONIAN SEDIMENTATION

The general pattern of tectonic activity and sedimentation that prevailed during deposition of the Huronian Supergroup was one of uplift, weathering and erosion of a source area that shed sediments southward onto a subsiding area that may have been an unstable platform, a basin, or a trough. Topography was rugged locally at the time that the Gowganda Formation was being deposited, but the lack of immature, sorted detritus elsewhere in the Huronian succession indicates that the source area had gentle relief.

A hinge-line between positive and negative areas passed 15 to 50 miles north of the present shoreline of Lake Huron. It had a northward convex, arcuate trend in this area, corresponding to trends of structures such as the Murray and Flack Lake Faults, which display post-Huronian movement but may have been formed earlier. This line migrated northward during deposition of successive Huronian groups and marks the northern paleolimit of deposition or preservation of these groups. Each group includes sediments deposited in quiet water but large parts of them are fluvial sediments, so that the hinge line generally marks the northern edge of an area of continental sedimentation rather then the margin of a shallow marine basin or trough.

Huronian sedimentation was dominated volumetrically by spreading of thick to very thick, submature to mature arenite sheets in a high-energy environment, presumably a great alluvial plain. These were deposited cyclically after abrupt downwarps of the basin and deposition of paraconglomerates and fine-grained sediments in a marginal marine environment. The lack of carbonate deposits is a distinguishing feature of the Huronian succession, these being essentially confined to the Espanola Formation. The expansion of the basin in Cobalt time was much the most pronounced of the succession of northerly shifts during sedimentation. Following Gowganda deposition, prolonged source-area uplift and erosion ensued, during which an enormous amount of weathered debris was swept from the source areas. The coarsest and most resistant portion of this was laid down in the Lorrain, and in the Bar River Formations.

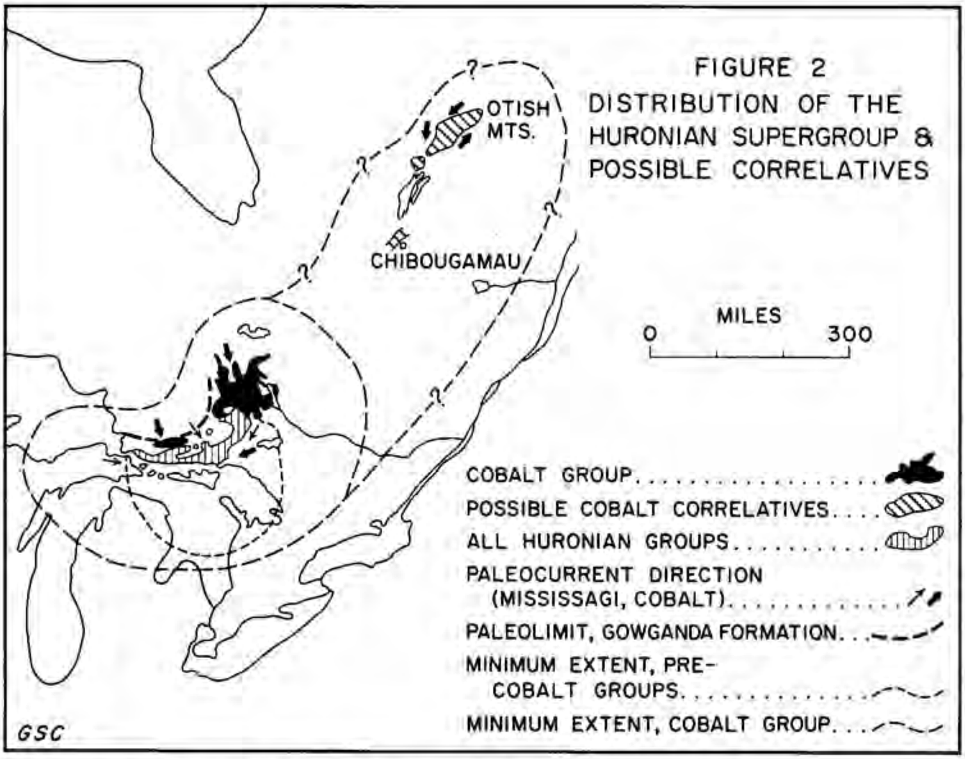
The fluvial formations of each group thicken southward and consist of coarse, cross-bedded quartzites containing variable amounts of feldspar and interstitial sericite (originally clay). As presently preserved, these formations cannot contain more than a small fraction of the total material that was eroded from Archean rocks during the interval when the sands were deposited. Most of the calcium, magnesium and iron were carried seaward in solution. Clay and silt would be the major detrital materials derived from the weathering of granitic source rocks as well as from Archean metavolcanic and metasedimentary rocks. Moreover, streams must have carried large amounts of fine to medium-grained sand-sized particles southward, beyond the present belt of preserved Huronian rocks. The arenaceous formations thus represent the coarsest-grained, sourceward parts of clastic wedges that must have extended far to the south.

The original volume of Huronian sediments, and of Archean rocks eroded to form them, must have been very great. The Cobalt Group alone contains more than 20,000 cubic miles of quartzite within Ontario, and it likely extended 500 miles farther northeast into central Quebec. If one reconstructs all of the major Huronian clastic wedges, as well as possible, and adds other important formations - McKim, Pecors, Espanola, Gowganda - to these, it becomes apparent that the total volume of Archean rocks eroded during Aphebian time may have exceeded 1,000 cubic miles per mile of length of Huronian fold belt. If we accept a correlation and continuity of the thick Chibougamau and Otish Mountain Groups with the Gowganda and Lorrain Formations in Ontario, the Huronian Supergroup, or at least the major group within it - the Cobalt Group - must have extended about 1,000 miles northeasterly from Lake Superior. This suggests that the total volume eroded from the Superior Province, therefore, could have been one million cubic miles or more.

In detail, patterns of Huronian sedimentation are not as simple as might appear from the generalization given above. Directions of sedimentary transport were not uniformly from north to south, thicknesses do not increase uniformly to the south, and the area of sedimentation did not sink or warp downward in a uniform manner. Cross-bedding in the Mississagi Formation, for example, indicates an easterly transport direction in the Bruce Mines area (McDowell, 1957), and a southwesterly direction near Sudbury. According to Card (1969), some of the Bar River sands south of Sudbury were carried northwestward from an area within the present Grenville Province. Card has also found that paleocurrents were deflected around the present area of the Sudbury Irruptive, indicating that it was relatively positive during part of Huronian sedimentation. Other localities within the deposition area such as the Thessalon and Chiblow anticlines also appear to have been relatively positive. Some thicknesses and facies changes are very rapid and local. The Huronian Supergroup was apparently deposited on a somewhat unstable, segmented platform.

It has been postulated by Roscoe (1969, pp. 77, 78) that recurrent epochs of glaciation influenced the pattern of Huronian sedimentation, and Casshyap (1969) suggested that they were actually the dominant influence. Periodic incursions of the sea into areas of continental sedimentation accompanied and followed the deposition of tillitic beds of the Ramsay Lake, Bruce, and Gowganda Formations. In this glacial-control model, these marine advances may be attributed to glacial loading and depression of the continent followed by rapid rise of sea-level due to melting of the ice-caps. Isostatic uplift due to unloading then resulted in a return to the dominant pattern of non-marine fluvial sedimentation. This cycle occurred three times - the last time on a truly grand scale (Fig. 1). Such uplifts and sea-level changes can amount to many hundreds of feet and result in migrations of coastlines for many hundreds of miles across low-lying parts of continents. Some of the coarsest Huronian clastics may have been deposited hundreds of miles away from the coast. The ore-bearing conglomerate beds, in particular, may have been deposited on the piedmont rather than on an alluvial plain. They are certainly not deltaic deposits as has been suggested.

Figure 2 shows the distribution of Huronian strata and suggested minimal limits for the original extent of these sediments. The present distribution of the strata of the Elliot Lake, Hough Lake and Quirke Lake Groups (see ruled area) indicates that their depositional basin extended at least from Sault Ste. Marie to the vicinity of Turner Township some 50 miles north of Sudbury. Based on this, an interpretation of the total area of distribution of the pre-Cobalt groups is shown in the figure. Prior to Cobalt deposition the basins became larger, and the figure shows two alternative interpretations depending on



whether or not the rocks in the Mistassini district, previously mentioned, are considered to be equivalent to the Cobalt Group. The broader interpretation suggests that an early Aphebian depression formed along the southeastern margin of the Superior Province. Tectonic control of this original depression may have been reflected later during the Hudsonian Orogeny, when formation of the Grenville Front was initiated.

Although some initial features had a northeasterly trend within the Huronian belt, it is by no means certain that the depositional areas as a whole had this trend, or that it was even trough-like. It may have been equidimensional, or perhaps elongated in a southerly or even a southeasterly direction. It is clear however that the depression was much larger than the present area of distribution and was a crustal feature of major proportions. Apart from the presence of till-like units and other extraordinary features that have been discussed, the Supergroup resembles deposits of some younger basins such as the Carboniferous basin of the Maritime Provinces. It may have been one of a number of similar, dominantly non-marine successions deposited at the beginning of Aphebian time in exogeosynclines near the perimeter of an Archean protocontinent. Rocks near Padlei, N.W.T., which include pyritic conglomerates and till-like units, may be remnants of another similar succession. There is evidence that Huronian rocks near Sudbury were metamorphosed and intruded by granite about 2.1×10^9 years ago. This intra-Aphebian event may have been widespread and it is possible that many areas of early Aphebian rocks were uplifted and removed prior to deposition of presently much more extensive younger Aphebian strata which are characterized by dolomites, iron-formations, other marine sediments and in some cases red beds.

In conclusion, the Huronian Supergroup is a sedimentary-volcanic assemblage distinct from other Aphebian successions, and was perhaps deposited under unique conditions. Also noteworthy is its contrast with the type of sediments found in Archean belts. This first development of platform-type sedimentation in the early Aphebian is ascribed to post-Archean changes in crustal behaviour brought about by crustal thickening and by increased crustal stability following the Kenoran Orogeny.

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Comment by: K.D. Card,
Ontario Dept. Mines

Re: Basins and Geosynclines of the Canadian Shield

The writers have concluded (see especially Fig. 2) that most of the Huronian sediments were deposited in a non-marine environment by fluvial and alluvial agencies. In my opinion, much if not most, of this material was deposited in a shallow-water marine environment.

Evidence for this includes:

(1) Possible marine glacial origin of the Gowganda, Ramsay Lake, and Bruce Formations.

(2) Facies changes in many of the formations from north to south indicating a possible non-marine - marine transition. For example:

Lower Gowganda in the north is mainly massive tilloid, while in the south it consists of bedded tilloid, sorted conglomerate, and turbidite-type sandstone.

Gowganda argillite in the north is mainly of the "varved" type whereas in the south it is irregularly laminated. Lindsey (1969) interprets this as a non-marine - marine transition - appearance of limestone and calcareous siltstone units in the Serpent in the south.

(3) Thickening of the entire section, and especially of the pelitic units from north to south. Much of this thickening appears to occur across the Murray Fault system and it is possible that this zone marks an important hinge-line within the basin.

(4) Character of formations such as the Lorrain and Bar River, which are very extensive sheets composed of well-rounded, mature quartz sands.

(5) The paleocurrent data, although far from complete, can be characterized as unimodal in the north and polymodal in the south, indicating non-marine - marine transition.

In summary, the available evidence indicates that the northern part of the sequence (Elliot Lake and Timagami basins) was deposited in shallow, epi-continental seas covering metastable Superior cratonic blocks. The southern part (Penokean fold belt) was apparently deposited in deeper water, possibly in a marginal basin produced by major rift faulting. Locally, as around Sudbury, still deeper-water conditions existed and turbidite-type sequences were deposited.

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Response by: M.J. Frarey and S.M. Roscoe

Dr. Card's suggestion that we underrated the importance of marine sediments to non-marine sediments in the Huronian succession parallels comments by Prof. F.J. Pettijohn. We have discussed this point at considerable length in our comments on Pettijohn's paper.

Comment by: J.A. Robertson,
Ontario Dept. Mines

At this stage of investigation, it is pertinent to consider the course of future Huronian studies. In this connection I have two suggestions: If available drill core could be properly stored, study of special Huronian features such as the regolith would be facilitated. The storage of appropriate drill core data in coded form for computer use is also desirable. Regional studies by formation are now possible, as government surveys have the stratigraphic and structural framework established. We would like to see the universities more in with detailed studies.

Response by: M.J. Frarey and S.M. Roscoe

We endorse and support the ideas expressed by Mr. Robertson regarding future studies in the Huronian Supergroup. Efforts toward the collection and storage of Huronian drill core and drill core information should receive all possible encouragement.

Comments by: E.H. Chown,
Loyola College, Montreal, P.Q.

I agree with the authors that the Otish-Mistassini clastics are more similar in type to the Lorrain than any other Huronian formation, and the age appears to be in the same range. However, I hesitate to accept their suggestion that complete continuity between the two areas is implied, as the Mistassini carbonate succession lies squarely between the two. I would also like to clarify the possibly confusing paleocurrent direction indicated on their map. The two southeast trending arrows indicate the trend in the arkose and conglomerate units, whereas the northeast arrow represents the trend in an intercalated orthoquartzite-subarkose unit. (I have subsequently confirmed this in conversation with Dr. Roscoe.)

Response by: S.M. Roscoe and M.J. Frarey

The Mistassini carbonate succession is younger than the Chibougamau Group which resembles the Gowganda Formation. We considered that it is also younger than the Otish clastics which resemble the Lorrain Formation. The direction of transport to the thick coarse Otish grits, conglomerates and most of the quartzites was to the southwest towards the northeast part of Mistassini Group as little as 20 miles away. It is difficult to imagine that the grits could grade and interfinger not into finer grained quartzite, siltstone or shale but into a carbonate succession in such short distances. We understand that Chown has some interpretations for relationships between the Chibougamau, Papaskwasati, Otish, and Mistassini rocks that differ from those suggested by Roscoe (1967, 1969) after very limited reconnaissance work, but we cannot comment on these until they are published and we have the opportunity to study them in detail.

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THE HURWITZ GROUP - A PROTOTYPE FOR DEPOSITION ON METASTABLE CRATONS

R.T. Bell
Department of Geological Sciences, Brock University
St. Catharines, Ontario

Abstract

Aphebian rocks of the District of Keewatin are described with reference to paleocurrents, depositional phases, and an interpretive sedimentation model.

The first three phases are represented by complex continental deposition in small depressions, with transition and continued transgression into autogeosynclinal and platform conditions. A possible glacial suite occurs in the second phase.

The next and most extensive succession, comprising supermature orthoquartzites with remarkable persistence of sedimentary detail and thickness, suggests a broad stable platform. Subsidence in this phase was probably slow and subequal over the craton, keeping pace with shallow water conditions.

The fifth phase is typified by rapid transition to pelitic sediments, including greywacke, carbonates and volcanics, illustrating deepening and transformation from a stable platform to a geosyncline, and to deeper water conditions.

The final phase is represented by break-up of the earlier pattern compounded by extensive cannibalization of the earlier units. Transition to this phase appears to have been continuous in the major depression. The last phase may be a harbinger of the Hudsonian orogenic events - an "exogeosynclinal" condition or early stage "molasse".

This synthesis and model for the Hurwitz succession, interpreted through a structural and metamorphic veil, is proposed for sedimentation on metastable cratonic Archean platform, for comparison with those directly on or marginal to stable cratons.

DISCOVERY AND MAPPING

Robert Bell (1885, 1887), after visiting the west coast of Hudson Bay, described and named the quartzites near Rankin Inlet after Marble Island. He reported on hearing of similar quartzites further inland, presumably near Whiterock Lake. J.W. Tyrrell (1897) described quartzites in the Kaminak and Whiterock belts and extended Bell's terminology "Marble Island Quartzites" to include these. No further reference to this name has been made, save some brief comments by the author (1968) on the history of the Hurwitz Group. The writer suggests dropping Robert Bell's terminology because of lack of use, because the name is misleading (Marble Island is made up of orthoquartzite), and because it has been used elsewhere.

The reconnaissance surveys of Lord (1953) and Wright (1955, 1967) outlined Aphebian belts of sediments typified by orthoquartzite (Fig. 1). To these, Wright applied the name "Hurwitz Group" throughout the southern and central district of Keewatin. Eade (1964, 1966) mapped the Kognak River area immediately north of Hurwitz Lake and subdivided the Hurwitz Group into mappable units, separated out a few scattered patches of Archean greywacke, and identified a small basin of sedimentary rocks clearly lying between Hurwitz and Archean rocks near Montgomery Lake. Subsequent work has added a few subdivisions to Eade's units.

The author concentrated his studies for the Geological Survey of Canada, east and northeast of Kognak River area during the summers of 1967 to 1969. He benefited greatly from Eade's constructive assistance and criticism, as well as from that of A. Davidson, W.W. Heywood, G.M. Wright and S.M. Roscoe.

AGE RELATIONS AND LITHOSTRATIGRAPHIC UNITS

Robert Bell and J.W. Tyrrell both suggested that rocks of the Hurwitz Group were Huronian. More generally (if we restrict Huronian to the Huronian Supergroup) the unit would be Early Proterozoic or Aphebian (Lord, 1953; Wright, 1967; Stockwell, 1964). The author considers that the Hurwitz is not Huronian, but more likely Late Aphebian.

Wright (1955, 1967) included all Aphebian sedimentary rocks in the Hurwitz Group. Because the Amer, Schultz and other basins north and west of Baker Lake are greatly separated from the semicontinuous basins of the Rankin-Ennadai greenstone belt, the author restricts the usage of Hurwitz Group areally to the Rankin-Ennadai greenstone belt (Fig. 1) and refers informally to the others as "Amer Lake Sediments". Significant stratigraphic detail distinguishes Hurwitz and Amer sediments (Bell, 1969) and it is probable that these groups are not traceable into each other, although radiometric dating may indicate some measure of correlation.

Minor refinements to Eade's stratigraphy have been presented earlier (Bell, 1968, and in press) and they, with the description of these units are summarized in Table 1. Paleocurrent data for the northeastern half of the Rankin-Ennadai greenstone belt have been compiled systematically during the present study. Brief visits were made to the Kognak and Watterson basins.

GENERAL INTRODUCTION

The Hurwitz and Montgomery Lake strata overlie Archean volcanic and sedimentary rocks and late Archean plutonic complexes in a broad belt between Ennadai Lake and Rankin Inlet. The Archean basement has been referred to as the "Hudson Protocontinent" by Goodwin (1968). The author proposes to call this the "Hudson Metastable Craton" with respect to Aphebian sedimentation. In contrast to Huronian and Labrador trough rocks which lie on and marginal to fully stable Archean cratonic elements, rocks of the Hurwitz Group lie entirely within a structural province that has undergone extensive Hudsonian remobilization. Figure 2a summarizes the distribution and intensity of the Hudsonian orogeny on the Hurwitz.

PHASE I - First Terrestrial Basin, Montgomery Lake Sediments

The "Montgomery Lake Sediments" lie directly on Archean basement rocks and comprise drab, immature to mature arenites, quartz-pebble orthoconglomerates and siltstones. A basal paraconglomerate was observed at several localities. The Montgomery Lake sediments are crossbedded and ripple marked; in the Padlei belt, a few scattered observations suggest a westerly component to sediment transport. A pyritic, slightly radioactive conglomeratic zone on the south flank of the Padlei synclinorium shows a decrease in pebble size to the southwest and a concommitant decrease in uranium content and in U/Th ratio. The environment was probably continental and fluvial.

An almost total lack of hematitic beds, an abundance of pyritic arenites and rudites in what is considered to be a continental environment suggest that the environment was anoxic (Roscoe, 1969) or not oxidizing. Deposition was probably restricted to a small area in the Kognak and Padlei areas (Fig. 2a). An unconformity separates the Hurwitz and Montgomery strata, but

TABLE I LITHOSTRATIGRAPHIC NOMENCLATURE OF HURWITZ GROUP

Bell (in press)	Bell (1968)	Eade (1964)	Description
---	Hurwitz G	unit 12	Pale brown, feldspathic, lithic, dolomitic, quartz arenites; minor varicoloured siltstone & shale; 1000-2000' thick.
---	---	10 & 11	Post Ameto complex: grey & green greywacke, siltstone, argillite, & dolomite; at least 3 dolomite units in Watterson basin, but none in McConnell belt; 2000-5000' thick.
Ameto Fm. Happotiyik Member	" Fa	---	Massive to pillowed volcanics; minor tuff & chert; 1000' (plus) at Kaminak Lake, thins to 80' in Padlei belt, apparently absent elsewhere.
Ameto Fm.	" E	9	Grey, black, & green slate, siltstone and fine greywacke; minor red slate and dolomite; about 2000' thick
Kinga Fm. Whiterock L. Member	" D	8	Thin-bedded, white, grey, & pink fine-grained orthoquartzite; ripple marked; 850' thick from Hudson Bay to Padlei, 1100' in Kognak basin, thins to 600' in Watterson basin.
Kinga Fm. Maguse Member	" C	8	Thick-bedded, white, grey & pink, medium to coarse-grained orthoquartzite to feldspathic quartz arenite; grit and orthoquartzite locally abundant; 1000' Padlei, 2000' at Ameto L, as much as 4000' Kognak basin.
Padlei Fm.	" A & B	8a	Grey & green slate, siltstone, polymict boulder to pebble conglomerate and conglomeratic mudstone to sandstone; irregular, may be as much as 1500' thick.
"Montgomery Lake Sediments"	---	7	Green, grey, feldspathic to chloritic quartz arenites, siltstones & pyritic quartz-pebble conglomerate, with basal polymict boulder conglomerate; 3000' thick in Padlei belt.
Archean Complex	---	---	Kaminak Group (volcanic & sedimentary rocks) cut by granitic & dioritic plutons.

RTB - 1970

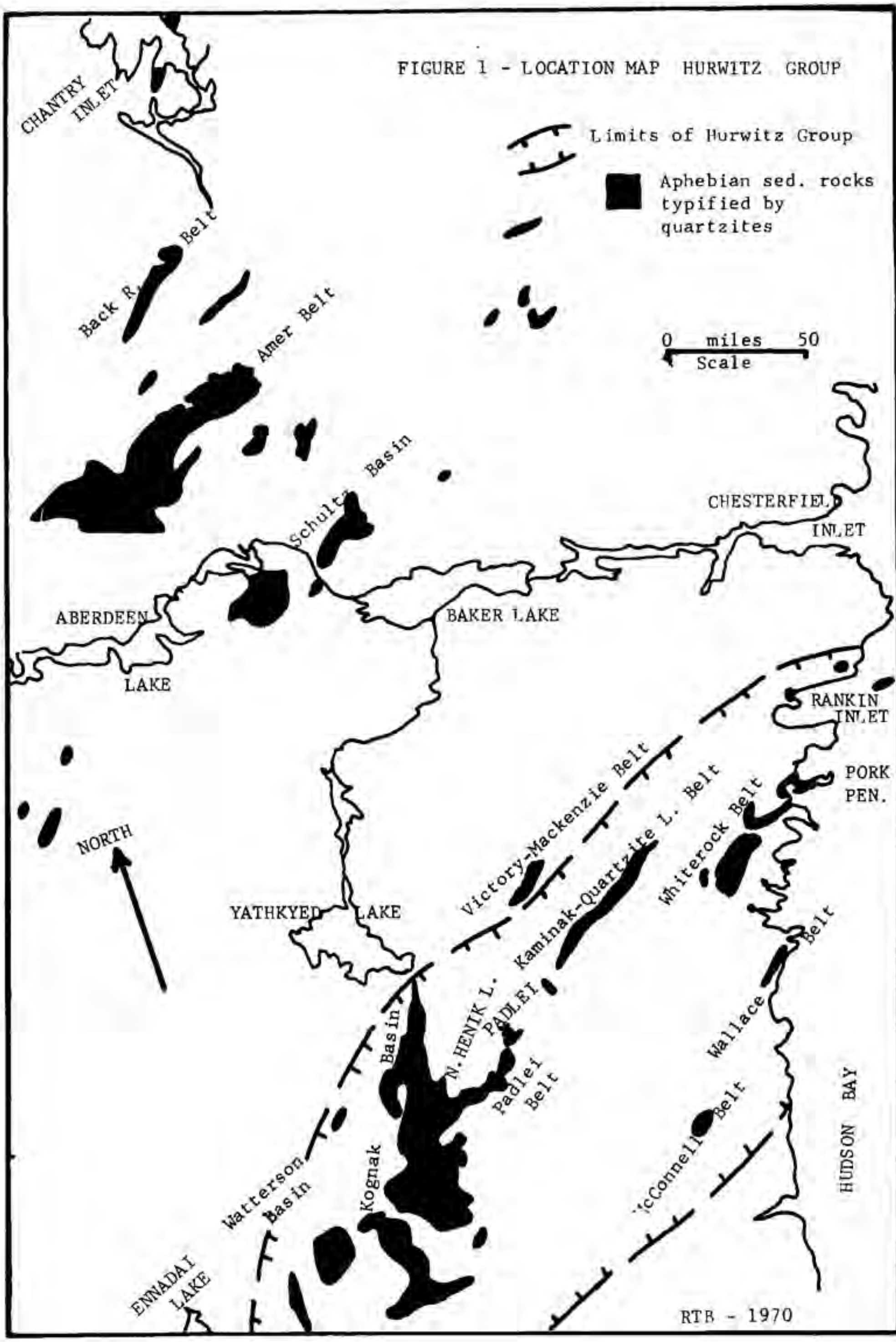
FIGURE 1 - LOCATION MAP HURWITZ GROUP
Limits of Hurwitz Group
Aphebian sed. rocks typified by quartzites
0 miles 50
Scale
CHANTRY INLET
Back R. Belt
Amer Belt
Schultz Basin
CHESTERFIELD INLET
ABERDEEN LAKE
BAKER LAKE
RANKIN INLET
PORK PEN.
NORTH
Victory-Mackenzie Belt
Kaminak-Quartzite L. Belt
Whiterock Belt
YATHKYED LAKE
N. HENIK L.
PADLEI
Basin
Padlei Belt
Wallace Belt
McConnell Belt
HUDSON BAY
Watterson Basin
Kognak
ENNADAI LAKE
RTB - 1970

the degree and type of deformation is yet unclear. In the Padlei belt, 3,000 feet of Montgomery strata are cut out in a northwesterly direction on a distance of three miles.

PHASE II - Second Terrestrial Basin - Padlei Formation

The Padlei Formation is a heterogeneous assemblage of paraconglomerates grading upwards into thinly bedded varved mudstones and siltstones. The resemblance between the Padlei and Gowganda Formations tempts one to ascribe a mudflow and/or glacial origin to the Padlei, although none of the critical glacial indicators have yet been found in the Padlei belt.

The basin of accumulation is similar to, and slightly smaller than that of the succeeding Maguse Member (Fig. 2b). The Padlei Formation generally occurs as a continuous sheet 500 to possibly as much as 1,500 feet thick, but is missing locally. Phenoclasts reflect the presence of an Archean basement; no pebbles clearly derived from the Montgomery Lake sediments have been identified.

PHASE III - Third Terrestrial Basin - Maguse Member of Kinga Formation

The backbone of the Hurwitz Group is the Kinga Formation which is readily divisible into two Members: Maguse and Whiterock Lake Members.

The Maguse Member is typified by thick bedded, coarse- to medium-grained pink and white orthoquartzite with somewhat arkosic strata near the base. It closely resembles the Lorrain Formation, even to the extent of carrying jasper-bearing orthoconglomerates. Crossbedding and ripple marks are locally present. Thickness reaches 4,000 feet in the Kognak Basin and decreases rapidly to the west and somewhat less rapidly to the east (Figs. 2b and 4).

Current directions in the northeastern part of the basin have a strong northeasterly component but more documentation is necessary to establish if this pattern is prevalent in the member.

The presence of locally abundant metamorphic andalusite in quartzites near Padlei and Kaminak Lake strongly suggests that the original non-quartzose components of the sediments were gibbsite and/or boehmite, probable products of intense tropical weathering. This and the maturity contrast with arenites of the Montgomery Lake sediments probably reflect a much subdued and distant source area. Pebbles in the orthoconglomerates may have been derived from Montgomery Lake sediments.

PHASE IV - Platform Sediments - Whiterock Lake Member of Kinga Formation

The Whiterock Lake Member is typified by well-bedded, supermature, fine-grained, pink to white orthoquartzites and is the most extensive unit (in exposure and area) of the Hurwitz Group. The unit is spectacularly ripple marked. Crossbedding is present in and east of the Kaminak-Quartzite Lake belt. Thickness is remarkably consistent. East and northeast of Padlei, the Whiterock Lake Member is about 850 feet thick; from Padlei west it thickens to about 1,100 feet; and then thins again west of the Kognak basin to about 600 or 700 feet at Watterson Lake.

Both Eade (1964) and the author have made extensive measurements on ripple marks in the Kinga Formation. Figure 3a summarizes more than seven hundred measurements made by the author. The sedimentary strike is northeasterly, spectacularly uniform, and the bulk of the current ripples indicates transport to the northwest (certainly as far west as Henik Lake). Eade's observations west of Henik Lake (pers. com. 1969) suggest a more complex pattern for both members of the Kinga Formation. The author feels that separation into two members is possible and necessary.

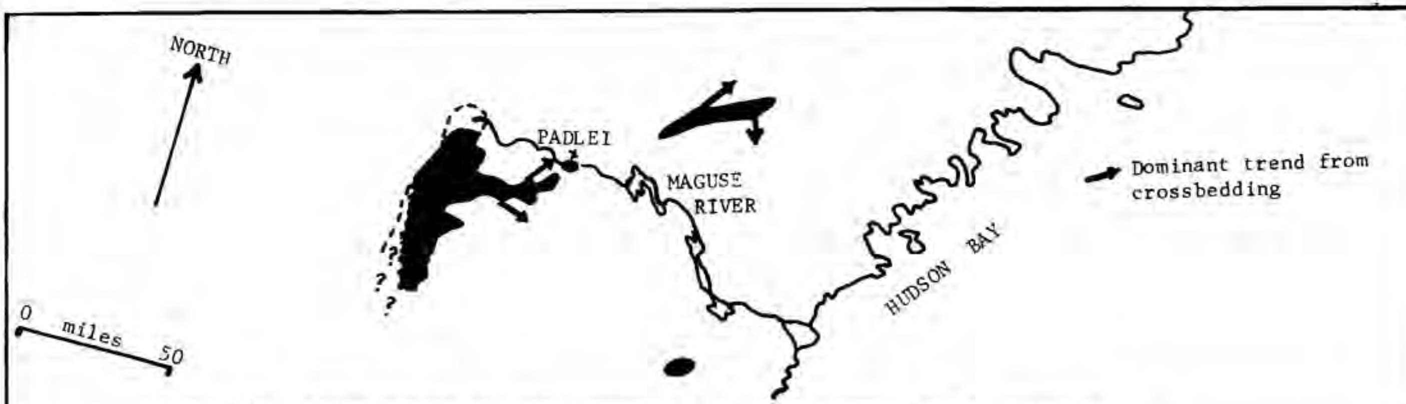

FIGURE 2b DISTRIBUTION AND PALEOCURRENTS - MAGUSE MEMBER, KINGA FORMATION

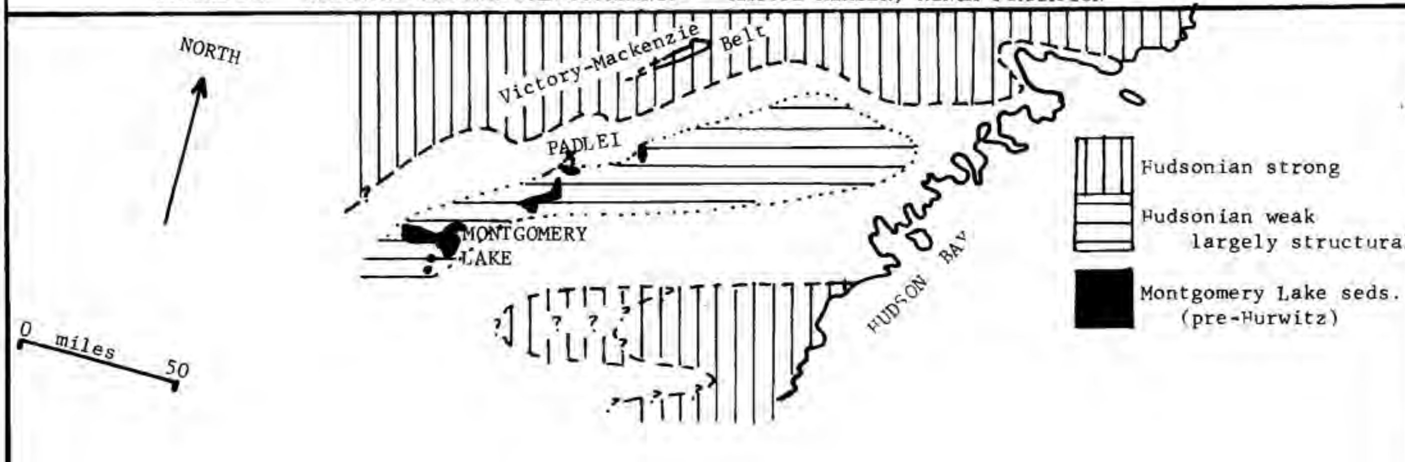

FIGURE 2a DISTRIBUTION OF MONTGOMERY LAKE SEDIMENTARY ROCKS, AND EFFECTS OF HUDSONIAN OROGENY

RTB - 1970

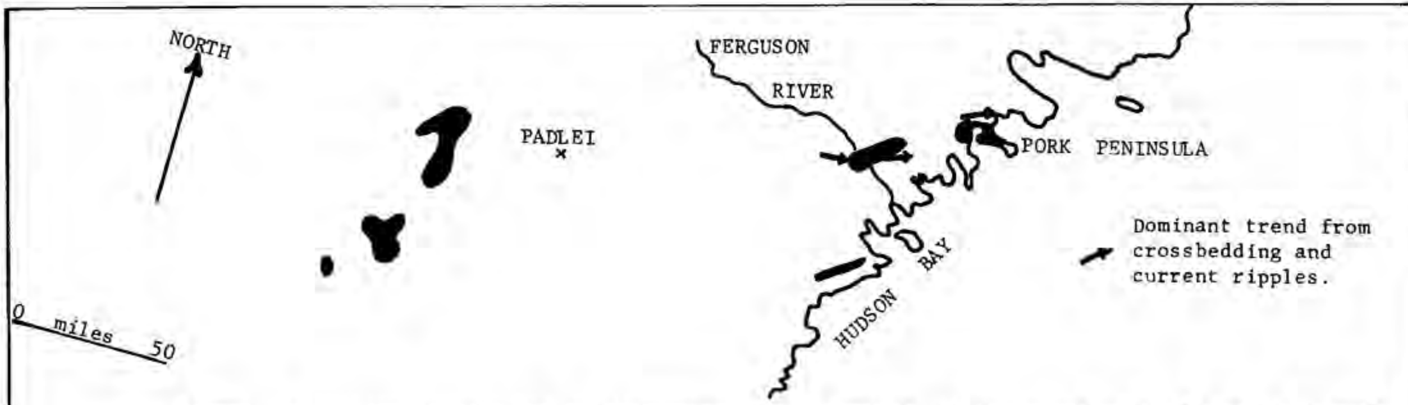

FIGURE 3b DISTRIBUTION AND PALEOCURRENTS - "HURWITZ G" (youngest unit of Hurwitz Group)

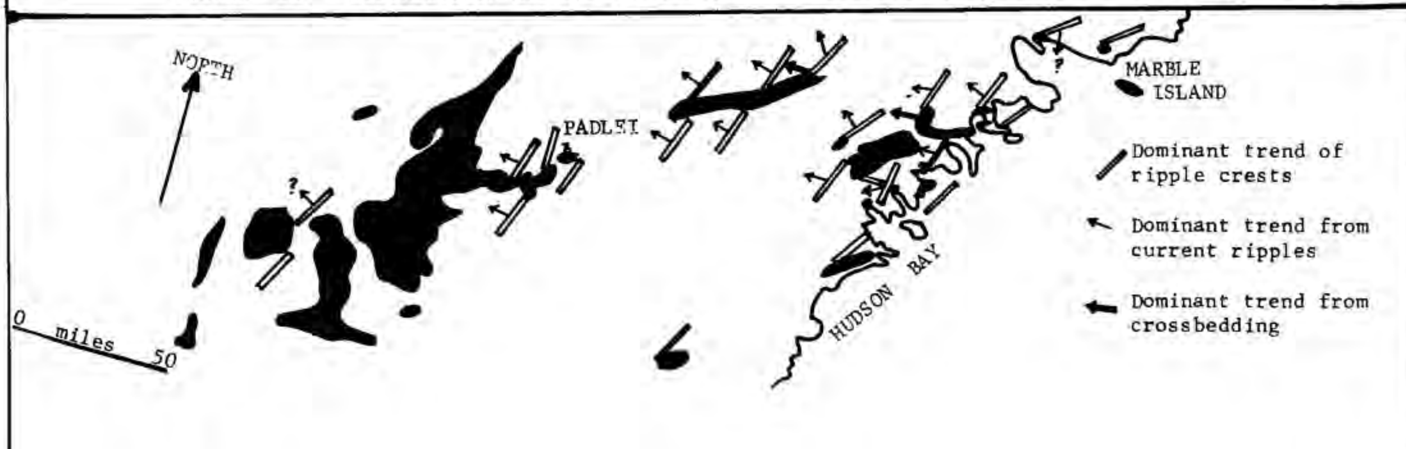

FIGURE 3a DISTRIBUTION AND PALEOCURRENTS - WHITEROCK LAKE MEMBER, KINGA FORMATION

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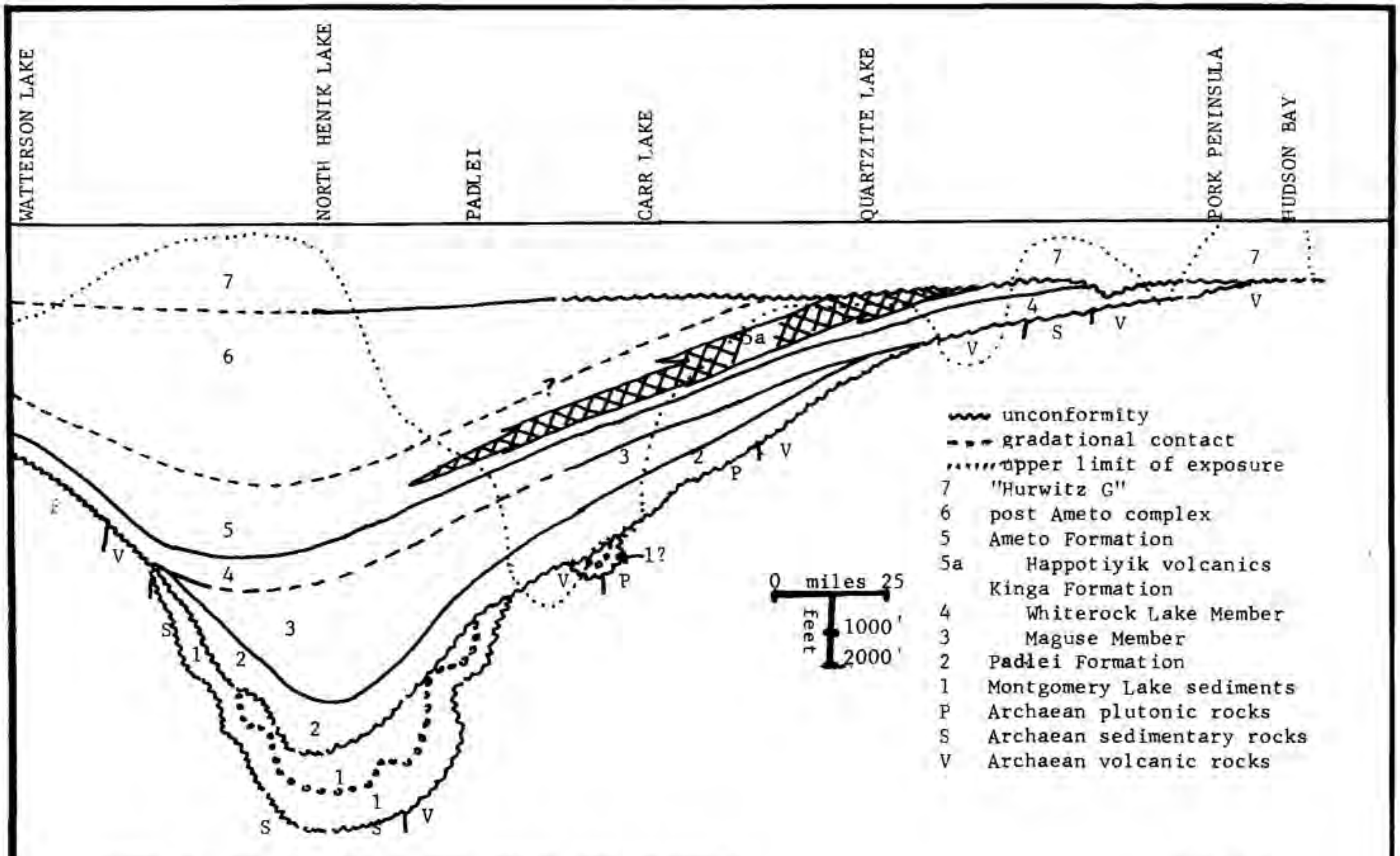

FIGURE 4 SKETCH OF HURWITZ BASIN AND FLATFORM

RTB-1970

The fine-grained nature and sedimentary features of the Whiterock Lake orthoquartzite suggest that a broad blanket of sand was laid down in shallow water on a stable cratonic platform associated with slow, steady transgression across the Canadian Shield. A few thick planar crossbedded units near and east of Whiterock Lake may represent beach deposits. Mud cracks are rare but are nevertheless present near the top of the unit which is marked by intensely red orthoquartzites. In several widely separated localities where the complete transition to the overlying Ameto Formation is observable, breccia, probably of primary origin, is present. A period of emergence and weathering might have terminated the Whiterock phase.

PHASE V - Geosynclinal Stage - Ameto Formation and Post-Ameto Units

The Ameto Formation is extremely poorly exposed. Eade and the author have been able to piece together an assemblage formed dominantly of grey, greenish grey and black slates with some varicoloured slates, small stromatolitic carbonate reefs, volcanics and lean iron formation.

The volcanic unit (Happotiyik Member) is about 1,000 feet thick in the Kaminak-Quartzite Lake belt and thins to four flows with an aggregate thickness of 80 feet near Ameto Lake. The enclosing fine-grained greywackes are tuffaceous. Associated with the flows are thick gabbroic sills. Similar sills are present at approximately the same level throughout the area from Wallace River to Watterson Lake. The sills are believed to be coeval with the Ameto Formation and probably with the uppermost part of the Happotiyik Member. Quite possibly, some of the greywackes in the Ameto Formation and in succeeding units are derived in part from concurrent vulcanism.

The post-Ameto complex in the Kognak basin was divided into two units by Eade (1964, 1966): a lower (dominantly dolomite and siltstone) member, and an upper (dominantly greywacke) unit. In the Watterson basin, at least three carbonate units are interbedded with greenish slates and fine-grained greywackes. In contrast, no carbonate units were observed in the McConnell basin, where the Ameto and post-Ameto complex above the gabbro sill grade upwards to increasingly coarser greenish metagreywacke. Probably because of poor exposure, Eade (1964, 1966) was locally unable to distinguish the two units that are then represented by one or the other unit or by a combined unit. These units might merely have local significance (5 to 20 miles extent) and the Ameto and post-Ameto complex might be a laterally and vertically varying mixture of marine mudstones, siltstones, carbonate banks, greywackes, and volcanic rocks with associated sediments, representing an abrupt deepening, broadening and growing instability of the Hudson metastable craton. It is impossible to determine whether more than one broad trough or basin were present. Certainly the region west of Henik Lakes has the thickest exposed section of the Ameto Formation; accordingly the main axis of the basin may have been west of Henik Lakes and even west of Watterson Lake.

PHASE VI - "Exogeosynclinal" or "Molasse" Stage - "Hurwitz G"

The "Hurwitz G" is defined (Bell, 1968) in the Whiterock-Pork Peninsula region near Hudson Bay coast (Fig. 3b). It comprises immature, micaceous, feldspathic, locally dolomitic, quartzose arenites, with minor subarkose, orthoquartzite, orthoconglomerate, and red and drab siltstones.

The base of Hurwitz G in the Whiterock belt is an orthoconglomerate with quartzite pebbles lithologically similar to the Kinga Formation. The base of the unit is about 1,000 feet above the Kinga Formation at the south end of the Whiterock belt and gradually cuts down section to basement rocks in the Pork Peninsula. The thickest expression of Hurwitz G is in the Pork Peninsula (3,000 to 4,000 feet) where the upper half of the formation contains abundant chips and a few cobbles of greenstone as well as orthoquartzite cobbles, and where it shows an inverse reflection of progressive erosion in the source area through the Hurwitz to basement rocks. Progressive coarsening upwards may reflect increasing diastrophism in the source area. Sediment transport appears to be from the southwest (Fig. 3b) and is not in accord with the regional dip of the unconformity. Uplift or graben structures could possibly account for this.

On lithological grounds, the author has proposed (Bell, 1968), that the highest member in the Kognak area is correlative with Hurwitz G (this proposed correlation is likely a weak link). In this area, the unit is gradational downwards to greywackes of the post-Ameto complex (Eade, pers. com., 1968). If the correlation is correct, we have a gradual transition from marine to nonmarine conditions in the major downwarped portion of the Hurwitz basin, concommitant with uplift and progressive cannibalization certainly at least in the eastern region and possibly elsewhere. A parallelism with the development of Phanerozoic exogeosynclines can be attempted, suggesting that Hurwitz G is a harbinger of the Hudsonian orogeny.

AGE CONSIDERATIONS

The Hurwitz and Montgomery Lake sediments lie with profound unconformity upon Archean volcanic and sedimentary rocks (Kaminak Group - A. Davidson, 1970) and upon late Archean plutonic complexes, dominantly diorites and quartz monzonites older than about 2,500 m.y. (K-A method on hornblende, Pb_{206}-Pb_{207} on sphene, A. Davidson and R.K. Wanless, pers. com., 1969). The Archean complex is cut by porphyritic diabase dykes which have been dated at about 2,300 m.y. (whole rock, K-A method, A. Davidson, pers. com., 1969). These dykes are clearly older than the Hurwitz Group. However, their relation to the Montgomery Lake sediments is not clear. The Hurwitz Group and basement rocks have been involved to varying degrees in the Hudsonian orogeny (about 1,700 m.y. ago). Accordingly, the limits on age of the Hurwitz strata are 2,300 and 1,700 m.y. For comparison, the limits on Huronian strata are 2,500 and 2,100 m.y.

Red beds, or hematite-bearing arenites and lutites are present from the Maguse Member upwards - essentially from early Hurwitz time on, but none were observed in the Montgomery Lake sediments. Red beds are also almost totally lacking from the pre-Cobalt section of the Huronian. Roscoe (1969) has discussed this "anoxic" character of Huronian rocks. Considerations regarding evolution of the atmosphere strongly suggest that a transition from nonoxidizing (anoxic/reducing?) to oxidizing conditions occurred before Helikian times. It appears that this transition occurred in pre-Lorrain, and in this case pre-Maguse time. Hence, if the above speculation is sound, the Hurwitz is largely, if not entirely, younger than the Huronian.

The picture concerning the deposition model shows a transition in late Hurwitz time to early diastrophism, associated by definition with an orogenic episode, presumably the Hudsonian event that terminated the Aphebian era.

SUMMARY

Deposition of Hurwitz and Montgomery Lake sediments on the Hudson Protocontinent or metastable craton is represented by the following stages:

(a) Terrestrial basin stage where deposition of the three lowest units is reflected in the Padlei-Kognak basin and reflects growing stability and geomorphic maturity of the craton;

(b) Stable Platform stage with deposition of the Whiterock Lake Member "blanket-sand" and transgression onto the craton;

(c) Geosynclinal stage with deepening and broadening of the stable platform, introduction of volcanic material, possibly indicating that the crustal underpinnings were removed at the end of Whiterock time; and

(d) "Exogeosynclinal" stage with developing diastrophism in the now unstable craton, cannibalization of parts of the older Hurwitz Group, and transition to the Hudsonian orogeny.

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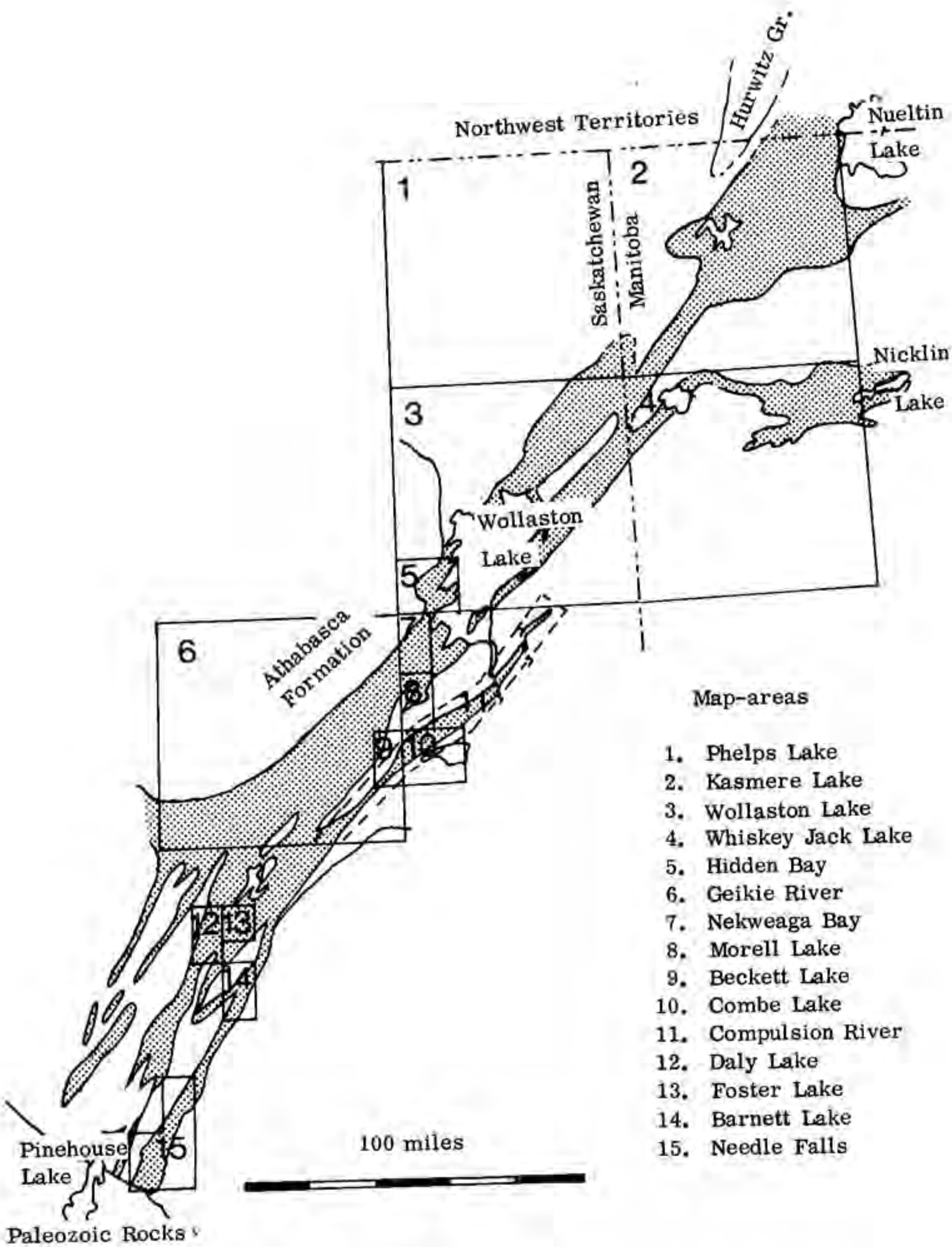

Figure 1. Known extent of the Wollaston Lake fold belt (dotted).

THE WOLLASTON LAKE BELT, SASKATCHEWAN, MANITOBA, NORTHWEST TERRITORIES

P.L. Money, Texas Gulf Sulphur Company, Toronto.
A.J. Baer, Geological Survey of Canada, Ottawa.
B.P. Scott and R.H. Wallis, Saskatchewan Department of Mineral Resources, Regina.

Abstract

The Wollaston Lake Belt consists mainly of supracrustal rocks. It extends in a north-northeasterly direction from the southern edge of the Canadian Shield near Pinehouse Lake into Manitoba and the Northwest Territories. It is at least 400 miles long, reaches a maximum width of about 50 miles, and is generally bounded by plutonic rocks. Available evidence suggests the presence everywhere (?) beneath the belt of an Archean sialic basement.

The Aphebian supracrustal rocks of the belt are divisible into at least four assemblages. The predominant one consists mainly of "meta-arkose" and biotite gneiss and has been assigned locally to the Daly Lake Group. The Hidden Bay assemblage, characterized by marble and calc-silicate rocks, appears to be correlative with the Daly Lake Group although atypical of it. A third assemblage differs from the first by the presence of hornblende-biotite gneiss, amphibolite, and "acidic meta-volcanic (?) rocks". It is probably confined to the eastern margin of the belt and it has been assigned locally to the Sandfly Lake Group. A fourth assemblage, assigned locally to the Meyers Lake Group, consists of quartz-pebble meta-conglomerate, quartzite, and pelitic schist.

The thicknesses of the Daly Lake, Sandfly Lake, and Meyers Lake Groups are at least 5,000 feet, 1,500 feet, and 1,500 feet respectively. More realistic estimates for the Daly Lake and Sandfly Lake Groups are probably 10,000 feet or more and 5,000 feet respectively.

The Sandfly Lake Group, probably the oldest group, includes rocks probably derived from immature sediments such as arkosic wacke and greywacke and also rocks of probable volcanic origin. It appears to have been deposited unconformably on the Archean basement under conditions of maximal crustal instability for the area. It is overlain, probably unconformably, by the Meyers Lake Group, which consists of relatively mature rocks and is inferred to have been deposited during a period of comparative crustal stability. The Daly Lake Group was probably deposited conformably on the Meyers Lake Group to the east and unconformably on the Archean basement to the west. The rocks comprising it were probably predominantly derived from arkosic arenites, greywacke, and shales. They are intermediate in maturity between those forming the other two groups. Limited evidence indicates shallow water marine deposition for at least some of the rocks assigned to each group.

Correlation with rocks of the Hurwitz Group and other rocks outside the belt is uncertain.

INTRODUCTION

The Wollaston Lake Belt is a zone of predominantly supracrustal rocks extending in a northeasterly direction from the southern edge of the Canadian Shield near Pinehouse Lake, Saskatchewan at least as far as Nueltin Lake on the Manitoba-Northwest Territories boundary, and possibly farther (Fig. 1). It has a minimum length of about 400 miles and varies in width from a minimum of seven miles in northwestern Manitoba to an apparent maximum of about 45 miles near Highrock Lake at latitude 57°N. Because of inconsistencies between map-units in adjacent map-areas, the exact width of the northern part of the belt is still unknown.

A number of subordinate belts extend out from the main one. They include the Compulsion River Belt (Møller, 1969) south of Wollaston Lake, the Misty Lake-Nicklin Lake Belt (Money, 1968) and an ill-defined area west and southwest of Nueltin Lake in northern Manitoba. For the purposes of this paper, these belts and a few remnants of supracrustal rocks between Highrock Lake and the southern edge of the Shield are all considered to be part of the Wollaston Lake Belt.

The Wollaston Lake Belt is contained entirely in the Churchill structural province of the Canadian Shield. Supracrustal rocks within it consist mainly of metamorphosed shales, arkosic arenites and greywackes. Variable proportions of acidic to basic volcanic rocks occur along the eastern margin of the belt. All of these rocks have been tightly folded into steeply dipping isoclinal or nearly isoclinal folds, and most are now in the amphibolite facies of metamorphism (Abukuma type or low pressure series). K-Ar ages on metamorphic micas suggest that the latest folding and metamorphism of the belt occurred during a Hudsonian event (1,570-1,820 m.y., *see* Table III, p.176).

To the west and southwest of Wollaston Lake, unmetamorphosed and little deformed Athabasca sandstone rests unconformably on rocks of the belt. Elsewhere, the belt is bounded mainly by granitic rocks of uncertain age. Granitic rocks also occur within the belt, generally as elongated, northeasterly trending bodies. Some of them represent Archean material in part remobilized during the Aphebian, whereas others were perhaps first emplaced during the Aphebian (Hudsonian orogeny). Distinction between the two types is difficult, but crude estimates of relative proportions of "old" and "young" components suggest that most granitic rocks are remnants of an Archean basement on which were deposited the Proterozoic supracrustal rocks.

Interest in the Wollaston Lake fold-belt goes back at least to 1952, when geologists of Eldorado Mining Corporation discovered radioactive occurrences in the Foster Lakes area. In 1953, a boulder of chalcocite-bearing "meta-arkose" was found in the same area, and a systematic search for base metals began. Discovery of Pb-Zn mineralization in other boulders led mining companies to explore the Compulsion River Belt, south of Wollaston Lake. Late in 1968, a drill hole southwest of Wollaston Lake intersected some promising uranium-mineralization in rocks of the Wollaston Lake Belt, causing renewed interest in the mineral potential of the area. Meanwhile, following the initial discovery of 1952, the Saskatchewan Department of Mineral Resources was systematically mapping the belt at a scale of one inch equals one mile (Mawdsley, 1957; Money, 1961, 1965, 1966; Chadwick, 1966, 1967; Scott, 1969, Scott, in preparation; Wallis, in preparation) and the Geological Survey of Canada was undertaking reconnaissance work in other parts of the belt (Fahrig, 1958; Tremblay, 1960; Currie, 1961; Fraser, 1962; Baer, 1969).

This paper is based mainly on the better known part of the belt, between the southern edge of the Canadian Shield and Wollaston Lake (Fig. 2). It should be considered as a progress report on the present state of knowledge of the Wollaston Lake Belt. Many geologists have contributed to this paper by their discussions or by supplying unpublished information. They include G.S.C. officers W.L. Davison, J.A. Fraser (northern Manitoba) and K.E. Eade (District of Keewatin), H. Baadsgaard and R.A. Burwash of the University of Alberta, and J.J. Brummer and S.K. Møller (Compulsion River Belt) of Falconbridge Nickel Mines Ltd. Their contributions are gratefully acknowledged.

ARCHEAN AND PROBABLY ARCHEAN ROCKS

Introduction

Probable Archean granitic rocks can be recognized in most areas where detailed geological studies have been undertaken in recent years (Table I). As our knowledge of the Wollaston Lake fold-belt increases, more extensive areas of Archean rocks will be recognized, and the "patchy" distribution presently apparent (Fig. 2) will be considerably modified.

From the southwest to the northeast, rocks of probably Archean age have been described from the Needle Falls area, the Daly Lake area, and east of Geikie River as far north as Wollaston Lake.

Rocks believed to be Archean are principally granites and quartz monzonites. In spite of their distribution in a 200 mile long belt, all known bodies are lithologically almost identical and will be described together. The Pederson Lake complex of the Daly Lake area (Money, 1966) is considered separately because it includes rocks in the granulite facies of metamorphism.

Granitic Rocks

Granitic rocks are commonly pink, more rarely grey (Needle Falls[1]) and they weather pink, with occasional rusty spots (Needle Falls, Morell Lake). They are generally medium-grained or coarse-grained, but fine-grained varieties occur in the Needle Falls and Beckett Lake areas. With the exception of the Needle Falls area, the rocks are strongly foliated and all descriptions indicate the presence of pervasive cataclastic deformation in planes parallel to the northeasterly regional trend. Flaser structure (Hidden Bay, Geikie River) and coarse-grained augen in a finer grained matrix (Morell Lake) appear to be common. Foliation appears to be less well developed in the Needle Falls area.

Major minerals are quartz, microcline, plagioclase and biotite. Hornblende is rare, and epidote, magnetite-ilmenite, sphene, apatite and zircon are accessory minerals.

Quartz is commonly in small anhedral grains and displays undulatory extinction. Potassic feldspar is almost entirely microcline, although Carlsbad twins have been reported from the Needle Falls area, and Baveno twins from the Geikie River area. String perthite, patch perthite and replacement perthite are common. Plagioclase varies in composition from calcic oligoclase to sodic andesine, except for rocks in the Needle Falls area, where sodic oligoclase (An_{20} or less) is common. In the Nekweaga Bay area, Chadwick (1966, p. 18) has reported albite rims around oligoclase grains in contact with microcline. Biotite is nowhere abundant (up to six per cent in most thin sections examined). It is commonly pleochroic in shades of brown to pale yellow, although greenish pleochroism has been reported by Baer (1969) from the Geikie River area. Green hornblende has been reported from a few localities. As shown in Figure 3 these rocks are dominantly of granitic to quartz monzonitic composition.

Pederson Lake Complex

The Pederson Lake complex has been identified by Money (1966, p. 12) in the Daly Lake map-area as a group of mappable rock units: granitic rocks, charnockitic rocks, and minor hypersthene amphibolite and plagioclase amphibolite. The granitic member forms over eighty per cent of the complex. Its composition is similar to that of previously described granitic rocks, but it differs from most of them by its deep reddish colour, and by the scarcity of microcline twinning and the abundance of exsolution perthites in potassic feldspar. Charnockitic rocks intimately associated with the granitic rocks differ from them by their buff to brown colour, both on fresh and weathered surfaces, and by the presence of up to fifteen per cent brown, weakly pleochroic, poikiloblastic hypersthene. Some hornblende replaces hypersthene, and some biotite replaces either hypersthene or hornblende. The rocks are unfoliated or weakly foliated and equigranular, except where they contain megacrysts of hypersthene. Their composition ranges from that of a charnockite to that of an enderbite. The hypersthene amphibolite and plagioclase amphibolite may represent Archean supracrustal rocks.

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[1]See Table II for references.

Structural relationships

Cataclastic structures are characteristic of most Archean rocks in the Wollaston Lake fold-belt. Contacts with supracrustal rocks have rarely been observed. All authors agree that contacts generally appear conformable, and that, in most areas, schistosity or foliation of the granitic rocks is parallel to the schistosity of surrounding supracrustal rocks. Baer however (1969) has observed that the original foliation of the granitic rocks in the Geikie River area is locally oblique to the schistosity. In the Needle Falls area (Money, 1965, 1967), the probable Archean rocks contain rotated inclusions of rocks which are lithologically similar to rocks belonging to the Daly Lake and Sandfly Lake Groups. They also may form the intrusive component of migmatites developed from rocks of the Sandfly Lake Group and rocks probably correlative with the Daly Lake Group. It is suggested that these relationships are due to remobilization during the Hudsonian orogeny. Outside of this area, evidence of intrusive relationships is limited to two occurrences, one in the Beckett Lake area, where a granitic vein cuts some meta-arkose, and one in the Hidden Bay area, where a 3-foot-wide vein transects a gneissic inclusion in the granite. In the Hidden Bay area, well exposed contacts show the presence of a pegmatitic zone between the granitic rocks and a biotite gneiss or granofels. The pegmatite is a few feet wide and cuts both other rock-types. Similar relationships exist south of Big Sandy Lake in Geikie River area, where granitic rocks and biotite gneiss are separated by apparently intrusive pegmatitic material.

Age and correlation

Arguments of structural, petrological, stratigraphical and radiometric nature have been brought forward to suggest an Archean age for the granitic rocks described above.

Where it is visible, the cataclastic foliation of the granitic rocks is parallel to the schistosity and probably to the bedding (Morell Lake area) of surrounding supracrustal rocks. Comparable cataclastic features are rare in the supracrustal rocks, suggesting that cataclasis was more or less synchronous to the folding of the supracrustal rocks, and thus that the granitic rocks were emplaced prior to the Hudsonian deformation. In the Daly Lake area, the Pederson Lake complex occupies an anticlinal position relative to rocks of the Daly Lake Group and the lack of evidence for migmatisation or intrusion along the contact suggests that the complex is older than the metasedimentary rocks resting on it. Money (1966, 1968) has discussed the metamorphism of the Pederson Lake complex and writes (1968, p. 1491): "This complex has undergone granulite facies regional metamorphism and later retrogressive amphibolite facies metamorphism, whereas the Daly Lake Group has undergone amphibolite facies metamorphism only. It is suggested that the granulite facies metamorphism must have occurred prior to deposition of the Daly Lake Group."

No angular unconformities have been observed between Archean granitic rocks and supracrustal rocks, but in the Compulsion River area (Møller, 1969) a basal conglomerate containing granitic boulders rests on the granitic rocks and grades upward into feldspathic quartzite. Similar relationships are described by Scott (1969). More generally, the areal distribution of arkosic rocks in the Geikie River area is closely related to that of Archean granitic rocks. The arkosic rocks may represent an immediate weathering product of these granites (Baer, 1969).

Radiometric measurements indicate the presence of at least two distinct thermal events in the Wollaston Lake belt, a Kenoran event and a Hudsonian one. All K-Ar dates on rocks situated in or close to the Wollaston Lake Belt (12 in all) indicate an approximate Hudsonian age (1,570-1,800 m.y.), but none come from the assumed Archean basement rocks. Two groups of analyses by the whole rock Rb/Sr method have been completed, one by Money (1968, p. 1495) in the Needle Falls area, the other by Wanless (personal communication) in the Geikie River area. In both cases the analysed rocks belong to the granitic rocks described above. Money established a best-fit isochron at 2,220 ± 70 m.y., based on four samples. Out of eleven samples, Wanless establishes an isochron at 2,470 ± 65 m.y. (preliminary results). Two of Money's samples plot exactly on the 2,470 m.y. isochron, and if all the data are considered together, the latter date is the best available age for the rocks under discussion. It corresponds very well with the 2,480 m.y. mean age proposed by Stockwell (1969) for the Kenoran orogeny.

Correlations between dated and undated granitic rocks is based on the great similarities of lithology and structural relationships. Table I and Figure 2 indicate for each map-area which rocks can reasonably be considered Archean. In other areas, Archean rocks are probably also present but available information is not sufficient to delineate such bodies at the present time.

Table I: Probable and possible Archean basement rocks, Wollaston Lake fold belt.

Area and Reference	(A): Archean or probably Archean; (B): possibly Archean
Ile-à-la-Crosse (Frarey, 1950)	(A) granite, granite-gneiss, pegmatite northwest of Sandfly Lake; (4)[1] in part.
Needle Falls (Money, 1965, 1967, 1968)	(A) "western granitic rocks", mainly quartz monzonite (12, 12a).
Barnett Lake ($W\frac{1}{2}$) (Money, 1961)	(B) granitic rocks northwest of Barnett Lake, mainly granodiorite and quartz monzonite, (8) in part.
Daly Lake ($E\frac{1}{2}$) (Money, 1966)	(A) "Pederson Lake complex" (1,2,3), granitic and charnockitic rocks, minor amphibolite. (B) "Roper Bay quartz monzonite" (11).
Geikie River (Baer, 1969)	(A) "pink gneissic granite with flaser structure" (4).
Beckett Lake ($E\frac{1}{2}$) (Scott, 1969)	(A) "Johnson River batholith" (10) in part.
Combe Lake (Scott, in prep.)	Foliated fine-grained granite flanking the Courtenay Lake-Cairns Lake fold belt; (A) northwestern side; (B) southeastern side.
Morell Lake ($W\frac{1}{2}$) (Chadwick, 1967)	(B) fine grained and coarse grained "schistose granitic formation" (5) and (6).
Compulsion River fold belt (Møller, 1969)	(B) "basement granite" (2), flaser gneiss (14 in part), red granitic gneiss (13), pink porphyritic granite (9).
Nekweaga Bay ($W\frac{1}{2}$) (Chadwick, 1966)	(B) "southwestern granitic rocks (6).
Hidden Bay (Wallis, 1969)	(B) coarse grained strongly foliated granite (9).

Note 1. Numbers in brackets refer to map-units of each original paper.
2. From Wollaston Lake to the northeast, available information does not allow a distinction between Archean and younger granitic rocks.

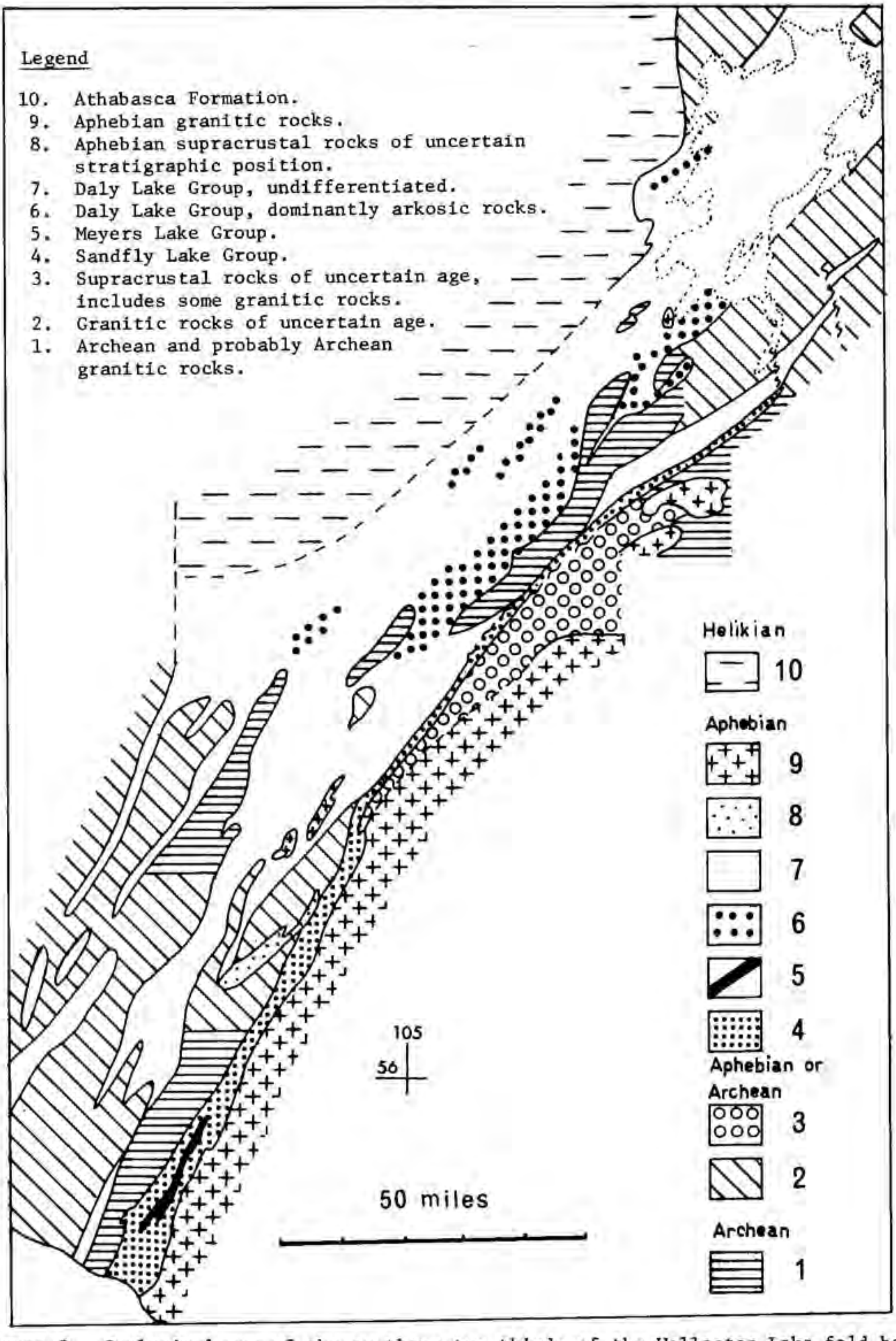


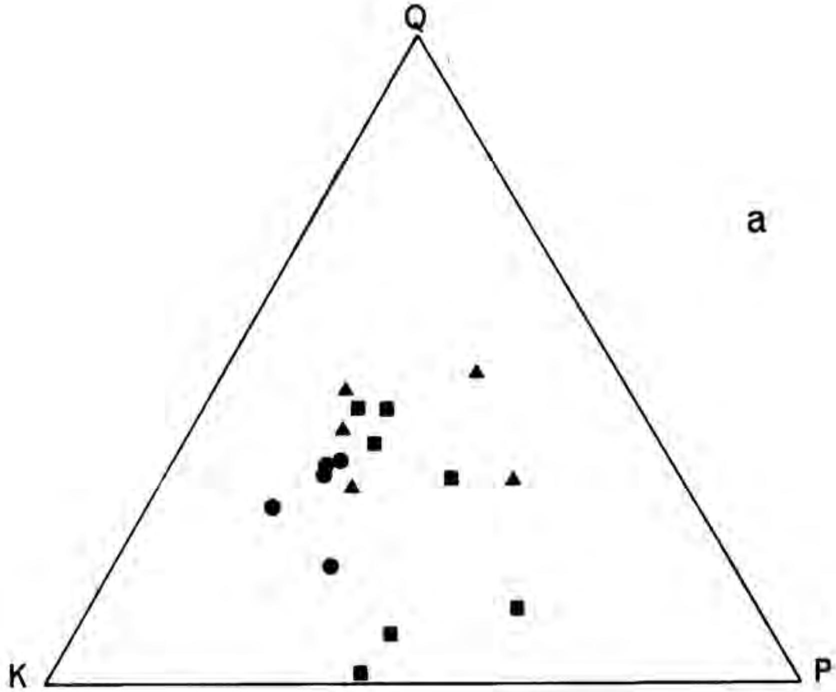
Figure 2. Geological map of the southern two thirds of the Wollaston Lake fold belt.

Table II: Correlation chart of supracrustal rocks, Wollaston Lake fold belt.

Area and reference	Rocks assigned to or probably correlative with: Daly Lake Group	Sandfly Lake Group	Meyers Lake Group	Rocks not correlative with these groups or of uncertain stratigraphic position
Ile-à-la-Crosse (Frarey, 1950)	Garnet-quartz-biotite gneiss (3b)[1]	Hornblende-plagioclase gneiss (3a); Units (1) and (2) in part	Units (1) and (2) in part	
Needle Falls (Money, 1965, 1967, 1968)	Biotite-cordierite-sillimanite rocks and associated rocks (5); derived migmatite (11d)	"Acidic meta-volcanic (?) rocks" (4), biotite and biotite-plagioclase rocks (3) in part, hornblende-biotite rocks and amphibolite (2) in part, "meta-arkose" (1), and derived migmatite and augen gneiss (11a, 11b, 11c, 11e, 11f) TYPE ASSEMBLAGE	Quartzite (8), biotite-muscovite-quartz schist (9), quartz-pebble meta-conglomerate (7) TYPE ASSEMBLAGE	Hypersthene amphibolite (5a), clino-pyroxene amphibolite (5b); biotite rocks, hornblende-biotite rocks, and amphibolite (2) and (3) and derived migmatite and augen gneiss where occurring as inclusions in granitic rocks
Barnett Lake (W½) (Money, 1961)	Biotite-(garnet)-cordierite-sillimanite rocks and associated rocks (4b and 4c)	Hornblende-biotite rocks and amphibolite and interlayered biotite rocks (3), "arkosic" meta-conglomerate (2), "meta-arkose" (1), and derived migmatitic rocks (6) and augen gneiss (7)		Biotite rocks (4, 4a)
Daly Lake (E½) (Money, 1966)	Calc-silicate rocks (9), hornblende-biotite rocks and amphibolite (8) "meta-arkose" and quartzo-feldspathic rocks (7), arkosic meta-conglomerate (6), biotite rocks (5), biotite-cordierite and associated rocks (4) TYPE ASSEMBLAGE			Plagioclase amphibolite (1a) and hypersthene amphibolite (1b) belonging to the Pederson Lake complex of probable Archean age.
Middle Foster Lake (Mawdsley, 1957)	Biotite "granulite" (1), garnet-cordierite-sillimanite gneiss (2), "calcareous metamorphics" (3), (meta-) conglomerate (9a), augen gneiss (4), "banded granitic" gneiss (5)			
Janice Lake (Rath, 1969; Rath and Morton, 1969)		Meta-volcanic? rocks, associated hornblende-biotite rocks		Biotite gneisses and associated rocks (B), "hornblende-biotite" gneiss and associated rocks (A), Juno Lake "meta-arkose"
Geikie River (Baer, 1969)	Biotite gneiss (3), "meta-arkose" (2) except east of Johnson River, hornblende gneisses north of Robertson Lake	Hornblende-biotite gneiss (1) and "meta-arkose" (2) east of Johnson River (in part?)		Hornblende-biotite gneiss (1) in part?
Beckett Lake (E½) (Scott, 1969)	Calc-silicate rocks (7), biotite-cordierite gneiss (6), "meta-arkose" (5) in part, biotite gneiss and "granulite" (4)	"Meta-arkose" (5), hornblende-biotite rocks and amphibolite (1) (all in part)		Hornblende-biotite "granulite" (3), "meta-arkose" (2), and amphibolite (1) in part
Combe Lake (Scott, in preparation)	"Meta-arkose" (9), "mixed meta-sedimentary formation" (8) (in part = Scott, 1969, unit 4)	Courtenay Lake Group Meta-volcanic amphibolite (5), "meta-arkose" and meta-conglomerate (4), biotite "granulite" and gneiss (3)	Souter Lake Group Biotite-sericite-andalusite-quartz schist and "argillite" (7), quartzite (6)	Quartzo-feldspathic rock (2), hornblende "granulite" (1)
Morell Lake (W½) (Chadwick, 1967)	"Mixed meta-sedimentary formation" (4), hornblende-biotite gneiss (3), biotite gneiss (2), "meta-arkose" (1), arkosic meta-conglomerate (1b)			
Compulsion River fold belt (= Courtenay Lake-Cairns Lake fold belt (Møller, 1969)	"George Lake Group" "Slate", "argillite", "arkose", and "schist" (= Scott, in prep., units 8 and 9)	"George Lake Group" "Arkose" ("basal" unit in south part of belt). Equivalent to Scott (in prep.) unit (4). Amphibole gneiss (= Scott, in prep., unit 5)	"George Lake Group" Feldspathic quartzite, quartzite (= Scott, in prep., unit 6)	
Nekweaga Bay (W½) (Chadwick, 1966)	Pyroxenite and associated rocks (4), hornblende-biotite gneiss (3), biotite paragneiss (2), cordierite-biotite paragneiss (2a), "meta-arkose" (1)			
Hidden Bay (Wallis, in prep.). Numbering of units from Wallis (1969)	Migmatized biotite "granulite" (6), biotite paragneiss (5), biotite-plagioclase-quartz "granulite" (4), feldspathic quartzite (3), "meta-arkose" (2) amphibolite "meta-arkose" (1) "Hidden Bay Group" Garnet-actinolite rocks (8), marble and calc-silicate rocks (7)			
Wollaston Lake (Fahrig, 1958)	Meta-sedimentary rocks (1) and migmatized equivalents (3 and 4 in part); hornblendic rocks (2) in part			
Whiskey Jack Lake (Currie, 1961)	Various biotitic rocks and associated cordierite-bearing and other rocks (1, 2, and 3) in part (?)			Feldspathic quartzite (4)
Phelps Lake (Tremblay, 1960)	Quartz-feldspar-biotite gneiss (4) in southeast corner of area (?)			Hurwitz Group (?) Argillite, greywacke, quartzite, and limestone (1)
Kasmere Lake (Fraser, 1962)	Biotite-quartz-feldspar schist, in part cordieritic (4), granite gneiss (5) in part (?), "meta-arkose (1a)			Hurwitz Group (?) Argillite, greywacke, quartzite, and limestone (1b, 2, and 3)
Southeastern Barren Grounds (Wright, 1967)	Undifferentiated supracrustal rocks (12) near Nueltin Lake (?)			Hurwitz Group Units 8, 9, 10, 11

[1]Numbers in brackets refer to map-units of each original paper.

[2]Granulite is frequently used as a textural term (= granofels) in the publications cited. It denotes a granoblastic metamorphic rock with little or no foliation or lineation. "Granofels" is preferred and is used to denote such rocks throughout this paper.



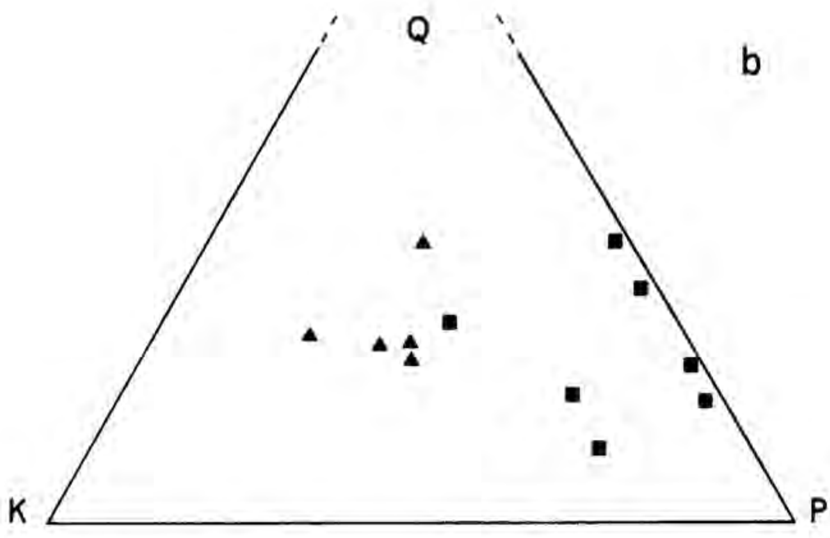


Figure 3. Modal quartz (Q), plagioclase (P), and potassic feldspar (K) of granitic rocks; a: Archean and probably Archean rocks (Kenoran); b: Aphebian rocks (Hudsonian). Dots: Beckett Lake area, triangles: Nekweaga Bay area, squares: Needle Falls area.

Comparison with Proterozoic granitic rocks

Proterozoic granitic rocks are briefly described here for comparison with the Archean granitic rocks.

Proterozoic granitic rocks are commonly grey, more rarely pink, medium to coarse-grained. They are generally homogeneous and lack foliation and schistosity. Major minerals are quartz, potassium feldspar, plagioclase, biotite and hornblende. Megacrysts of feldspar appear in numerous plutons (Nekweaga Bay, Barnett Lake, Needle Falls) but are not characteristic. Sphene, epidote, apatite and zircon are common accessory minerals. Composition of the rocks varies from quartz diorite to granite (Fig. 3).

Proterozoic granitic rocks differ from Archean ones essentially by their structural characteristics. They are commonly massive and display cross-cutting relationships with surrounding supracrustal rocks. Cataclastic structures, common in Archean rocks, are rare. Xenoliths of rocks probably belonging to the Sandfly Lake and Daly Lake Groups have been found in some granitic rocks (Chadwick, 1966; Money, 1967; Wallis, in prep.). Pegmatites are common in most areas studied, and although many were probably formed by segregation during regional metamorphism, others appear to be related to intrusive granites. They contain biotite and/or muscovite, not infrequently tourmaline, occasionally fluorite and locally radioactive minerals. Chilled contacts have been observed in the Needle Falls area (Money, 1965, plate 9B).

A potassium-argon age determination on muscovite from a pegmatite cutting metasedimentary rocks of the Compulsion River area is 1,575 ± 55 m.y. old (Wanless *et al.*, 1970). Biotite from an intrusive fluorite granite in Kasmere Lake area has given an age of 1,735 m.y. (Lowdon *et al.*, 1963) and another intrusive granite from the same region has been dated at 1,700 m.y. (Leech *et al.*, 1963). These dates presumably correspond to the time of emplacement of the granitic bodies.

PROTEROZOIC SUPRACRUSTAL ROCKS

Introduction

The supracrustal rocks within the Wollaston Lake fold-belt are divisible into at least four lithological assemblages. Three assemblages have been included in the Daly Lake, Sandfly Lake and Meyers Lake Groups locally (Money, 1968). A fourth assemblage had been tentatively assigned to the Hidden Bay Group (Scott and Wallis, 1969) but this is now considered to be part of the Daly Lake Group. The rocks assigned to or probably correlative with each group are shown in Table II. The table also includes rocks of uncertain stratigraphic position, and indicates the probable relationship of the Courtenay Lake and Souter Lake Groups (Scott, in prep.) and of the "George Lake Group" (Møller, 1969) to the previously named groups.

Sandfly Lake Group and probably correlative rocks

Introduction

The Sandfly Lake Group and probably correlative rocks are confined to the eastern margin of the Wollaston Lake fold-belt. They occur from the southern end of the fold-belt (Sandfly Lake Group proper) to near the north end of the Courtenay Lake-Cairns Lake branch of the fold-belt, a distance along stike of about 180 miles. Mapping is not sufficiently advanced to prove that these rocks are continuous and they have been referred to as the Courtenay Lake Group (Scott, in prep.) and George Lake Group (Møller, 1969) north of 57°30'N. The name "George Lake Group" is rejected here because the group is named after an informally named "George Lake" and not after the "official" George Lake which occurs in another part of the Wollaston Lake Belt. The "George Lake Group" may include part of three separate and distinct groups (Table II). The name Courtenay Lake Group should be retained pending further mapping and is extended to include the appropriate part of the "George Lake Group".

Lithology

The rock types occurring within the Sandfly Lake Group and probably correlative are shown in Table II. The lithology and inferred origin of each type are discussed below:

1. "Acidic meta-volcanic (?)" rocks (Needle Falls)[1]; meta-volcanic (?) rocks (Janice Lake).

These rocks are exceptionally fine grained (average grain size 0.1 to 0.2 mm, Needle Falls area), "dense" to cherty in appearance, and have very close jointing and an almost conchoidal fracture in places. No textural features definitely indicative of origin are preserved. Layers, generally 1/10 to 1/4 inch thick, occur locally in the Needle Falls area. Similar layering is reported in the Janice Lake area. A plot of quartz and feldspar (Fig. 4) for thin sections show very little scatter (except for one unusually biotite-rich section) and a nearly constant quartz-feldspar ratio. These rocks are quartzo-feldspathic and may have been derived from arkoses or felsic volcanic rocks or tuffs. The textural features and the nearly constant quartz-feldspar ratio, so unlike the highly variable ratios in "meta-arkose" of the fold belt, strongly suggest a volcanic origin. Two chemical analyses are available (Rath, 1969; Money, 1967) and both indicate that the rocks have the composition of rhyodacite. The biotite-rich thin-section may represent tuffaceous material. The fine layering may be sedimentary layering in tuffs or may be flow banding.

2. Biotite and biotite-plagioclase rocks (Needle Falls); biotite rocks (Barnett Lake); biotite gneiss and "granofels"[2] (Combe Lake).

These rocks are generally equigranular but porphyroblastic rocks occur in the Needle Falls area. The average grain size in the equigranular varieties is about 0.6 to 1.0 mm. Textures are due to metamorphism and provide no evidence of origin. Chemical analyses of two samples from the Needle Falls area suggest derivation from greywacke (Money, 1967). A similar origin is assumed for samples from the Combe Lake area which are mineralogically similar and have little or no K-feldspar. Samples from the Barnett Lake area are quartz-rich (44 to 47%) and contain appreciable K-feldspar (12 to 15%) and may have been pelites or argillaceous sandstones originally. Further work may demonstrate that the biotite rocks are an intercalated greywacke-pelite sequence.

3. Hornblende-biotite rocks (Needle Falls, Barnett Lake, Janice Lake, Geikie River, Beckett Lake).

These rocks are texturally similar to the equigranular biotite rocks. They have undergone complete metamorphic recrystallization and no sedimentary or volcanic textures have been observed. Chemical analyses of two samples from the Needle Falls area suggest that the most plausible origin is derivation from intermediate volcanic rocks (andesite or dacite) or tuffs or greywacke derived from or contaminated by volcanic rocks. Large variations in SiO_2 content suggest that the hornblende-biotite rocks are, at least in part, tuffaceous or that they are greywackes. The composition of both analysed samples, although within the range reported for igneous rocks, is characterized by unusually high normative ab:an ratios for rocks of their normative quartz content and some Na-metasomatism or albitization is suspected (Money, 1967).

4. Amphibolite (Needle Falls, Barnett Lake, Beckett Lake); meta-volcanic amphibolite (Combe Lake); amphibole gneiss (Compulsion River fold belt).

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[1]See Table II for reference to original publications, and Figure 1 for location of map-areas.

[2]Referred to as "granulites" in the original publication.

These rocks have typical metamorphic textures and generally have not retained any textural features indicative of their origin. However, in the Combe area, Scott (in prep.) has reported pillows, volcanic bombs, and blocks. The blocks are in a matrix of biotite granofels (greywacke?). Before this evidence was found, it had been concluded (Money, 1967) on the basis of two chemical analyses that the amphibolites in the Needle Falls area were probably derived from alkali basalts. It is suggested that the amphibolites are in part or mainly derived from basaltic volcanics, but the possibility cannot be excluded that they may consist in part of andesitic volcanics and/or tuffaceous rocks.

5. "Meta-arkose" and "arkosic" meta-conglomerate (Needle Falls, Barnett Lake, Geikie River, Beckett Lake, Combe Lake); "arkose" (Compulsion River fold belt).

These arkosic rocks contain local conglomeratic horizons and show local weak colour-layering (= bedding?) throughout most of the Wollaston Lake fold belt. In the Combe Lake area crossbedding and ripple-marks are preserved and one local unconformity has been recognized. This consists of conglomerate, locally rich in muscovite-schist cobbles, overlying a grey muscovite schist. Cobbles, pebbles, and boulders elsewhere are generally quartzitic or quartzo-feldspathic. Some consist of quartz, others of fine- to coarse-grained unfoliated granitic rocks (quartz monzonite, granodiorite or pegmatite), and others of various fine-grained quartzo-feldspathic rocks. Some of these fine-grained rocks may be aplite. However, in part they also consist of arkosic material or are acidic tuffs. Evidence for this is provided by the presence of detrital quartz grains and by an abnormally high quartz content for an igneous rock. For example, one cobble from the Barnett Lake area contains 54% quartz. It is uncertain whether such cobbles result from local erosion of the Sandfly Lake Group and probably correlative rocks or are derived from older rocks. However, in the Combe Lake area they are mica-poor and unlike any known "meta-arkose" within the Courtenay Lake Group. A cobble or lens of amphibolite was observed in "meta-arkose" in the Needle Falls area. The "meta-arkose" itself is characterized by the presence of up to 40% muscovite and a lack of or very low content of biotite, in contrast to the "meta-arkose" of the Daly Lake Group which lacks muscovite and is characterized by biotite with or without amphibole or clinopyroxene. Quartz-feldspar ratios (Fig. 4) show comparatively little variation, suggesting that only a limited sorting took place. The high muscovite content of most samples and high Al_2O_3 content of two chemically analysed samples from the Needle Falls area suggest that the rock originally contained an appreciable amount of clay minerals. It is inferred that most of the "meta-arkose" was an arkosic wacke (Gilbert *in* Williams, Turner, and Gilbert, 1955) or a feldspathic greywacke (Pettijohn, 1957). Mica-poor varieties may have been arkosic arenites.

Relationships of rock types

In the Needle Falls area, hornblende-biotite rocks and amphibolite are invariably interlayered and cannot be mapped separately. Individual layers are about one-half inch to at least 50 feet in thickness and generally are traceable for the full length of outcrops, i.e. up to several hundred feet. Both rock types occur interlayered with the "acidic meta-volcanic (?) rocks", the biotite and biotite-plagioclase rocks, and the "meta-arkose". In the case of the "acidic meta-volcanic (?) rocks" the individual layers are about one-half inch to 30 feet thick. Layers of the other rocks are generally of the order of 20 feet or more. The "meta-arkose" and biotite rocks are interlayered locally but neither has been seen in areal association with the "acidic meta-volcanic (?) rocks". Contacts between all rock types appear to be conformable. Relationships of rock types appear to be similar in other areas except for one locality in the Combe Lake area (Scott, in prep.) where amphibolite and associated rocks may overlie crossbedded "meta-arkose" with angular unconformity. The presence of a local unconformity within the "meta-arkose" in the Combe Lake area and of a possible cobble of amphibolite in "meta-arkose" of the Needle Falls area may indicate that there are other unrecognized local unconformities within the group.

In the Needle Falls area, a crude estimate of relative abundance of different rock types in the Sandfly Lake Group is 55% amphibolite, hornblende-biotite rock and derived migmatite and 15% of each of the "acidic meta-volcanic (?) rocks", "biotite and biotite-plagioclase rocks", and "meta-arkose". In the Barnett Lake area the most abundant rock types are hornblende-biotite rocks and amphibolite. They form as much as 80% of the group if derived migmatite is included with them. "Meta-arkose" and meta-conglomerate form about 20% of the group. Farther north, in the Combe Lake area, the abundant rock types are "meta-arkose" (about 80%) and meta-volcanic amphibolite (about 20%). In addition to the apparent trend from abundant volcanic rocks in the south to abundant "arkosic" rocks in the north, there are other gross lithological changes. Biotite rocks occur in only minor amounts north of the south half of the Needle Falls area, unless the rocks of map-units (4) and (4a) in the Barnett Lake area belong to the Sandfly Lake Group. "Acidic meta-volcanic (?) rocks" appear to occur locally in the southern part of the group and have not been mapped north of the Janice Lake area.

Thickness

Thickness of the Sandfly Lake Group and probably correlative rocks is difficult to estimate because of isoclinal or nearly isoclinal folding and a general lack of marker horizons. In the Needle Falls area the present thickness may be as little as 1,500 feet locally and is unlikely to exceed 5,000 feet anywhere (Money, 1968). In this area the group is overlain, probably unconformably, by another group of rocks and may have undergone considerable erosion. In the Barnett Lake area, the thickness may be of the order of 8,000 feet. In the Combe Lake area, depending on the structural interpretation, the thickness may be as little as 1,000 feet or as much as 5,000 feet. In this area the group may have been eroded before deposition of subsequent groups and the base of the group may have been assimilated by granitic rocks. For the Sandfly Lake Group, an original maximum thickness of perhaps 5,000 to 10,000 feet appears reasonable.

Inferred depositional environment

The Sandfly Lake Group and probably correlative rocks were probably derived from basic to acidic volcanic rocks, arkosic wacke or feldspathic greywacke, and greywacke. It may be inferred that deposition was rapid and took place in an area of crustal instability. There is evidence for sialic basement (Archean (?) plutonic rocks, see p. 5 sq.). It is impossible to explain the abundant arkosic rocks within the group unless older sialic rocks were exposed nearby. The acidic and intermediate volcanic rocks may be attributed to partial fusion of the basement or to contamination of basic magmas rising through it. Greywacke (biotite and biotite-plagioclase rocks) and volcanic rocks appear to be more abundant toward the south and arkosic wacke (with cross-bedding) becomes predominant to the north, facts which may indicate a change in the depositional environment.

Meyers Lake Group and probably correlative rocks

Introduction

The Meyers Lake Group and probably correlative rocks have been found in the Needle Falls area and in and northeast of the Combe Lake area. They appear to be absent throughout all or most of the intervening area. They have been referred to as the Souter Lake Group in the Combe Lake area. The name Souter Lake Group is retained pending further mapping and this group is extended to include the appropriate part of Møller's (1969) "George Lake Group" (*see* Table II).

Lithology

The Meyers Lake Group, in the Needle Falls area, consists of quartz-pebble meta-conglomerate, various quartzites, and biotite-muscovite-quartz schist with or without andalusite or sillimanite. The Souter Lake Group lacks conglomerate. It consists mainly of quartzite virtually identical to that in the Meyers Lake Group of the type area, with lesser amounts of biotite-sericite-andalusite-quartz schist and 'argillite'.

The quartz-pebble meta-conglomerate is oligomictic except for the presence of possible 'pebbles' of pink feldspar. Some and perhaps all of these 'pebbles' are porphyroblasts. The quartz pebbles are glassy, grey, or white and appear to have been derived from vein or pegmatitic quartz. The matrix consists either essentially of quartz and biotite or of quartz, feldspar, and muscovite.

Common varieties of quartzite include "pure" quartzite, feldspathic quartzite, and actinolitic quartzite. All thin sections studied, except very actinolite-rich ones, contain 70 per cent or more of quartz, and the 'pure' quartzite contains 90 per cent or more. Other major components are feldspar, predominantly K-feldspar, and muscovite and in some cases biotite. Bedding is recognizable in places within the quartzites and crossbedding was observed at one locality. A cobble of amphibolite has been reported in quartzite of the Combe Lake area (Scott, in prep.). The "pure" quartzite appears to be metamorphosed orthoquartzite and the feldspathic quartzite is metamorphosed feldspathic sandstone or feldspathic arenite (Fig. 4). The actinolitic quartzite represents variants of the feldspathic sandstone and perhaps of the orthoquartzite that originally had a calcareous cement.

Pink feldspathic quartzites also occur, that are mineralogically like the "meta-arkose" belonging to the Sandfly Lake Group. They were probably derived from arkosic wacke or feldspathic greywacke.

The various mica schists are essentially biotite-muscovite-quartz rocks with or without andalusite and sillimanite. Feldspar is rare. The mineralogy, and three chemical analyses of samples from the Needle Falls area, indicate derivation from quartz-rich, feldspar-poor pelitic rocks (Money, 1967).

Relationships of rock types

In the Needle Falls area a basal unit and an upper unit are distinguishable. The basal unit generally has an apparent thickness of about 100 to 400 feet but it is probably absent locally. It consists predominantly of quartz-pebble meta-conglomerate with minor interlayered quartzite and mica schist. The upper unit is predominantly quartzite (about 80%) with interlayered mica schist (about 20%) and minor quartz-pebble meta-conglomerate. The quartzites are estimated to be at least 65% "pure quartzite" and not more than 35% feldspathic and actinolitic varieties. The pink feldspathic quartzite is not known to occur more than 300 feet above the top of the basal unit. It is uncommon and is always in layers less than 5 feet thick. Similar quartzite or "meta-arkose" occur at the base of the basal unit in places but it is uncertain whether this is part of the Sandfly Lake Group or of the Meyers Lake Group.

In the Combe Lake area quartzites make up about 70% of the group and mica schist and argillite about 30%. The rock types are interlayered but three main schist units, separated by quartzites, are distinguished. Argillite is confined to the northwestern part (top?) of the group.

Thickness

In the Needle Falls area the Meyers Lake Group has an estimated thickness of at least 1,500 to 2,400 feet. The maximum present thickness is of the order of 3,000 feet. In the Combe Lake area maximum thickness is in the order of 2,500 feet.

Inferred depositional environment

The Meyers Lake Group consists in large part of relatively mature sedimentary rocks and bears a closer resemblance to assemblages characteristic of a stable shelf than to other standard lithological assemblages (Pettijohn, 1957; Krumbein and Sloss, 1958). Nevertheless, the presence of abundant feldspathic arenites and minor arkosic wacke indicates much more rapid deposition or a less stable environment than the typical shelf. Quartz-pebble (oligomictic) conglomerates are commonly deposits of a transgressive beach over a surface of low relief (Pettijohn, 1957), although similar conglomerates may also be fluvial. The extent of the conglomerates along strike (at least 16 and possibly 30 miles), the interlayered pelites, and the extent of the group as a whole (approximately 160 miles if the Souter Lake Group is correlative) suggest marine deposition for the group. One recognizable example of crossbedding is preserved, indicating deposition above the wave base. The possibility remains that the group consists in part of aeolian (beach dune sands) or fluviatile quartzites. In any case it was deposited, probably in or near shallow water, during a period of relative stability of this area.

Daly Lake Group and probably correlative rocks

Introduction

These rocks form most of the Wollaston Lake fold-belt except for its eastern margin. Part of the probably correlative rocks have been included in the "George Lake Group" (Møller, 1969) and the "Hidden Bay Group" (Scott and Wallis, 1969). Both names are now rejected. The "Hidden Bay Group" and associated feldspathic quartzite will be referred to as the Hidden Bay assemblage in the following discussion.

Lithology

The lithology and inferred origin for each rock type included within or probably correlative with the Daly Lake Group are discussed below:

1. Biotite, biotite-cordierite, biotite-cordierite-sillimanite, and related rocks (most map-areas).

This division includes a wide variety of equigranular to porphyroblastic biotite gneisses and granofels. No sedimentary textures have been observed in most areas other than compositional layering which probably parallels bedding. In the Hidden Bay area, a rock occurs which has been interpreted as a metamorphosed mud-flake conglomerate. This consists of interlayered biotitic and arkosic rocks about 6 inches thick. The biotitic layers contain dispersed angular, flat, biotite-rich fragments. The fragments are separated from one another, but probably not tectonically, and are explainable as mud flakes incorporated in a silty matrix. Polymictic meta-conglomerate also occurs within the biotitic rocks in the Hidden Bay area. The clasts are mainly granite or arkose (80-90%) but include biotite amphibolite, quartzite, and biotitic meta-sedimentary rocks. Mineralogically, the biotite rocks fall into two main subdivisions, cordierite-bearing rocks and cordierite-free rocks. The cordierite-bearing rocks may contain garnet and commonly contain sillimanite, particularly in the southern third of the fold-belt. Chemical analyses of two samples from the Needle Falls area indicate that these rocks were pelitic (Money, 1967). The cordierite-free rocks generally lack garnet and sillimanite but contain amphibole or rarely clinopyroxene locally. No chemical analyses are available. Money (1966) and Scott (1969) suggest that varieties of the biotite rocks poor in K-feldspar may have been derived from greywacke.

2. "Meta-arkose" and quartzo-feldspathic rocks (most map-areas); arkosic meta-conglomerate (Daly Lake, Middle Foster Lake, and Morell Lake areas).

This division includes rocks consisting essentially of quartz and feldspar and containing less than ten per cent biotite. The quartz-feldspar ratio (Fig. 4) is highly variable and indicates that the "meta-arkose" belonging to this group has undergone more sorting than that belonging to the Sandfly Lake Group. Certain distinctive and characteristic varieties contain small green spots of amphibole or rarely clinopyroxene. On close examination those are commonly found to be skeletal porphyroblastic crystals, probably derived from a calcareous cement. Weak colour-layering may locally indicate bedding. Undoubted bedding is indicated in places by the presence of layers with markedly different concentrations of amphibole and clinopyroxene. Crossbedding is recognizable locally in the Morell Lake and Nekweaga Bay areas. The "meta-arkose" is, at least in part, derived from arkose. The low mica content suggests that these were predominantly arkosic arenites but they may have been in part arkosic wackes. The highly variable quartz-feldspar ratio confirms the postulated sedimentary origin of the "meta-arkose". However, the possibility cannot be entirely precluded that this unit may include minor amounts of acidic volcanic rocks.

The arkosic meta-conglomerate is of very limited extent. It contains boulders and cobbles of milky quartz, of a fine-grained aplite or arkose, and of fine- to coarse-grained igneous rocks. These are estimated to range in composition from granite to quartz monzonite to diorite or quartz diorite.

3. Feldspathic quartzite, Hidden Bay area.

Feldspathic quartzite occurs in the Hidden Bay area immediately below and intercalated with the calc-silicate rocks and marble assigned to the Hidden Bay assemblage. Modal analyses indicate that the quartz content is about 60 to 75 per cent. The dominant rock type is fine to coarse grained and consists of well rounded quartz grains in a matrix of feldspar. This is overlain by a finer grained variety containing layers of quartz-pebble meta-conglomerate that are rarely more than one foot thick. These can be followed for up to 50 feet along strike. Graded bedding from coarse angular grains to small rounded grains is locally discernible in the finer grained variety of quartzite. The unit may be the equivalent of rocks forming part of Fahrig's (1958) unit (1) and described as: "... pebbly conglomerate layers consisting of white quartzite pebbles in a biotitic matrix ..." and farther: "The conglomerate generally is interbedded with rusty-weathering meta-quartzite".

4. Calc-silicate rocks (Daly Lake, Beckett Lake); "calcareous metamorphics" (Middle Foster Lake); pyroxenite and associated rocks (Nekweaga Bay); garnet-actinolite rocks, marble, and calc-silicate rocks (Hidden Bay).

Calc-silicate rocks are widespread within the Daly Lake Group but in many areas doe not form mappable units. They are highly variable in mineralogy and composition, even within a small area. Many of the rocks mapped as calc-silicates are simply unusually calcareous "meta-arkose" or biotite rocks and were probably originally limey pelitic rocks, arkose, or perhaps greywacke. No rocks are known south of the Hidden Bay area which may have been pure carbonate rocks. Pyroxenite in the Nekweaga Bay area is of uncertain origin but may be derived from calcareous sediments. Calcareous rocks in the Hidden Bay area appear to be stratigraphically equivalent to the Daly Lake

Group although forming part of the distinct Hidden Bay assemblage. The calcareous rocks within this sequence include marble in layers as much as 150 feet thick.

5. Hornblende-biotite rocks and amphibolite (Daly Lake, Morell Lake, Nekweaga Bay); hornblende gneisses north of Robertson Lake (Geikie River).

These rocks are somewhat variable in composition. No textural features indicative of origin have been noted. In part they are probably hornblende-rich calc-silicate rocks of sedimentary origin but they may be in part of volcanic origin.

6. "Mixed metasedimentary formation" (Combe Lake, Morell Lake) and equivalent rocks in the Beckett Lake area and Compulsion River fold-belt.

The "mixed metasedimentary formation" consists of a wide variety of thinly interlayered metasedimentary rocks. Individual layers in the Morell Lake area are commonly of the order of eight inches in thickness. The rock types making up the unit include rocks described (Chadwick, 1967) as grey-brown pelitic schists, black graphitic schists, pale grey-white and grey-green quartzites, and blue-grey, quartz-rich semi-pelites. Calcareous varieties of all rock types occur. Scott (in prep.) describes the unit as consisting predominantly of thinly laminated biotite "granulite" (= granofels) and argillite of mixed pelitic and greywacke ancestry, with minor arkose, feldspathic quartzite, and calc-silicate rocks. The "arkose" and feldspathic quartzite are commonly calcareous and the biotite granofels and argillite are calcareous in part. Syngenetic(?) sphalerite, galena, and pyrite occur locally (Pyke and Partridge, 1967; Møller, 1969). Scott has noted one possible example of graded bedding accompanied by features indicative of primary slumping and faulting in unconsolidated or weakly consolidated sediments and one case of primary (?) thrust faulting in such sediments. The differences in terminology between Chadwick (1967) and Scott (in prep.) appear to be largely due to differences in texture and not in composition. Chadwick's quartzites and quartz-rich semi-pelites are probably equivalent to Scott's feldspathic quartzites. No chemical analyses are available but the postulated origin of these rocks as a finely laminated sequence of partly calcareous pelites, semi-pelites (greywacke?), feldspathic quartzite, and arkose appears to be reasonable.

Relationships of rock types

Throughout most of the Wollaston Lake fold-belt, the Daly Lake Group and probably correlative rocks consist predominantly of biotite-rich rocks and "meta-arkose". Together these generally form well over 90% of the group. They generally occur in subequal amounts but either may locally form up to two-thirds of the group. The cordierite-bearing biotite rocks are gradational to the cordierite-free biotite rocks which in turn are gradational to calcareous variants of the biotite rocks and to the "meta-arkose". In the Daly Lake area, where the rocks are comparatively well exposed and traces of axial planes can be delineated, lithological units consisting predominantly of each of the major rock types are probably repeated several times throughout the stratigraphic column. Such units appear to be about 500 to 2,000 feet thick.

The "mixed metasedimentary rocks" are of uncertain relationship to the rocks probably correlative with the Daly Lake Group in the Combe Lake area. In the Morell Lake area pelitic schists belonging to this unit are gradational to biotite paragneiss which is in turn gradational to meta-arkose, both of which are characteristic of the Daly Lake Group. It is on the basis of this relationship and of the calcareous nature of many of the rocks within the "mixed metasedimentary rocks" that they are considered to be probably correlative with the Daly Lake Group.

The Hidden Bay assemblage underlies most of the Hidden Bay area northwest of Hidden Bay. The assemblage contains more than 50% of calcareous rocks and feldspathic quartzites and less than 50% of biotite granofels, and is decidedly atypical for the Daly Lake Group. It appears nevertheless, to be overlain and underlain by rocks typical of this group.

Thickness

In the Daly Lake area the thickness of the Daly Lake Group can hardly be less than 5,000 feet and a more realistic estimate is probably in the order of 10,000 feet or more. Estimates elsewhere are of little value because axial planes of folds have not been located. There is no evidence to suggest that probably correlative rocks are considerably thinner or thicker anywhere in the fold belt.

Inferred depositional environment

The typical Daly Lake Group and probably correlative rocks consist mainly or wholly of sedimentary rocks, mainly pelites and arkosic arenites but probably including arkosic wacke and greywacke. Many rock types are somewhat calcareous. Volcanic rocks are rare or non-existent. Deposition is inferred to have taken place in an area of crustal instability, in close proximity to a sialic landmass, and probably on a sialic basement, and at least in part in a marine environment. Crossbedding in arkoses suggests shallow water deposition.

Too little is known about the Hidden Bay assemblage for a profitable speculation on its depositional environment. Nevertheless, the presence of units of carbonate rocks and oligomictic quartz-pebble conglomerate suggest that the assemblage was deposited in a comparatively stable area or during a period of greater crustal stability than the rest of the Daly Lake Group and probably correlative rocks.

Other supracrustal rocks

Introduction

Most of the supracrustal rocks within the Wollaston Lake fold-belt are probably correlative with one or other of the previously discussed groups. However, there are rocks of uncertain stratigraphic position in a number of map-areas (Table II) and also some Archean (?) supracrustal rocks (previously discussed) near the fold-belt. The rocks of uncertain stratigraphic position are, in part, in areas that have not been mapped in detail. Generally, so little information is available about them that it is futile to discuss them. Rocks of uncertain stratigraphic position that occur in areas which have been mapped in comparative detail are briefly discussed in the following sections.

Needle Falls area

In this area, a variety of rocks occur as inclusions or isolated patches in granitic rocks both east and west of the branch of the Wollaston Lake fold-belt named the Barnett Lake-Needle Falls fold belt (Money, 1968). Certain of these rocks, including biotite rocks, hornblende-biotite rocks, and amphibolite, are mineralogically similar to rock belonging to the Sandfly Lake Group, and such rocks may belong to this group wholly or in part. Hypersthene amphibolite and clinopyroxene amphibolite also occur. It is possible that the hypersthene amphibolite is Archean as it has undergone granulite facies metamorphism like the probably Archean Pederson Lake complex, but its occurrence as a small remnant surrounded by granitic rocks precludes any positive correlation. Chemically and mineralogically it is unlike any of the rocks assigned to the Sandfly Lake Group or other named groups. Clinopyroxene

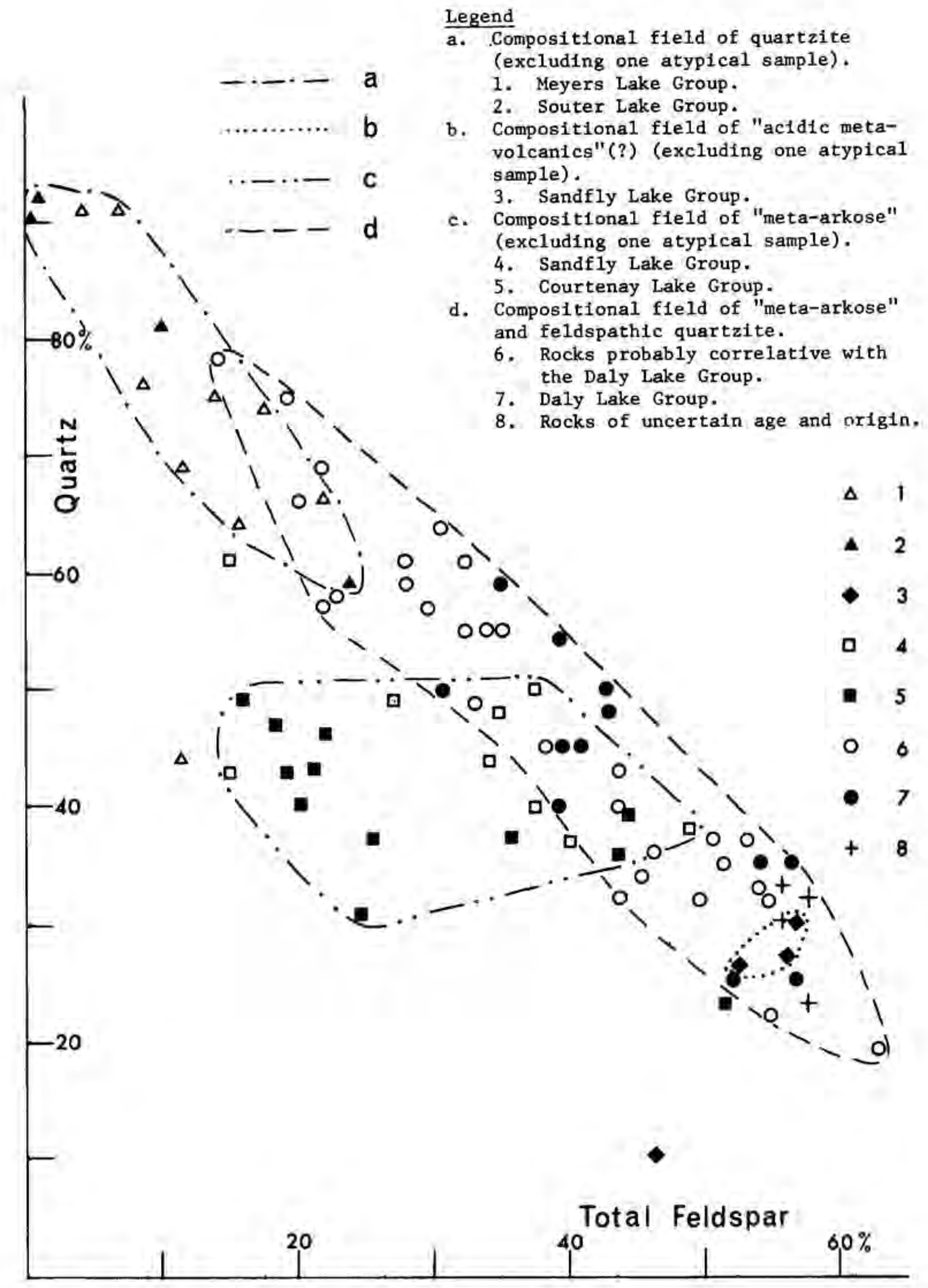

Figure 4. Modal quartz : feldspar ratios in quartzo-feldspathic supracrustal rocks.

amphibolite occurs on a small island in Sandfly Lake. A chemical analysis suggests derivation from a sedimentary calc-silicate rock. This rock type could belong to the Daly Lake Group.

Barnett Lake area, west half

Biotite rocks (4, 4a) occur in a small area surrounded by granitic rocks near the southeast corner of the map-area and also along the west side of the Barnett Lake-Needle Falls fold-belt. The rocks in the southeast corner are of unknown stratigraphic position. The rocks along the western side of the fold-belt contain garnet and appreciable plagioclase and are unlike typical biotite rocks in either the Sandfly Lake or Daly Lake Groups. They have been considered as being possibly gradational between the biotite rocks of both groups. The groups would thus be interpreted as representing different but contemporaneous sedimentary facies (Money, 1968). This interpretation probably is no longer acceptable (*see* p. 28). The stratigraphic position of the biotite rocks is uncertain.

Janice Lake area

The Janice Lake area comprises about six square miles and is at approximately 50°50'N, 104°56'W. The extreme southeast corner of the area is underlain by rocks probably correlative with the Sandfly Lake Group. The rocks in the rest of the area are a probable metasedimentary assemblage subdivided by Rath (1969) into (1) hornblende-biotite gneiss, "minor meta-arkose", and some conglomeratic horizons (map-unit A); (2) biotite gneiss, minor quartzitic biotite gneiss, feldspathic biotite gneiss, and laminated biotite gneiss (map-unit B); (3) calc-silicate rocks; and (4) Juno Lake meta-arkose. Modal analyses and one chemical analysis suggest that the rocks belonging to map-unit A were probably originally in part arkosic wacke and in part arkosic arenite. The hornblende-biotite gneiss is heterogeneous on a scale of a few inches, consisting of irregularly distributed quartz-rich or feldspar-rich patches. The associated meta-conglomerates have quartzo-feldspathic matrices and contain a variety of pebbles, cobbles and boulders. The most abundant clasts are quartz-rich, but "granitic" and hornblende-rich varieties occur. Thin sections of two "granitic" boulders indicate that one has the composition of a granodiorite and the other that of a quartz monzonite. Syngenetic (?) chalcocite occurs in the meta-conglomerate and associated arkosic rocks. The rocks comprising map-unit B and the Juno Lake "meta-arkose" appear to be mainly derived from arkosic arenite but may include some arkosic wacke. The entire assemblage may belong to the Daly Lake Group but correlation is considered to be uncertain until further mapping has been done. Like the Daly Lake Group, the rocks of the area were probably deposited rapidly in an area of crustal instability. They were probably derived from a mainly granitic terrane.

Geikie River, Combe Lake, and Beckett Lake (east half) areas

The rocks of uncertain stratigraphic position in these areas comprise hornblende-rich rocks variously referred to as hornblende granulite, (= granofels)[1] hornblende-biotite granulite (= granofels) or gneiss, or amphibolite, and quartzo-feldspathic rocks which have also been referred to as "meta-arkose". These rocks occur within a granitic terrane, as easterly-trending remnants and patches east of the area underlain by the Courtenay Lake Group. The origin of the hornblende-rich rocks is uncertain, as original textural features are not preserved, but a relative uniformity of composition and grain size and lack of obvious calc-silicate assemblages suggest they were volcanic. The quartzo-feldspathic rocks locally contain rounded quartz grains, suggesting a sedimentary or tuffaceous origin, but no primary structures have been

[1]See note Table II.

observed. They have a very limited range of quartz:feldspar ratios (Fig. 4, No. 8), like the "acidic meta-volcanic (?) rocks" belonging to the Sandfly Lake Group and unlike arkosic rocks, suggesting a volcanic or tuffaceous origin.

The hornblende-rich and quartzo-feldspathic rocks are spatially associated and locally are gradational to each other (tuffaceous?). They occur immediately east of the Courtenay Lake Group and are mineralogically very similar to the rocks forming the eastern part of the Sandfly Lake Group (interlayered amphibolite, hornblende-biotite gneiss, and "acidic meta-volcanic (?) rocks"). They may be correlative with the Courtenay Lake and Sandfly Lake Groups. If they are, the difficulty is to explain the easterly trend of their folding, which is characteristic of rocks folded during the Kenoran orogeny and unusual for rocks folded during the Hudsonian orogeny.

Relationships of the recognized groups

It is postulated that the oldest group within the Wollaston Lake fold-belt is the Sandfly Lake Group (= Courtenay Lake Group) and that this is overlain unconformably by the Meyers Lake Group (= Souter Lake Group) which in turn is overlain conformably by the Daly Lake Group.

In the Needle Falls area, crossbedding in quartzite immediately above the basal meta-conglomerate unit belonging to the Meyers Lake Group indicates that this group overlies the Sandfly Lake Group. Attitudes of minor folds indicate that cores of synclines in the Sandfly Lake Group are occupied by the Meyers Lake Group. An unconformity is suggested on the basis that (1) a meta-conglomerate is the lowest unit of most of the Meyers Lake Group; (2) in different places the Meyers Lake Group probably overlies hornblende-biotite rocks, amphibolite, biotite-plagioclase gneiss, biotite gneiss, and meta-arkose (suggesting an angular unconformity); and (3) the Meyers Lake Group and the Sandfly Lake Group were deposited in different sedimentary environments.

In the Combe Lake area, numerous top determinations based on the attitudes of pillows and of crossbedding in rocks belonging to the Courtenay Lake Group indicate that this group is overlain by the Souter Lake Group. The presence of an amphibolite cobble in the Souter Lake Group may indicate that the groups are unconformable. The Souter Lake and Courtenay Lake Groups are lithologically equivalent to and probably correlative with the Meyers Lake and Sandfly Lake Groups respectively. The relationship of the Souter Lake Group to the Courtenay Lake Group is the same as that of the Meyers Lake Group to the Sandfly Lake Group, confirming the suggested correlation. The Souter Lake Group is gradational to the "mixed metasedimentary formation", the contact zone consisting of alternating "argillite" and quartzite beds. Although sparse outcrop allows several possible structural interpretations, the most plausible is that the "mixed metasedimentary formation" overlies the Souter Lake Group. The attitude of the only known example of graded bedding in the "mixed metasedimentary formation" supports this interpretation. As the "mixed metasedimentary formation" is gradational to rocks typical of the Daly Lake Group (Chadwick, 1967), the Daly Lake Group or probably correlative rocks presumably overlie the Souter Lake Group (= Meyers Lake Group) conformably.

Age of the supracrustal rocks

All of the supracrustal rocks within the Wollaston Lake fold-belt were involved in the Hudsonian orogeny. K-Ar dates for metamorphic micas (Table III) indicate an age of approximately 1,750 m.y. for the last main phases or phase of this orogeny.

Table III: K-Ar ages of metamorphic micas, Wollaston Lake fold-belt.

Sample	Mineral	Rock Type	Apparent age (m.y.)	Reference
GSC 60-67	Biotite	Sillimanite paragneiss (Daly Lake Group?)	1,670 ± 90	Lowdon, 1961
614-30-8a	Muscovite	"Meta-arkose" (Sandfly Lake Group)	1,730 ± 90	Money, 1968
614-47-9	Muscovite	Biot.-musc.-qtz. schist (Meyers Lake Group)	1,780 ± 90	Money, 1968
	Biotite		1,820 ± 90	Money, 1968
614-S3	Biotite	Hornblende-biotite gneiss (Sandfly Lake Group?)	1,570 ± 80	Money, 1968

The age of sedimentation of the supracrustal rocks is unknown. In view of their occurrence in a fold belt with a northeasterly "Hudsonian" trend, their predominantly metasedimentary nature, the abundance of arkosic rocks, and evidence for the presence of an Archean granitic basement, it seems probable that they were deposited during Aphebian time. It is possible that the Sandfly Lake Group and Courtenay Lake Group, i.e. the partly volcanic assemblages along the eastern margin of the belt, are Archean. The writers do not favour this interpretation.

BASIC PLUTONIC ROCKS

Although some basic plutonic rocks are known from the gneissic terrane flanking the Wollaston Lake fold-belt, occurrences within the belt itself are scarce. The "epidiorite" of the Needle Falls area, a medium to coarse grained massive rock consisting of hornblende and plagioclase with minor biotite, forms the largest body. It is about six miles long and 3/4 mile wide. In the Geikie River area, a group of massive outcrops of gabbro has been reported. The rock is formed in about equal parts of plagioclase and pyroxene or hornblende. Isolated outcrops of pyroxenite in the Nekweaga Bay area are formed almost entirely of diopside. Their origin is uncertain, as according to Chadwick (1966, p. 17), they might be metamorphosed calcareous sedimentary rocks. Scapolite gabbro is known only from the Daly Lake area. It displays igneous layering and is formed essentially of clinopyroxene and scapolite. Diorite and metagabbro have been reported from the Combe Lake area and the Beckett Lake area respectively. In both cases, however, localities fall outside the Wollaston Lake Belt itself, in what is considered here to be, in part at least, remobilized Archean material.

Recrystallization and deformation have affected basic rocks of the Needle Falls, Beckett Lake and Daly Lake areas. Cross-cutting relationships show that at least some of these rocks are younger than sedimentary rocks of the belt. They intrude the Sandfly Lake Group (Needle Falls, possibly Beckett Lake), the Meyers Lake Group (Needle Falls) and the Daly Lake Group (Daly Lake). In the Needle Falls area, granitized epidiorite is found, indicating that emplacement of the diorite preceded the granitization. A similar age might tentatively be assigned to most basic intrusive rocks of the Wollaston Lake Belt. Their emplacement would post-date most or all of the sedimentation in the belt, but pre-date metamorphism and deformation. In this hypothesis, they would be assigned a late Aphebian age. In the Daly Lake area, essentially undeformed and unmetamorphosed clinopyroxene-plagioclase pegmatites are younger than all other known basic intrusive rocks.

SUMMARY OF PRE-OROGENIC HISTORY OF THE BELT

The oldest recognized rocks in the vicinity of the belt are the probably supracrustal rocks (amphibolites) that occur as inclusions in the probably Archean Pederson Lake complex of the Daly Lake area. Other possibly Archean supracrustal rocks include amphibolites and quartzo-feldspathic rocks of the Geikie River and nearby areas. These rocks have an easterly (Archean?) trend and are probably predominantly of volcanic origin. They occur adjacent to the Sandfly Lake Group or probably correlative rocks and are lithologically similar to this group except that they may lack arkosic rocks. They may be correlative with the Proterozoic Sandfly Lake Group or they may be Archean.

The Archean and probably Archean granitic rocks are widespread in and near the belt (Fig. 2) and it is postulated that they underlaid most or all of the area of deposition of the Proterozoic supracrustal rocks. Burwash and Krupička (1969) have presented evidence of polycrystalline deformation in the Precambrian basement of Western Canada and have postulated a zone (the Athabasca mobile zone) that consists in large part of pre-Hudsonian crystalline rocks. If the boundaries of this zone are projected onto the exposed part of the Shield the Wollaston Lake Belt lies within it. The writers consider that the "pre-Hudsonian crystalline rocks" of Burwash and Krupička are at least in part equivalent to the Archean and probably Archean granitic rocks of this paper.

The probably Proterozoic supracrustal rocks deposited on and above the sialic basement are divided into the Sandfly Lake, Meyers Lake, and Daly Lake Groups and probably correlative rocks. An atypical assemblage that is now considered to belong to the Daly Lake Group is referred to as the Hidden Bay assemblage. The Sandfly Lake Group is probably the oldest group. It is overlain, probably unconformably, by the Meyers Lake Group which is probably overlain conformably by the Daly Lake Group.

The Sandfly Lake Group is at least 1,500 feet thick. An original maximum thickness of 5,000 to 10,000 feet appears reasonable. The group and probable correlatives consist mainly of rocks probably derived from basic to acidic volcanics, arkosic wacke, and greywacke. Crossbedding and volcanic pillows are preserved locally and at least one local unconformity has been recognized. Deposition probably took place rapidly, on a sialic basement, in part, at least, in a shallow-water marine environment.

The Meyers Lake Group has an estimated thickness of about 1,500 to 3,000 feet. The group and probably correlative rocks consist mainly of quartzite, oligomictic quartz-pebble meta-conglomerate, and pelitic schist. The assemblage bears a considerable resemblance to assemblages characteristic of a stable shelf but the presence of feldspathic arenite and minor arkosic wacke indicates a less stable environment than the typical shelf. Crossbedding is preserved locally. Shallow-water marine deposition is postulated.

The Daly Lake Group has a thickness of at least 5,000 feet and a more realistic estimate is probably 10,000 feet or more. A meta-sedimentary assemblage consisting mainly of rocks probably derived from shale, greywacke, and arkosic arenite forms most of the group and of the probably correlative rocks. Rocks derived from limestone and feldspathic quartzite (Hidden Bay assemblage) are probably part of the group. They occur locally near the western margin of the Wollaston Lake Belt. Crossbedding and graded bedding are preserved locally within rocks belonging to the group. Deposition of the group is postulated to have taken place, at least in part in a shallow-water marine environment, unconformably on a sialic basement to the west and conformably on rocks belonging to the Meyers Lake Group to the east.

If the rocks assigned to all three groups were deposited in the order postulated during the depositional phase preceding the Hudsonian orogeny, then a period of time must have occurred during which the area was comparatively stable. Rocks of the Meyers Lake Group, deposited during this time, separate rocks probably deposited under conditions of great crustal instability

(Sandfly Lake Group) from rocks deposited under conditions of moderate crustal instability (Daly Lake Group). The probable unconformity between the Meyers Lake and Sandfly Lake Groups indicates an erosional and perhaps deformational interval separating rocks belonging to what may be two depositional cycles (i.e. Sandfly Lake and Meyers Lake-Daly Lake).

Within the belt, no pre-orogenic supracrustal rocks are known which seem to be younger than the Daly Lake Group. Basic to granitic plutonic bodies occur in the belt. They may be in part pre-orogenic but in general are probably younger.

REGIONAL CORRELATION

The supracrustal rocks of the Wollaston Lake Belt are generally separated from other belts of supracrustal rocks by granitic areas and hence correlation is difficult. The problem is compounded by the fact that, with the exception of the Hurwitz Group, other supracrustal rocks of this general area have not yet been assigned to any groups.

The Hurwitz Group underlies a part of the "Southeastern Barren Grounds" (Wright, 1967). Probably correlative rocks occur in the Phelps Lake area of northeastern Saskatchewan (Tremblay, 1960, unit 1) and in the Kasmere Lake area of northwestern Manitoba (Fraser, 1962, units 1, 2, and 3). All known occurrences are north, northwest or northeast of the Wollaston Lake Belt. The Hurwitz Group is described in this publication by R.T. Bell. The various formations comprising this group show little lithological resemblance to any of the assemblages of supracrustal rocks within the Wollaston Lake Belt, with the possible exception of the Hidden Bay assemblage. The Hidden Bay assemblage could represent a sedimentary facies transitional between rocks in the Phelps Lake and Kasmere Lake areas (which may be correlative with the Hurwitz Group) and rocks typical of the Daly Lake Group. Hence these groups might be broadly contemporaneous. The Hidden Bay assemblage occurs on or near the western margin of the Wollaston Lake Belt, i.e. spatially between rocks probably correlative with the Hurwitz Group and rocks typical of the Daly Lake Group.

In view of our present lack of knowledge of the supracrustal rocks within and outside of the Wollaston Lake Belt north of latitude 58°N no other correlations can be attempted in this area. An attempt at correlation with supracrustal rocks south of this latitude and west of the belt is considered to be futile because of the occurrence of predominantly granitic terrane immediately west of the belt and because no adequate description of the supracrustal rocks west of the granitic terrane is available. To the east of the belt there is a terrane of young granites and migmatites and east of this is a belt consisting in large part of supracrustal rocks. This belt may be informally termed the La Ronge Belt. It may be traceable for about 125 miles from Lac La Ronge to and beyond the southern end of Reindeer Lake. It is beyond the scope of this paper to formally define this belt or to subdivide its rocks into groups. However, it should be noted that rhyolitic to basaltic meta-volcanic rocks are abundant and at least locally these are spatially associated with arkosic meta-sedimentary rocks (e.g. Padgham, 1960) forming an assemblage with some lithological similarity to the Sandfly Lake Group. Elsewhere in this belt occurs a meta-sedimentary assemblage that consists mainly of biotite-rich rocks with minor calc-silicate rocks and meta-arkose (e.g. Mawdsley and Grout, 1951). This has some lithological similarity to the Daly Lake Group. No correlations are considered to be warranted until further work has been done, but the possibility exists that some rocks of the La Ronge Belt may be equivalent to those of the Wollaston Lake Belt.

RECOMMENDATIONS FOR FURTHER WORK

The Wollaston Lake Belt is a major feature of the western part of the Churchill structural province and, as such, warrants thorough investigation. As this paper illustrates, our present knowledge of the belt is minimal. From a stratigraphic viewpoint it is suggested that priority in future work should be given to radiometric dating, chemical analyses, and detailed mapping.

Rb-Sr whole rock dating and U-Pb dating of zircons should be undertaken. It is particularly important to date by these methods the volcanic rocks and the granitic rocks of probable or possible Archean age. Dating of the volcanic rocks could establish the age of the Sandfly Lake Group and correlative rocks and perhaps of the various easterly-trending rocks east of the belt, in and near the Geikie River area (*see* Table II). Sampling for dating of the Sandfly Lake Group and probably correlative rocks should be carried out in and near the Janice Lake area where the metamorphic grade is comparatively low (lower amphibolite facies) and where acidic meta-volcanic (?) rocks are present. Dating of Archean (?) granitic rocks is essential to establish the extent of the Archean basement. This is necessary to understand the environment of sedimentation of the supracrustal rocks.

Chemical analyses of the supracrustal rocks are needed to determine the origin of the meta-sedimentary rocks and to classify and determine variations within the meta-volcanic sequence. Analyses, by rapid chemical methods, of a limited number of samples of the various biotite gneisses and granofels are recommended. Numerous analyses of the meta-volcanic rocks should be carried out as an aid to their classification.

Selected areas should be mapped at the scale of one inch equals one mile or better. Priority should be given to areas along and extending from the western margin of the belt to the northeast, to the area underlain by map-unit (1) in the Phelps Lake area (Tremblay, 1960), and to areas near the eastern margin of the belt. The western areas should be mapped to determine the extent of the Hidden Bay assemblage and the relationships of the Daly Lake Group and probably correlative rocks to Tremblay's (1960) map-unit (1) (= Hurwitz Group (?)). If the Hidden Bay assemblage is extensively developed parallel to the western margin of the belt it would suggest that there was a comparatively stable area to the west and that the present outlines of the belt are not an accident of folding but are parallel to the margins of the original depositional trough. Mapping near the eastern margin should be concentrated in areas where rocks belonging to the Sandfly Lake Group may be in contact with rocks belonging to the Daly Lake Group (*see* Fig. 2). Two areas appear to be especially promising for determining the relationships of these groups. These are immediately south of the Daly Lake area and east and northeast of the Middle Foster Lake area (including the Janice Lake area).

Valuable geological information about the belt might also result from "lead-lead" radiometric dating of syngenetic (?) sulphide mineralization (Pyke and Partridge, 1967; Rath and Morton, 1969; Møller, 1969) and U-Pb dating of uraninite, pitchblende and other minerals that occur in pegmatites and in vein-type deposits within the belt.

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Comments by: R.A. Burwash,
Department of Geology, University of Alberta, Edmonton

For more than ten years I have been searching for relict Kenoran material within the Hudsonian orogenic belt. In the light of this search, three comments seem pertinent to the paper given by Dr. Money. Two new age determinations support his thesis, newly calculated regional average chemical data suggest a note of caution, and a field project completed in 1969 provides some contradictory evidence.

Zircon U/Pb Dating

In 1968, bulk samples of the igneous and metamorphic rock units suspected to be of Archean age were collected by Peter Money from the Needle Falls and Daly Lake areas. These samples were sent to the University of Alberta, supplying the material for six zircon concentrates prepared between July and December, 1968. Uranium-lead dating has been carried out by Dr. H. Baadsgaard. Unfortunately, to date only two of these samples have been analyzed for lead isotopes, and only one for uranium isotopes.

The completed sample is from the Roper Bay quartz monzonite, near the south boundary of the Daly Lake map-area (56°30'30"N, 105°37'04"W). Zircons from this sample give a Pb^{207}/Pb^{206} ratio of radiogenic lead of 0.1521, for an age of 2,405 million years. A sample from the northwestern corner of the Daly Lake map-area has a slightly lower Pb^{207}/Pb^{206} ratio, but after determination of uranium will probably yield a similar or slightly younger age.

Four zircon separates prepared from three bulk samples from the Needle Falls area have, as yet, not been analysed. We hope to complete this work shortly.

Regional K and Rb Metasomatism

Evidence of K-metasomatism has recently been suggested for a large area of the subsurface Precambrian underlying the western Canada sedimentary basin (Burwash and Krupička, 1969). The cataclastically deformed and metasomatized area, the "Athabasca mobile zone" is mainly a southwestern extension of the lithologic assemblages exposed north of Lake Athabasca. In addition to K_2O, a number of other components have been mobile. In the acid plutonic rocks (SiO_2 > 65%) the undeformed rocks give a regional average for the Rb/Sr ratio of 0.61. For the similar rocks deformed and metasomatized the ratio increases more than 2.5 times to 1.60. In view of this great change in Rb/Sr ratios the whole philosophy of "closed system" polymetamorphism is questioned, and the use of Nicolaysen-type Rb/Sr plots may not be valid.

Poly- vs Monometamorphism in "Western Granitic Rocks"

After detailed textural studies of the subsurface Precambrian of western Canada during the winter of 1968-1969, a field project along the Churchill River was undertaken in June 1969. The traverse from Needle Falls to Knee Lake crossed the Wollaston Lake Belt and some 50 miles of "Western Granitic Rocks". Careful field studies of textural and structural relations left both Dr. Jirí Krupička and me convinced that we were observing a migmatitic gneissic complex derived directly from a sedimentary-volcanic sequence. This does not argue directly against a relic Kenoran metamorphic complex in the Daly Lake area. The granulite facies rocks in the Daly Lake area may represent a culmination in the Wollaston Lake Belt where it crosses a pre-existing east-west Kenoran structure.

Response by: P.L. Money,
Texas Gulf Sulphur, Toronto

Dr. Burwash has commented on three points, zircon U/Pb dating, regional K and Rb metasomatism, and poly- vs. monometamorphism in the "western granitic rocks". The U/Pb dating extends the area of known Archean rocks (Roper Bay quartz monzonite, shown as of uncertain age on Fig. 2). It also confirms the Archean age of the Pederson Lake complex which had been inferred on indirect evidence, as the sample from the northwest corner of the Daly Lake area is from this complex. The problem of interpretation of Rb/Sr "ages" in view of possible Rb-metasomatism requires much further work. Nevertheless, it seems probable that the 2,470 ± 65 m.y. Rb/Sr age reported is at least a reliable minimum age for the plutonic rocks in question, as there appears to be no mechanism whereby addition of Rb (in conjunction with the Hudsonian orogeny?) could result in an increase in apparent age. If Rb were selectively removed, an increase in apparent age might be possible, but this does not appear to be the case. Drs. Burwash and Krupička have studied the "western granitic rocks" of the Needle Falls area in a traverse which extended some 50 miles to the west of Needle Falls. They interpret the rocks that they saw as a migmatitic gneissic complex derived directly from a sedimentary-volcanic sequence. If their interpretation is correct, the parent sedimentary-volcanic sequence could be Aphebian and the Rb-Sr age (Money, 1968) might simply indicate the presence of Archean detritus. A supporting argument used in the interpretation is that the "western granitic rocks" lack evidence of polymetamorphism, unlike rocks that Burwash and Krupička would interpret as "older basement" elsewhere. Money and co-workers do not consider that the "western granitic rocks", the southernmost "Archean or probably Archean granitic rocks" of Figure 2, extend fifty miles to the west. Their known width is only about ten miles. A rapid reconnaissance survey by one of the writers (Money) indicated that some of the "granitic" rocks farther west are metamorphosed arkosic rocks like those belonging to the Daly Lake Group. It thus appears that at least in part we are not considering the same rocks. The "western granitic rocks", as defined by the writers, are hardly a "migmatitic gneissic complex". A number of areas of migmatite occur, for example near Sandfly Lake, but outside of these fairly well defined areas, the "western granitic rocks" are remarkably monotonous and uniform. The composition of two typical samples (Money, 1967) falls in the low temperature trough of the system SiO_2-$NaAlSi_3O_8$-

$KAlSi_3O_8$ and it is considered that the "western granitic rocks" are magmatic in origin. It is possible that these rocks formed during the Hudsonian orogeny by fusion of Archean supracrustal rocks or that they are Archean plutonic rocks thoroughly remobilized during the Hudsonian orogeny. Either concept is compatible with the field relationship, the Rb/Sr data and the lack of evidence of polymetamorphism. Hudsonian mobilization of Aphebian supracrustal rocks consisting in large part of Archean detritus could also conceivably fit the Rb/Sr and other data. The presence of boulders of granitic rocks in the Sandfly Lake Group (Money, 1967) which are chemically like the "western granitic rocks" (both are characterized by unusually high Na_2O:CaO ratios for their bulk compositions) and may be derived from them suggests that the "western granitic rocks" may have existed as magmatic rocks before deposition of this group and, if so, were probably initially emplaced in the Archean.

Comment by: A. Ruffman

Both you (P.L. Money) and R.T. Bell earlier used the term "stable craton" and "metastable craton". Surely the term stable craton is redundant and we should use only "craton" and find a different term for "metastable craton" to avoid confusion.

Response by: P.L. Money

This question should be directed to R.T. Bell as to the best of my knowledge he originated the term "metastable craton". The term, in my opinion, is justifiable and is valuable in that a "metastable craton" is a useful concept. According to the AGI Glossary of Geology and Related Sciences, craton (Kraton) was first used by Stille and refers to a "relatively immobile part of the earth". Presumably it is immobile relative to a geosynclinal belt. The areas discussed by Bell and by Money, Baer, Scott, and Wallis at this workshop appear to have been generally immobile compared to a geosyncline and hence can be considered to be cratonic areas. A distinction is required between normal (?) cratonic areas that have undergone relatively mild deformation or only epeirogenic movement and those abnormal (?) areas that have undergone considerable subsequent deformation. Such a distinction may be made by referring to the former as "stable" and the latter as "metastable".

Comment by: P.L. Money on paper by R.T. Bell

This paper is a stimulating and valuable contribution. I do not consider that I am competent to discuss it in detail. Perhaps, however, I could suggest that while the Hurwitz Group may well be a prototype for deposition on a metastable craton, there are metastable cratons and metastable cratons. The Hurwitz Group has been deposited on an exceptionally (?) stable part of the metastable Archean craton, in contrast to the various groups within the Wollaston Lake Belt, which have been deposited on a generally less stable part, but certainly not under typical ensialic orthogeosynclinal conditions. Various criteria for relative stability support this view. These include the maturity and general nature of the supracrustal rocks, the degree and type of deformation that the supracrustal rocks underwent, and perhaps their metamorphic grade, as these are at least in part a function of the stability of the cratonic platform.

It might be possible, using these and perhaps other criteria, to set up a "cratonic stability series" in analogy to Read's "granite series". Such a series might make it possible, through a study of Aphebian supracrustal rocks, to delineate or crudely "contour" zones, belts, or areas of the Archean craton on the basis of their relative stabilities, thus providing some insight into the nature of the metastable part of this craton.

Response by: R.T. Bell to A. Ruffman and P.L. Money

The adjective "metastable" has been applied to craton, just as it is to "equilibrium". Possibly "stable craton" is redundant, but to create a new term for "metastable craton" could complicate the matter and could be considered presumptuous. I think that the White Rock Lake Member of the Kinga Formation demonstrates a platform-stable shelf situation. The succeeding pattern of sedimentation and of deformation demonstrates that this stable platform or craton was not fully stable but only "metastable".

A "cratonic stability series" has much merit. Geologic synthesis has traditionally focused on Phanerozoic mountain belts. Of late, our attention has been dramatically focussed on ocean basins and has born fruit in leading to a widely accepted dynamic model of the earth's crust. The time is ripe now to look at the stable cores of the continents and to assess the process of craton formation. An earlier attempt (with accretion) ran unfortunately into some dead ends.

Two events would appear to stand out in the earth's history: one is the Archean and its terminal "orogenies" and the other is the terminal Aphebian "orogeny". Subsequent events, I venture, are one order of magnitude smaller. Possibly sea-floor spreading (and attendant subduction zones, etc.) was only a minor thing prior to the Hudsonian event. If for example Hudsonian areas are allowed to be unfolded (by a 3 to 4 times extension) oceanic areas would be almost completely eliminated.

Archean events are possibly governed by changes from a homogeneous earth to a layered earth, coupled with higher radiogenic heat sources. The Hudsonian event is possibly the reflection of the first grand sweep of seafloor spreading, accompanying the beginning of high-level convection in a plastic mantle.

THE CORONATION GEOSYNCLINE OF APHEBIAN AGE, DISTRICT OF MACKENZIE

P.F. Hoffman, J.A. Fraser and J.C. McGlynn
Geological Survey of Canada, Ottawa, Ontario.

Abstract

There are four belts of Aphebian sedimentary and volcanic rocks (Great Slave, Epworth, Goulburn and Snare) in the District of Mackenzie. These belts are remnants of an arcuate, northerly trending orthogeosyncline convex to the west, which developed probably between 2,000 and 1,750 million years ago, between a craton to the east and a late Aphebian orogenic belt (the Bear Province) to the west. The orthogeosyncline is named the "Coronation Geosyncline", after its area of most complete preservation south of Coronation Gulf.

Stratigraphic thicknesses of 32,000 feet are attained in the geosyncline, in contrast to the 6,000 to 10,000 feet recorded in the contiguous strata over the craton. Four phases in the depositional history of the geosyncline are:

1. a pre-orogenic orthoquartzite-carbonate phase,
2. a transitional euxinic-volcanic phase,
3. an early syn-orogenic flysch phase,
4. a late syn-orogenic molasse phase.

The orthoquartzitic sandstones of the pre-orogenic phase are derived from the craton, whereas the syn-orogenic feldspathic greywacke and lithic sandstone are derived from the orogenic belt.

The Coronation strata are intensely deformed and metamorphosed only in the orogenic belt, but they have been displaced by many westerly dipping low angle thrust-faults adjacent to that belt.

The Coronation Geosyncline is, in general, similar to the classic Phanerozoic orthogeosynclines of North America.

INTRODUCTION

Late Aphebian sedimentary and volcanic rocks are preserved in four belts in the District of Mackenzie (Fig. 1).

1. The Great Slave Supergroup in the East Arm of Great Slave Lake.
2. The Snare Group between Great Slave and Great Bear Lakes.
3. The Epworth Group south of Coronation Gulf between the Coppermine and Tree Rivers.
4. The Goulburn Group along the Burnside and Western Rivers southwest of Bathurst Inlet, Coronation Gulf.

Study of the stratigraphy, paleocurrents and sedimentology of rocks of the Great Slave Lake area suggested the presence of a north-northwesterly trending orthogeosyncline (Hoffman, 1969). Predictions were made concerning the nature of the other three belts which, if verified, would support both the orthogeosynclinal hypothesis and the supposition that all four belts are remnants of the same succession.

Comparison of facies in the four belts is now sufficiently advanced to confirm the orthogeosynclinal hypothesis. The comparisons are based on the work of Fraser (1964), Bostock (1965), Tremblay (1968), Fraser and Tremblay

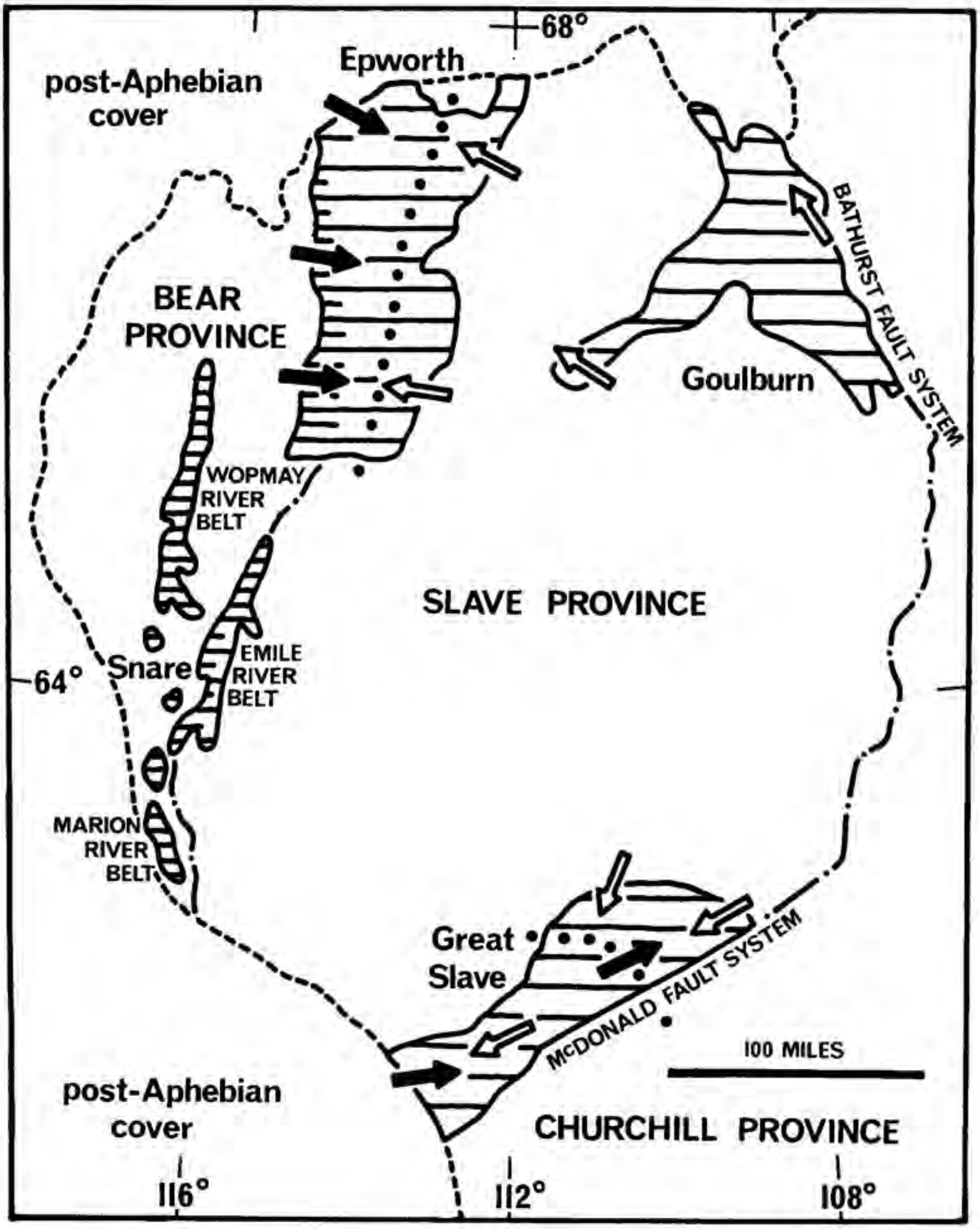

Figure 1. Northwestern part of the Canadian Shield. The four areas of Aphebian sedimentary and volcanic rocks are shown in the ruled pattern; widely ruled:unmetamorphosed; closely ruled:metamorphosed and penetratively deformed. The open arrows show paleocurrent directions in the pre-orogenic craton-derived sandstones, the solid arrows in the syn-orogenic flysch and molasse. The heavy dotted line is the approximate position of the hinge-line of the geosyncline.

(1969) in the Goulburn Group; Fraser (1960, 1966, in press), Fraser and Tremblay (1969), Hoffman (1970) in the Epworth Group; McGlynn (1957, 1964), Ross (1959, 1966), Ross and McGlynn (1965), McGlynn and Ross (1962, 1963) in the Snare Group; Stockwell (1932, 1936), Barnes (1953), Hoffman (1968, 1969) in the Great Slave Supergroup.

The orthogeosyncline is here named the "Coronation Geosyncline" from the region south of Coronation Gulf where it is most completely exposed. It has an arcuate northerly trend, is convex to the west and it developed between a craton (platform) to the east and a late Aphebian orogenic (mobile) belt to the west.

PRE-CORONATION BASEMENT

The basement beneath the Coronation Geosyncline consists mainly of Archean rocks covered only locally by older Aphebian strata. The Archean, exposed throughout most of the Slave Province, includes the variously metamorphosed volcanic and sedimentary rocks of the Yellowknife Supergroup (*see* McGlynn and Henderson, this publication) in northerly trending belts separated by granitic rocks which yield late Archean radiometric ages.

Pre-Coronation Aphebian strata, known only in the Great Slave belt, include the tightly folded Wilson Island Group, once believed to be Archean but now thought to overlie the late Archean granites (Reinhardt, 1969), and the much less deformed Union Island Group. The Wilson Island Group consists of basic to acid volcanic flows and tuffs overlain by arkosic conglomerate, orthoquartzite, dolomite and siltstone. The Union Island Group contains dolomite overlain by graphitic shale, basic pillow lava and flow breccia, and quartz-pebble conglomerate. Both sequences differ conspicuously from the overlying Coronation strata in their lack of stromatolites and red beds. As the two groups total at least 20,000 feet, the region must have had a long Aphebian depositional history, even before the beginning of the Coronation Geosyncline. The folding in rocks of the Wilson Island Group, much more intense than in the younger Aphebian strata, indicates a major mid-Aphebian structural event.

DEFORMATION OF THE CORONATION ROCKS

Rocks of the Coronation Geosyncline are intruded by late Aphebian granitic rocks and have suffered intense penetrative deformation in the Bear Province in the west. Here, the Snare and Epworth argillaceous sediments are converted to hornfelses and schists containing sillimanite, garnet, andalusite or cordierite. Also prominent are late Aphebian stocks, sheets, dykes and ignimbrite flows of vari-coloured acidic quartz-feldspar porphyry. The Bear Province, and its presumed northern and southern extensions beneath the post-Aphebian cover, is the orogenic belt of the geosyncline.

To the east of the Bear Province, deformation of the Coronation rocks becomes progressively less intense. The Epworth belt contains a 45-mile-wide zone of westerly dipping low angle thrust-sheets. Telescoping of the Epworth rocks accounts in part for the abruptness of stratigraphic thickening across the thrust zone, many of the Epworth sediments having been originally deposited perhaps tens of miles west of their present position. Along the eastern margin of the Epworth belt and in the Great Slave and Goulburn belts, there are no thrust faults. Here, the Coronation strata are only gently folded, with relatively sharp anticlines and broad synclines in part, around north and northeasterly trending axes. Dips are mostly less than thirty degrees and the strata are intruded only by dioritic laccoliths (in the Great Slave belt) and diabase dykes and sheets.

Unmetamorphosed Great Slave and Goulburn strata are in fault contact with gneisses of the Churchill Province along the McDonald and Bathurst fault systems. The Aphebian strata are locally tilted to high angles near these faults. Numerous northeasterly trending faults with right-lateral displacement cut the Epworth and Snare rocks and post-date thrusting.

Latest Aphebian or early Helikian continental red bed successions, up to 13,000 feet thick, lie unconformably over the folded Coronation strata and were deposited in grabens contemporaneously with faulting in the Great Slave Lake (Et-then Group) and Great Bear Lake (Cameron Bay Group) regions.

DEPOSITIONAL HISTORY OF THE CORONATION GEOSYNCLINE

The Coronation rocks in the Great Slave Lake and Coronation Gulf regions are compared in two east-west stratigraphic cross-sections (Fig. 2). Relatively thin strata cover the craton in the east and pass westward across a "hinge-line" into greatly thickened geosynclinal successions. Although they are three hundred miles apart along the trend of the geosyncline, both areas have a similar succession of facies, reflecting a similar depositional history.

In broad terms, the depositional history of the geosyncline can be broken down into four phases.

1. A pre-orogenic orthoquartzite-carbonate phase.
2. A transitional euxinic-volcanic phase.
3. An early syn-orogenic flysch phase.
4. A late syn-orogenic molasse phase.

The continental fault-basin sediments overlying the folded Coronation rocks may be termed a fifth, post-orogenic, phase. This phase is not covered in this paper.

Pre-orogenic Phase

The pre-orogenic phase consists of a basal terrigenous sequence (Odjick and Hornby Channel Formations), a cyclical dolomite-shale assemblage (Rocknest and Duhamel Formations), and a second terrigenous sequence (Burnside River and Kluziai Formations) restricted mainly to the craton but extending into the geosyncline in the Great Slave Lake region. Paleocurrents demonstrate that the terrigenous sediments are derived from the craton to the east.

The basal terrigenous sequence is the most heterogeneous and laterally variable stratigraphic unit in the geosyncline. It is dominated, in the Great Slave Lake region (Hornby Channel Formation), by crossbedded subarkosic sandstone and conglomerate of alluvial origin, but to the north toward

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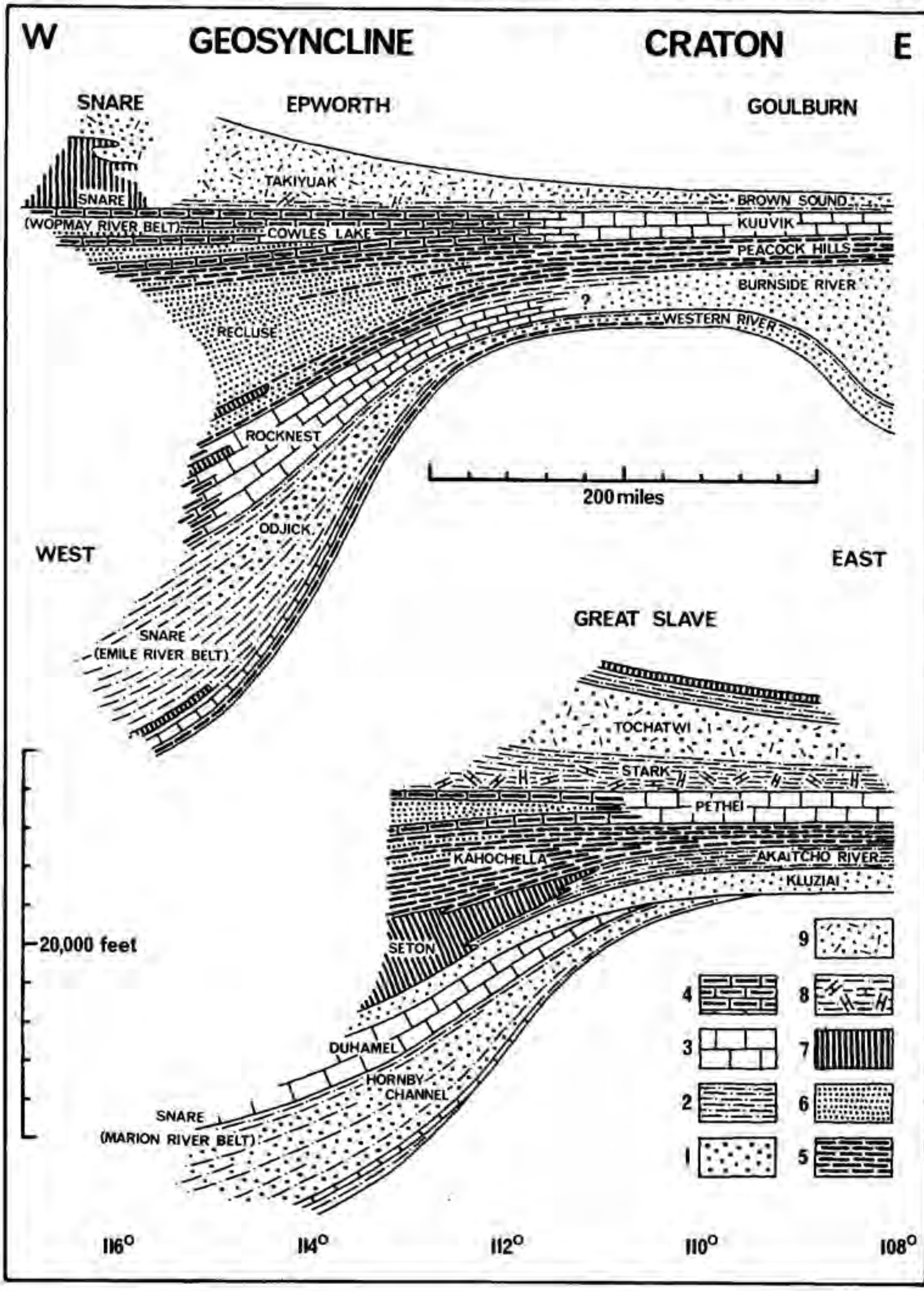
Figure 2 (opposite page).

East-west stratigraphic cross-sections through the Coronation Geosyncline and its contiguous cratonic cover in the Coronation Gulf region (top) and in the Great Slave Lake region (bottom).

Lithologic symbols:

1. mature crossbedded orthoquartzite to subarkose with minor conglomerate;
2. interbedded siltstone, shale and sandstone;
3. stromatolitic and oolitic dolomite, limestone and argillaceous carbonate;
4. non-stromatolitic interlaminated limestone and shale;
5. shale, concretionary shale and minor siltstone;
6. graded greywacke beds rhythmically interbedded with shale;
7. volcanic flows, breccias and tuffs;
8. carbonate mega-breccia with red mudstone matrix;
9. immature crossbedded red lithic sandstone.

the Coronation Gulf (Odjick Formation), vari-coloured siltstone and orthoquartzitic sandstone of mixed alluvial, coastal marine and aeolian origin are increasingly abundant. Stromatolitic dolomite is a minor, but persistent, component, particularly near the base. On the craton (Western River Formation), the sediments are similar but thin, whereas in the extreme west (Snare Group),



there is a thick monotonous succession of thinly interbedded dark siltstone and shale, probably of offshore marine origin, with scattered volcanic units.

The basal terrigenous sediments grade upward into a thick succession of stromatolitic and oolitic dolomite, cyclically interbedded with mud-cracked calcareous shale. In the Great Slave Lake region (Duhamel Formation), thin units of crossbedded white orthoquartzitic sandstone also occur. The cyclical succession was deposited in lagoons and on adjacent tidal flats which covered a vast shallow marine shelf. Along the western margin of the Epworth belt (Rocknest Formation), the mud-cracked calcareous shales largely disappear in a shelf-margin sequence of crossbedded dolomite and thick stromatolite units in which cycles are lacking. Farther to the west, the carbonates pass abruptly into dark off-shelf marine shales with beds of carbonate mega-breccia, possibly of submarine slump origin. Abrupt deepening of water westward of the shelf margin probably explains the absence of thick carbonates in the Snare Group in the extreme west.

The second terrigenous sequence is much more uniform than the first, being composed of crossbedded pink orthoquartzitic to subarkosic sandstone and minor quartz-pebble conglomerate, almost entirely of alluvial origin. Sandstones in the Great Slave Lake region (Kluziai Formation) cannot be distinguished in hand specimen from those of areas south of Coronation Gulf (Burnside River Formation). At the top, the uniform pink alluvial sandstones grade upward into mixed red and white sandstone, red and green siltstone and shale of deltaic origin (Akaitcho River Formation).

The pre-orogenic sandstones are abnormally thick in two areas. Along the southeastern margin of the Great Slave belt, sandstone in the basal terrigenous sequence reaches 5,000 feet in thickness; along the northeastern margin of the Goulburn belt, the second terrigenous sequence is as much as 9,000 feet thick. It is significant that these stratigraphic depressions (i.e. isopach maxima) coincide with the post-Coronation McDonald and Bathurst grabens.

Transitional Phase

In the Coronation Gulf region, the shallow shelf Rocknest carbonates are overlain conformably by a few hundred feet of laminated black pyritic shale (basal Recluse Formation). The transition is abrupt, the shale being separated from the carbonate by thirty feet or less of thin-bedded ripple-marked green siltstone. The black shales are interpreted as having been deposited slowly in a deep euxinic marine basin "starved" of sediment. The rapid subsidence which produced the euxinic basin marks the important transition between carbonate shelf sedimentation below and greywacke flysch sedimentation above.

In the Great Slave Lake region, no deep euxinic basin was established. Early in the transitional phase, subsidence was less rapid than to the north and alluvial sandstone from the craton (Kluziai Formation) extends westward across the hinge-line, over the shelf carbonates (Duhamel Formation). More importantly, subsidence was balanced by the extrusion of a thick complex of volcanic rocks (Seton Formation). These include massive and columnar andesitic flows and flow breccias, minor rhyolitic flows, and great volumes of predominantly basic lithic and vitric tuff. Several breccia-filled vents and associated cinder cones can be seen. The tuffs commonly have been reworked subaqueously, have mud-cracks, gypsum casts, paleosols and are interstratified with non-siliceous granular hematite ironstone and tuffaceous shale. Pillow lavas are absent and the centers of volcanism are thought to have formed islands on an otherwise shallow marine shelf.

It is possible that the 5,000 feet or more of volcanic rocks and associated sediments in the Great Slave Lake region are the time equivalents of less than 300 feet of euxinic black shale in the Coronation Gulf region to the north.

Flysch Phase

The deep basin, formed in the Coronation Gulf region during the transitional euxinic phase, became the site of accumulation of thousands of feet of rhythmically interbedded coarse grained feldspathic greywacke and dark green to black shale (Recluse Formation). The greywacke, extremely immature both texturally and compositionally, forms laterally continuous graded beds up to fifteen feet thick. The greywacke thins to the east and its feather edge does not extend past the hinge-line. The contiguous strata on the craton (Peacock Hills Formation) are red and green shales with ripple marks, convolute bedding and thin units of siltstone, dolomite and, along the eastern margin of the Epworth belt, granular hematite ironstone. The greywacke is derived from tectonic lands uplifted in the orogenic belt to the west and was deposited on submarine fans by currents which flowed both transverse (west to east) and parallel (north to south) to the trend of the geosyncline. The contiguous shales on the craton were laid down in a shallower epeiric sea or on a shelf, but whether the shaly sediment was derived from the same westerly source as the greywacke or from the craton is not known. Although there are numerous diastems within the shale sequence east of the hinge-line, there is no direct evidence that basement was exposed on the craton.

In the Great Slave Lake region, where no deep euxinic basin had been established, greywacke is rare in the strata correlative with the Recluse Formation. Here, the geosyncline merely contains a thickened succession of red and green concretionary shales (Kahochella Group) lithologically similar to the contiguous strata on the craton. There is, however, a westward increase in the proportion of green, relative to red, shale.

Late in the flysch phase, a shallow water carbonate shelf was established on the craton in both the Great Slave Lake (Pethei Group) and Coronation Gulf (Kuuvik Formation) regions. The outer margin of the carbonate shelf is near the hinge-line, west of which deep-water flysch sedimentation continued, with greywacke and shale coming from the orogenic belt to the west and carbonate mud from the cratonic shelf to the east. At this time, greywacke is more prominent in the Great Slave Lake region (Blanchet Formation) than in the Coronation Gulf region (Cowles Lake Formation). The importance of recognizing lateral facies changes is well illustrated here by the juxtaposition of flysch sediments in the geosyncline with shelf carbonates only a few miles away on the craton.

Molasse Phase

Above the flysch sediments and their cratonic equivalents is a thick, broadly regressive succession of red beds. These are most completely preserved in the Great Slave Lake region where red siltstone at the base (Stark Formation) contains halite casts, thin stromatolitic carbonate beds, carbonate mega-breccias and syn-depositionally folded gravity-slide sheets up to a quarter of a mile across. Less extensive carbonate mega-breccias occur in red siltstone at the base of the succession in the Coronation Gulf region. The siltstones grade upward in both regions into crossbedded red lithic sandstone of alluvial origin (Takiyuak and Tochatwi Formations, samples of which cannot be distinguished). The sandstone is highly immature, being composed in large part of angular sedimentary and volcanic rocks fragments. Dolomite-pebble conglomerate occurs near the base of the sandstone in the Great Slave Lake region, and red mud-cracked siltstone with halite and gypsum casts at the top. Paleocurrents demonstrate that the molasse sediments, like the flysch greywackes before them, are derived from the orogenic belt to the west.

In the Great Slave Lake region, the molasse sediments are capped by a thin sheet of columnar basalt flows. Much thicker massive and columnar basalt is associated with red sandstone in the Wopmay River belt of the Snare

Group in the extreme west. Volcanism in the orogenic belt, contemporaneous with molasse sedimentation, probably provided most of the volcanic fragments in the red molasse sandstone.

Both the flysch and molasse phases of sedimentation are classed as syn-orogenic because they are genetically related to uplift and erosion in the orogenic belt of the geosyncline. The transition from pre-orogenic to syn-orogenic deposition is marked by the change from highly mature to highly immature sandstones, and by a 180 degree reversal in paleocurrent trend.

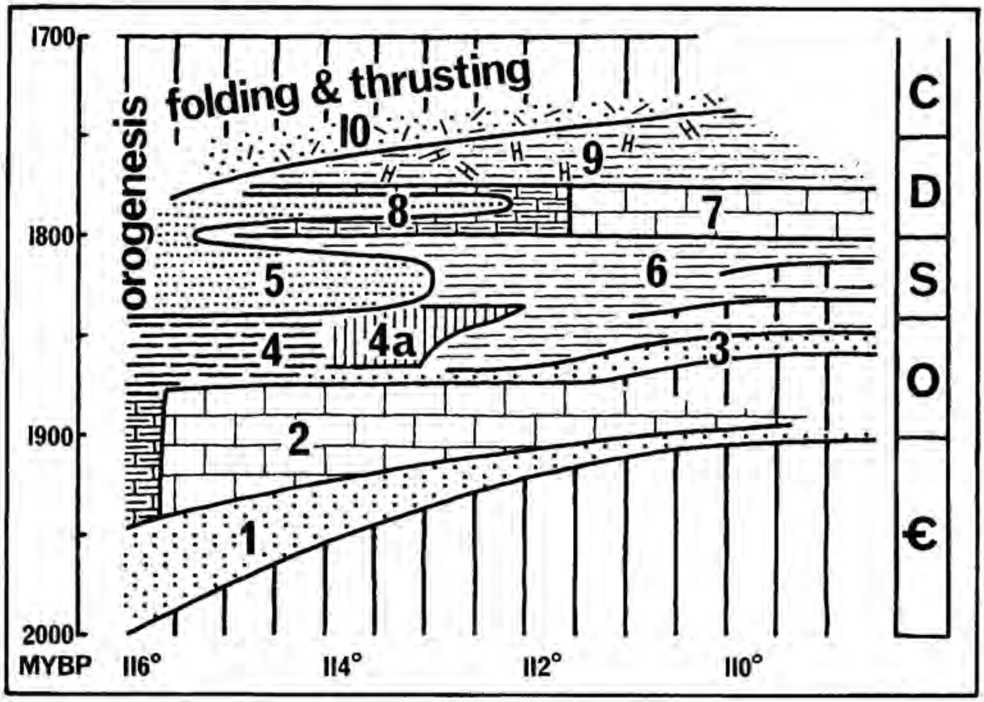
COMPARISON WITH PHANEROZOIC GEOSYNCLINES

The depositional history of the Coronation Geosyncline differs only in minor details from that of the classic Phanerozoic orthogeosynclines of North America (e.g. the Appalachian, Cordilleran, Franklin and Ouachita-Marathon Geosynclines). The sequence of facies is so similar to that outlined by Pettijohn (1957, p. 641) that his terminology has been used in this paper. In the geosynclinal models of Kay (1951, *see* p. 107 for definitions), the pre-orogenic phase would be classed as miogeosynclinal and the syn-orogenic flysch and molasse as exogeosynclinal. Those parts of the pre-orogenic and transitional phases containing volcanic rocks would be termed eugeosynclinal.

Recognizing the presence of an exogeosyncline ("clastic wedge" of King, 1959, p. 57) is most important as this is the sedimentary record of orogenesis in the geosyncline. The molasse is deposited synchronously with the major period of orogenic uplift. The best estimate of the absolute age of orogenic uplift in the Coronation Geosyncline comes from K-Ar dating of granitic and metamorphic rocks within the orogenic belt, dates which probably reflect uplift of the dated rocks above the critical depth (related to the geothermal gradient), below which radiogenic argon diffuses freely. Twenty-two such dates from the Bear Province, the orogenic belt of the geosyncline, average 1,750 million years, with all but two falling within 105 million years of the average. Therefore, 1,750 million years is taken as the best available estimate of the age of molasse sedimentation in the Coronation Geosyncline.

There is no good reason to believe that rates of sedimentation in the Proterozoic should be vastly different from those of similar facies in the Phanerozoic. Therefore, by estimating sedimentation rates in the Coronation Geosyncline from the known time-stratigraphy of the Phanerozoic geosynclines which it so closely resembles, a crude and highly interpretive, but nonetheless instructive, time-stratigraphy of the Coronation succession can be constructed (Fig. 3). This type of cross-section has the advantage over normal rock-stratigraphic ones in that horizontal lines are time horizons, and can therefore be used to reconstruct paleogeographic zonations. A significant conclusion to be drawn from this exercise is that the beginning of deposition in the Coronation Geosyncline may have occurred no earlier than 2,000 million years ago. As the Aphebian era began at least 400 million years before this, the Coronation succession may well be entirely younger than Aphebian strata in other parts of the Shield. It is, for example, very likely younger than the classic Huronian geosyncline, known from dating of intrusive dykes to be older than about 2,150 million years (Van Schmus, 1965), a conclusion previously reached by Roscoe (1969).

The similarity of the Coronation Geosyncline to the orthogeosynclines of Phanerozoic age in North America implies that the tectonic conditions responsible for the unique depositional and structural history of such geosynclines have been present, at least intermittently, since late Aphebian time. Non-actualistic models need not, and should not, be invoked.



Stratigraphic units

10. Takiyuak-Brown Sound-Tochatwi molasse sandstone;
9. Stark red mudstone with carbonate mega-breccia;
8. Cowles Lake-Pethei off-shelf carbonate and flysch;
7. Kuuvik-Pethei carbonate shelf;
6. Peacock Hills-Recluse-Kahochella shale;
5. Recluse flysch;
4. Basal Recluse euxinic basin;
4a Seton volcanic island complex;
3. Burnside River-Kluziai terrigenous shelf;
2. Rocknest-Duhamel carbonate shelf;
1. Odjick-Hornby Channel terrigenous shelf.

Figure 3. Time-stratigraphic model of the depositional history of the Coronation Geosyncline. The cross-section is constructed by taking 1,750 million years as the best estimate of the age of molasse sedimentation and extrapolating the depositional rates of the older units by comparison with similar lithofacies in Phanerozoic orthogeosynclines for which there is time-stratigraphic control. Note that the vertical extent of each stratigraphic unit is the estimate of its duration in time, not its measured thickness. Horizontal lines across the diagram are time-lines. They show that different units were deposited synchronously and that many units are diachronous. On the left is the time-scale in millions of years before the present; on the right is superimposed the Paleozoic time-scale as an aid in comparing the depositional histories of the Coronation and the familiar Paleozoic orthogeosynclines. The heavy vertical lines cover intervals of erosion.

PROBLEMS REQUIRING FURTHER STUDY

There are at least four major problems relating to the Coronation Geosyncline that cannot be resolved with existing data:

1. The geology of the Bear Province must be evaluated in the light of its regional position as the orogenic belt of the geosyncline. Relationships must be established between structural and magmatic events there, and sedimentation and deformation in the Epworth succession to the east.
2. The tectonic significance of the McDonald and Bathurst fault systems and associated folding is not clear. Both appear to be long-lived features which influenced depositional patterns in the Coronation sediments but have trends that diverge from the depositional strike of the geosyncline.
3. The contact zone between the Slave and Churchill Provinces must be re-evaluated as there is no evidence of Hudsonian orogeny in either the depositional history of the deformation of the Coronation sediments adjacent to the Churchill Province boundary.
4. What is the significance of the eastward displacement of the Coronation Geosyncline relative to the younger Cordilleran Geosyncline? Is this to be taken as evidence for westward continental growth, either by accretion of "new" sial or by fusion with a piece of "foreign", formerly independent, continental crust?

The story presented in this paper is both simplistic and naive. Its success will be measured by the energy expended in collecting data to prove that fact.

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Comment by: A. Ruffman,
Dalhousie University, Halifax.
Present address: Bedford Institute, Dartmouth.

If one wishes to interpret your paper in terms of plate tectonics, I believe a reasonable story could be put forward. Thus we would interpret the Bear Province and what you call the Coronation Gulf Geosyncline as the direct result of the interaction (call it collision if you will) between a Slave plate of Archean age and some other plate to the west, the interaction occurring about 2.0 to 1.7 m.y. ago. However there is a serious problem of how to explain the ages of 1.7 b.y. found in the Churchill Province immediately to the east of the Churchill. May I press you and ask you to speculate on what type of event imprinted the K-Ar ages on rocks of the Churchill Province east of the Slave Province?

Response by: P.F. Hoffman

The answer to your question regarding the Bear Province will have to come from the Phanerozoic record -- Do orthogeosynclines similar to the Coronation Geosyncline result from the collision of sialic plates? It is not yet clear to me that orthogeosynclines result from such collisions, nor that they are necessarily spatially related to continent-ocean plate interfaces.

Concerning the Churchill Province, the boundary of "Hudsonian" ages coincides, in part, with the McDonald and Bathurst fault systems. At the time in question, these faults appear to have been normal fault grabens indicating tensional stress resulting, perhaps, from regional epeirogenic uplift. Southeast of the McDonald Fault, there was uplift of at least 15,000 feet, and maybe as much as 35,000 feet, relative to the area northwest of the fault. The uniformly young K-A mica ages on the Churchill Province side of the fault may reflect epeirogenic uplift of basement rocks above the critical depth-temperature value, dependent on thermal gradient, at which radiogenic argon is retained. If such ages reflect regional epeirogenic uplift, then it is not surprising that large areas of the Shield yield similar ages unrelated to the true age of the rocks involved.

I stress again that the Bear-Slave boundary and the Churchill-Slave boundary are different.

HELIKIAN BASINS AND GEOSYNCLINES OF THE NORTHWESTERN CANADIAN SHIELD

J.A. Fraser, Geological Survey of Canada, Ottawa
J.A. Donaldson, Carleton University, Ottawa
W.F. Fahrig and L.P. Tremblay, Geological Survey of Canada, Ottawa

Abstract

The last mountain-building episode for most of the exposed Precambrian of North America was the Hudsonian orogeny. Occurrence of relatively undeformed, unmetamorphosed, predominantly clastic units of Helikian sedimentary rocks in widely separated regions of the Canadian Shield substantiate early transformation to craton of large parts of the Shield.

The rapid accumulation of thick redbed units, commonly associated with volcanic rocks, in basins bounded by active faults, represents either earliest Helikian or late Aphebian events. Increased tectonic stability is recorded by regoliths below blankets of more mature clastic sediments that, in places, overlap the redbed basins. The most extensive of these blanket deposits are Helikian sandstones of the Athabasca Formation, Thelon Formation, and Hornby Bay Group in the northwestern Canadian Shield. Sandstones of these units contain structures indicative of deposition in shallow water, and appear to be mainly fluvial. Each of the sandstones is overlain by stromatolitic carbonate, a relationship that suggests a change to shallow marine deposition.

The Athabasca and Thelon Formations have been locally faulted but have not been significantly folded. Lithologically similar clastic sediments of the Tinney Cove Formation occur in a fault-bound wedge in the Bathurst Inlet region. Lithology, geometry, and paleocurrent patterns show many similarities with the blanket sandstones, suggesting that the Helikian cratonic cover was once much more extensive.

The clastic-carbonate platform sequence thickens to the northwest and contains an increasingly greater amount of interbedded shale, marking the transition from platform to eugeosyncline facies. The hinge-line of the geosyncline in the northern part of the western Shield is in the Great Bear Lake region. Crustal instability in the hinge-zone is suggested by unconformities and lithologic variation in this area. The main axis of the Helikian eugeosyncline is about 200 miles west of this zone.

Absence of Helikian strata along the Precambrian-Phanerozoic contact between Lake Athabasca and Great Bear Lake indicates the presence of a broad northerly trending arch during the Paleohelikian-Cambrian interval. The term 'Slave Arch' is applied to this tectonic feature.

INTRODUCTION

Redbed deposits similar in lithology, geometry, structural setting, and age occupy intracratonic basins (Fig. 1) in widely separated parts of the northwestern Canadian Shield. Most of these sedimentary deposits are intercalated with, or are overlain by, volcanic rocks of alkalic or acidic composition. Resting unconformably on granites, gneisses, and folded Aphebian and Archean strata of the Churchill, Bear, and Slave Provinces, the redbed deposits show many features that signify rapid deposition and derivation from nearby source areas.

Areally more extensive but thinner sedimentary units in some regions overlap the redbed basins. Petrologically distinct from the redbeds, these younger strata are sufficiently similar to suggest that they are erosional remnants of more extensive, in part correlative, stratigraphic units.

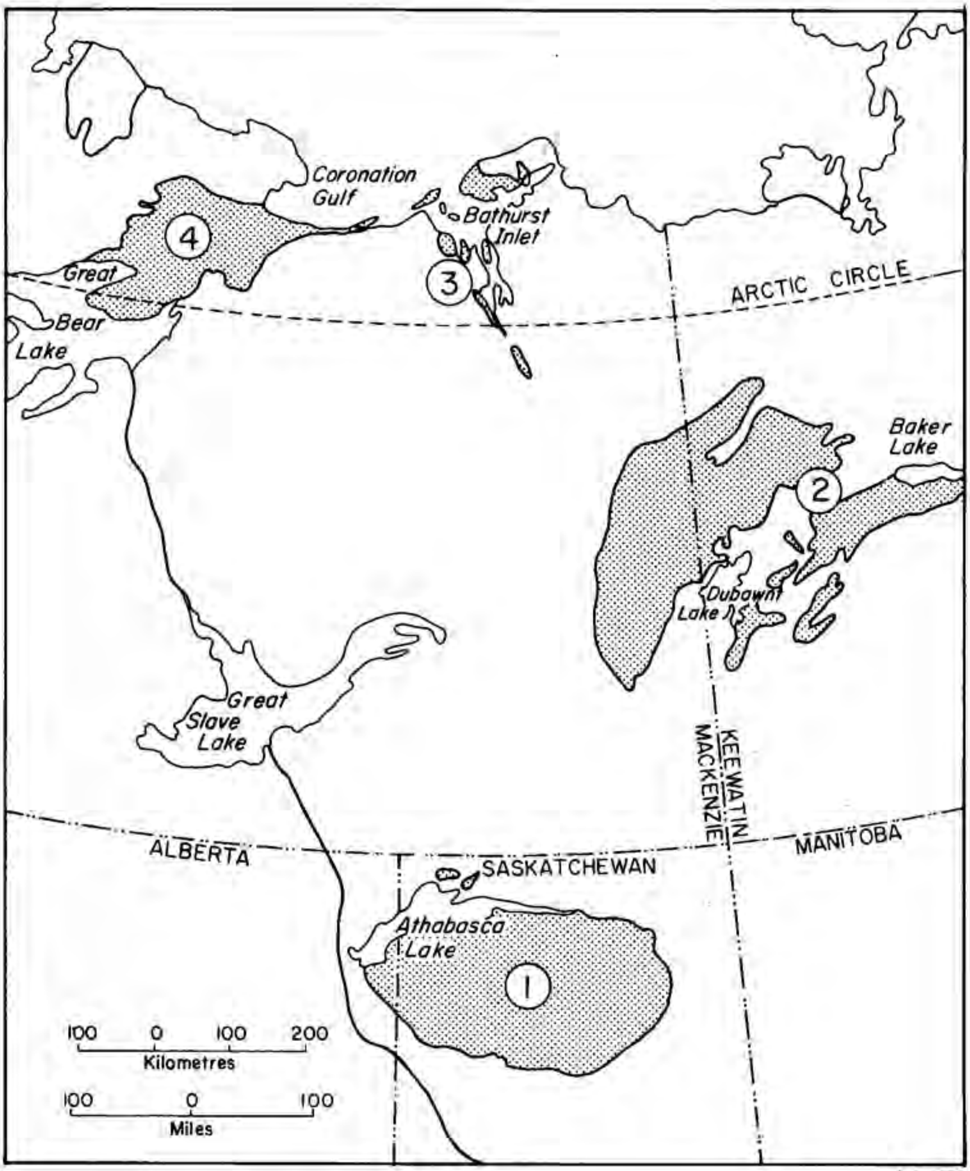

Figure 1. Distribution of Helikian basins in the northwestern Canadian Shield.

The purpose of this paper is to summarize data relevant to the origin and correlation of the redbed deposits and the overlying Helikian strata of the northwestern Canadian Shield. In keeping with the theme of this conference, data related to interpretation of tectonic history are emphasized.

Much of the information presented here is contained in numerous published reports. In addition, each author has contributed previously unpublished data, including information stemming from continuing studies in the Great Bear Lake (Donaldson) and the Lake Athabasca (Fahrig) regions.

The Rb/Sr isochron from the Dubawnt volcanics (Fig. 5) was provided by R.K. Wanless of the Geochronology Section of the Geological Survey of Canada. Figures 1 to 8 were drafted by Miss Anna Carlman, Carleton University.

LAKE ATHABASCA REGION: MARTIN FORMATION

Basin Geometry and Stratigraphy

The Martin Formation is a redbed and volcanic succession that outcrops in the Beaverlodge area, Saskachewan (Fig. 2), unconformably overlying gneisses and granites of the Tazin Group of probable Archean age. The formation occurs mainly in two areas 15 miles apart, the Martin Lake-Fredette Lake area, comprising about 48 square miles, and the Tazin Lake area, comprising about 37 square miles. Each area is in part bordered by normal faults. In most places the unconformity at the base of the formation is sharp but locally it is characterized by a regolith. The trace of the unconformity and lithology of the Martin rocks suggest deposition on a rugged erosional surface.

The Martin Formation is unmetamorphosed, fresh, and well consolidated. In the Martin Lake area where the succession is thickest and best exposed, it is at least 13,000 feet thick. The succession is as follows from bottom to top: basal conglomerate, lower arkose, volcanic flows and gabbro sills, upper arkose and siltstone. The flows and sills occur only in the Martin Lake area and are interbedded with some arkose. Conglomerate interbeds occur throughout the section. Late gabbro dykes are shown to cut only the Martin rocks below the flows. No younger rocks directly overlie the Martin Formation.

The basal conglomerate is made up of close-packed angular fragments in an arkosic to silty matrix. The fragments constitute 70% of the rock and are unsorted as to size, reaching a diameter of 3 feet at the base. Many fragments near the unconformity are derived from lithologies that correspond to the immediately underlying Tazin rocks. A few feet above the unconformity the fragments are mixed Tazin, i.e. mainly gneisses and granite. This lower unit is in part a talus breccia.

The arkose is crossbedded, ripple marked and in part thickbedded. Its grains are angular, closely packed and exhibit a bimodal size distribution with modes near 1 mm and between 0.1 and 0.5 mm. A cement composed of iron oxide and carbonate constitutes less than 5% of the rock. Immediately above the flows the arkose contains abundant grains and fragments of lavas. Throughout the section the arkose is interbedded with minor conglomerate and near the top of the section, with abundant siltstone.

The siltstone is brownish red and arkosic and is most abundant near the top of the section where it is interbedded with much arkose. Its grains have a bimodal size distribution with modes of 0.1 mm and between 0.02 and 0.05 mm. The amount of cement is small and is mainly iron oxide. The rock is thinly bedded, mud-cracked, and, in the Martin Lake area, has probable stromatolites near the top of the section.

The volcanic rocks are basaltic to andesitic and are somewhat spilitic. They are commonly amygdaloidal, most are porphyritic; and in rare outcrops they are pillowed. Gabbro sills occur below the flows and both the flows and sills are compositionally similar as shown by chemical and modal analyses.

Conglomerate interbeds are made up of about 50% fragments and abundant matrix. The fragments are well rounded with a maximum diameter of 15 cm. Most are typical Tazin rocks but a few are derived from the Martin Formation and possibly also from late gabbro dykes. At the south end of Beaverlodge Lake a related conglomerate is interbedded with white sandstone similar in appearance to the Athabasca Formation, ten miles to the south.

Paleocurrents

Directions of paleocurrents were obtained from 262 attitudes of crossbeds measured in five localities in the lower and upper arkoses of the Martin Lake succession (Fig.2). They broadly indicate that the material was transported in a south-southwesterly direction. Fahrig (1961) obtained the same general direction of transport on 41 readings. In detail, however, the data indicate that transport directions converged in the northern half of the Martin Lake area and that this area was a basin of sedimentation.

A few measurements of attitudes of crossbeds in conglomerate interbeds near the top of the Martin Lake area succession also suggest southwesterly transport, an inference further supported by the elongation of the Martin Formation of the Martin Lake-Fredette Lake area in a southwesterly direction (Fig. 2).

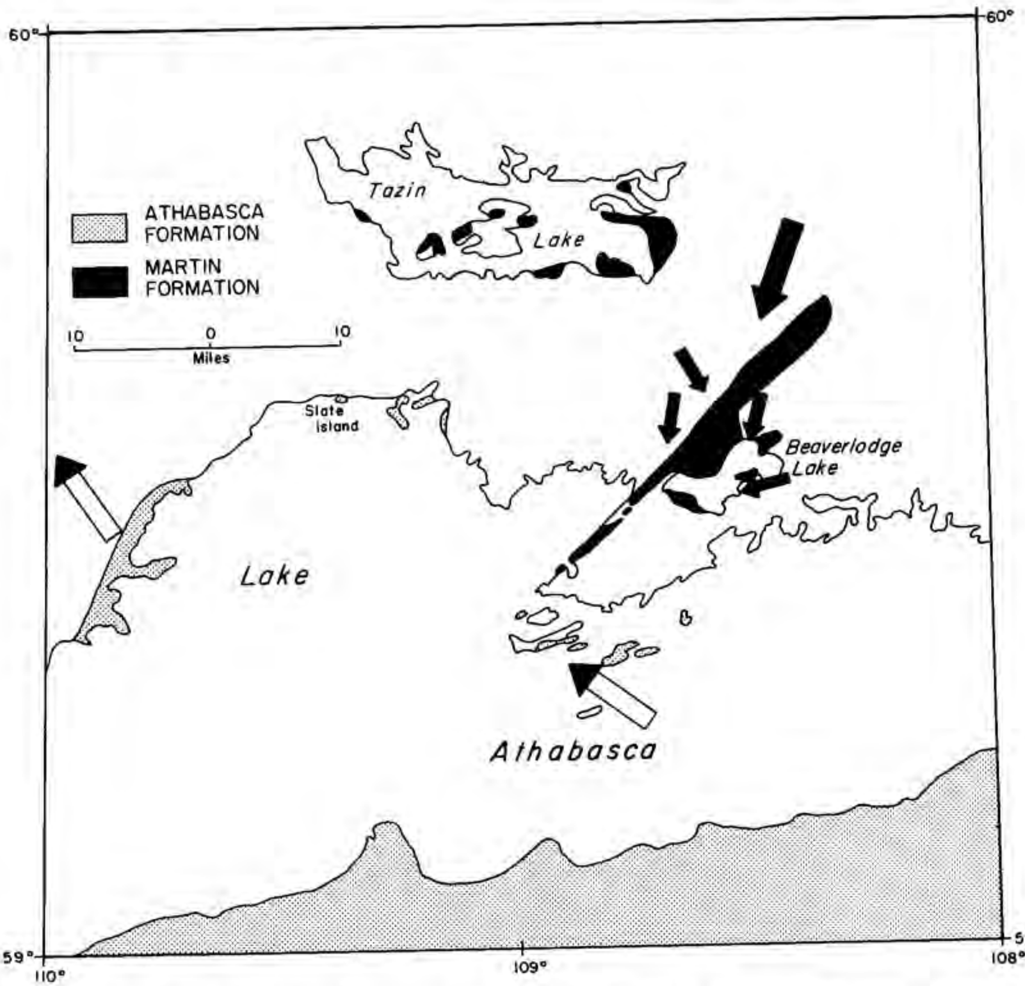

Figure 2. Paleocurrent directions in the Martin and Athabasca Formations near Beaverlodge Lake. Large arrows represent average trends.

Structure

The basins in which the Martin Formation occurs are both depositional and structural. Those in the Martin Lake-Fredette Lake area are elongated in a northeasterly direction whereas those in the Tazin Lake area are elongated east-west.

During sedimentation of the Martin Formation the basins had outlines similar to those that they exhibit at the present time. This is suggested by the basal conglomerate which has been largely derived from nearby subjacent rock, by the detailed pattern of transport indicated by the attitudes of crossbeds, and by the present bedding attitudes.

It is also apparent that these basins are structural features. Most of the major basins are bordered by faults and bedding planes along the basin margins presently dip as much as 90 degrees. The lower units have a pronounced dip whereas some of the beds in the upper part of the succession are almost flat-lying and little deformed, suggesting that deposition and deformation were contemporaneous. Although the shape and orientation of the amygdules and vugs in volcanic rocks indicate that these rocks were extruded on an almost horizontal surface, their present orientation bears evidence of tilting since their formation.

The basins containing the Martin Formation may therefore be regarded as taphrogeosynclines.

Age Relationships

The Martin Formation unconformably overlies metamorphosed rocks dated at 2,350 m.y. (Archean) by the Rb/Sr method (Aldrich and Wetherill, 1956). The Archean rocks have been subjected to retrograde metamorphism. Chloritized biotite so produced has been dated by the K-Ar method at 1,795 ± 90 m.y. These Archean rocks are cut by late gabbro dykes, one of which contains biotite yielding a K-Ar age of 1,835 ± 50 m.y. This dyke does not seem to cut rocks of the unmetamorphosed Martin Formation and thus these K-Ar ages put an upper limit on the Martin Formation of approximately 1,800 m.y.

The age of Martin Formation flows dated by the whole rock K-Ar method is 1,635 ± 180 m.y. A late gabbro dyke that cuts the basal conglomerate and the lower arkose has given by the same method a date of 1,490 ± 100 m.y. and finally, one of the sills below the flows has been similarly dated at 1,410 ± 100 m.y. This establishes a lower limit for the flows and the rocks of the Martin Formation below them of about 1,650 m.y. The rocks above the flows could be younger but because they are lithologically and structurally similar to the sediments below the flows they are probably of similar age.

Stockwell (1969) regarded the Martin Formation as Paleohelikian. Koeppel (1968) using Pb-U dates concluded that it formed between 1,930 ± 40 m.y. and 1,780 ± 20 m.y. which would place the Martin Formation in the Aphebian. The age of the Martin Formation is here regarded to lie between 1,830 m.y. and 1,650 m.y., this is, very early Helikian or very late Aphebian.

THE ATHABASCA TROUGH

Basin Geometry and Stratigraphy

Earlier work in the Athabasca Basin, in particular paleocurrent study (Fahrig, 1961), indicated that the Athabasca Formation was deposited in a trough structure within the Churchill (structural) Province. Convergence of fluvial paleocurrents suggests that material was transported into a platform depression. The dips of beds exposed at the surface were found, with few exceptions, to be very gentle, but surface data alone could not be used to define the present total geometry of the basin. Seismic work (Hobson and MacAulay, 1969) has since provided additional data on basin shape (Fig. 3). The seismic work does indicate that the formation now occupies a distinct basin in the underlying gneisses, and that within the basin there are two or three pronounced depressions. The present axes of these depressions trend northeasterly and the depth to the basement-complex refractor in the deepest of these is more than 5,000 feet. However, it should be emphasized that, because parts of the trough have been differentially tilted since the formation was deposited, these data relate to the sub-Athabasca surface and not to pre-Athabasca topography and thus the interior troughs and swells indicated by the seismic work probably were developed during or after sedimentation. A more detailed picture of the sub-Athabasca topography has been obtained locally from drilling. Figure 4 shows the logs of a series of drill-holes near the eastern edge of the basin, running in an east-southeasterly section, about parallel with paleocurrent direction. In this section the sub-Athabasca surface drops about 100 feet in nine miles; parallel sections have shown slopes up to 600 feet in seven miles.

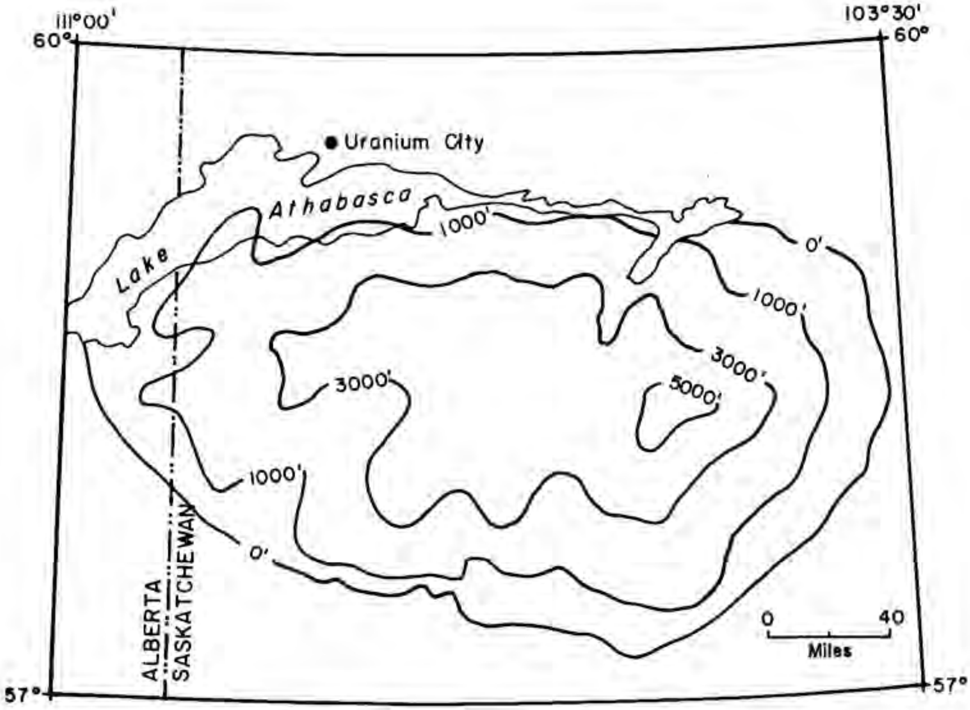

Figure 3. Isopach map of Athabasca Formation based on seismic data (from Hobson and MacAulay, 1969).

The bulk of the Athabasca Formation as it appears in outcrop is predominantly orthoquartzite with minor intercalations of shale and lenses of conglomerate. The conglomerate lenses appear to be most abundant in the basal part of the formation, particularly in the eastern part of the trough. Because paleocurrents suggest a trough-like basin of deposition, stratigraphic variation at depth within the interior of the basin seems likely. Figure 4 illustrates typical stratigraphy in the most easterly part of the basin. The

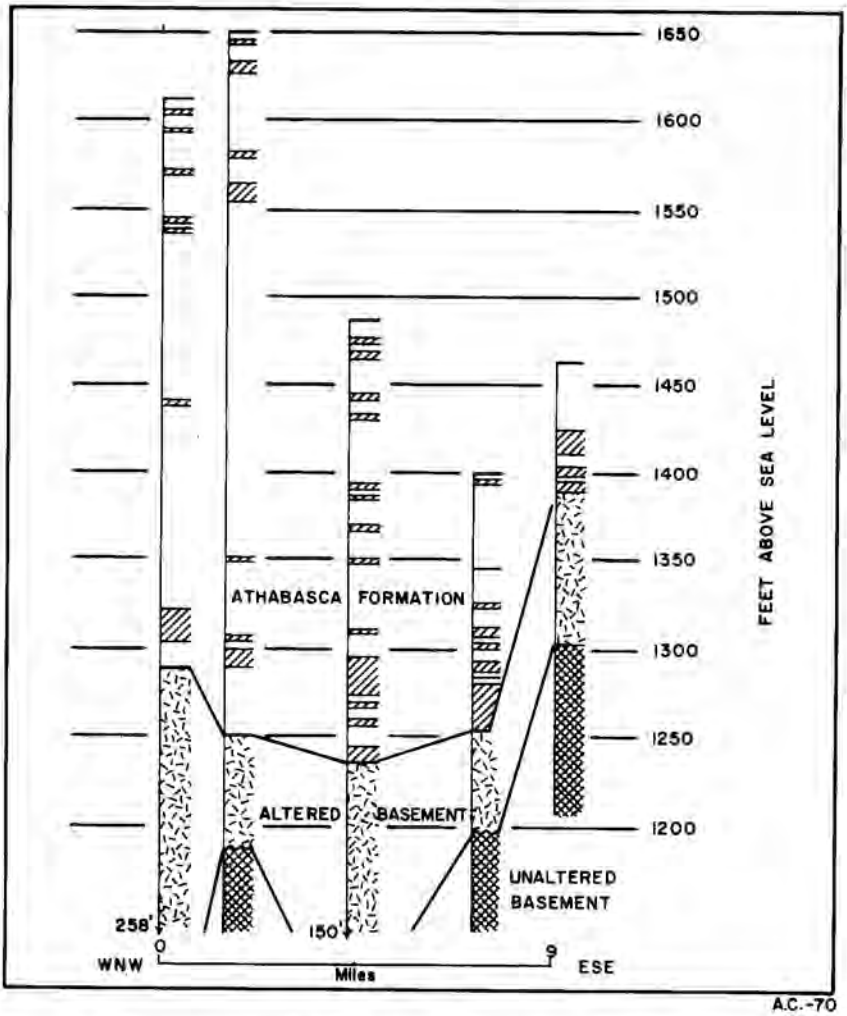

Figure 4. Lithology of Athabasca Formation from bore-holes along WNW-ESE section. Diagonal shading represents conglomerate, unpatterned intervals are sandstone. Data supplied by courtesy of Gulf Minerals Co.

cross-hatched intervals are predominantly poorly sorted conglomerates and the intervening beds are chiefly very coarse to coarse-grained sandstone, that contain pebbly lenses. The logs of these holes and of adjoining holes suggest no down current or lateral correlation except in a very general way. These and other cores from this area are virtually devoid of shale intervals although clay is a common interstitial component. The possible alignment of northeasterly trending basin depressions and swells with basement faults north of Lake Athabasca led Hobson and MacAulay (1969) to suggest that the depressions might be graben structures similar to those around Martin Lake. If so, the depressions would be expected to contain a basal redbed (molasse) sequence similar to the Martin Formation. Drilling in some of these deeper parts has, however, revealed a section consisting entirely of Athabasca sandstone with minor thin shale and shale-chip layers. This does not mean that molasse sedimentary basins are not present beneath the Athabasca Formation, but their existence has not yet been proven.

The clastic quartzose Athabasca fill is conformably overlain by the Carswell Formation, a carbonate unit preserved in a ring syncline as a result of cosmic intervention (Dence, 1965). Currie (1969) has divided the formation into two units. The lower consists of fissile dolomite with calcarenite and stromatolite zones; the upper consists of thickly bedded, massive dolomite. The preserved maximum thickness of the Carswell Formation is, according to Currie (1969) more than 500 feet. Because of the peculiar structure caused by meteorite impact in the Carswell Lake area, an entire section of underlying Athabasca Formation is here preserved. Seismic work indicates a total original thickness of Athabasca Formation in this area of about 2,500 feet.

Basin fill near the eastern margin consists of a roughly cyclical repetition of poorly sorted conglomerate layers containing predominantly subrounded to poorly rounded quartz pebbles up to five centimeters in diameter, and poorly to well sorted, predominantly coarse to very coarse sandstone. The cyclic units are not well defined because the conglomerates have sandy lenses and the sandstone has conglomerate lenses. All these units are typically crossbedded and interbeds of shale are virtually absent. Eighty miles to the west, the thickest preserved Athabasca section consists of deposits of cyclical sandstone and shale, equivalent at least in part to the conglomerate-sandstone cycles to the east. Even here, however, the shale forms less than 1% of the cycle. These sequences probably describe the arrangement of fill making up the Athabasca Formation in the eastern half of the Athabasca Basin, and may be typical of the entire basin. Both sequences are considered to be of fluvial origin. The conglomeratic sequence near the eastern limit of the basin resembles deposition in braided streams, whereas the sandstone-shale couple probably represents meandering river deposition. The greater variability of paleocurrent measurements (Fahrig, 1961) in the downstream direction implies a more complex meander pattern towards the western part of the basin. The shift from fluvial to marine conditions as evidenced by the change from the Athabasca Formation to the Carswell Formation might lead one to expect deltaic sedimentation in the upper western parts of the Athabasca Formation, but such a deltaic sequence has not yet been documented.

Drilling programs have provided useful information on the character of the sub-Athabasca rocks. Drill-holes near the eastern boundary have indicated that the basement gneisses are typically intensely altered for several feet below the unconformity. In places the gneisses now consist of grains and veins of quartz floating in a clay matrix. This intensive alteration gradually decreases with depth but geophysical probes in some holes can detect alteration to a depth of several hundred feet. The conglomerates in the eastern parts of the basin (Fig. 4) were probably derived locally by the rapid removal and dumping of material derived from the upper regolith zone. This suggests that the present eastern margin of the basin is not far from the eastern boundary of the basin existing when these conglomerates were deposited. The character of the bulk of the sands forming the Athabasca Formation, however, indicates a multicycle origin.

The paucity of shale even at depth in the Athabasca Basin is of some interest. The ubiquity of interstitial clay suggests that the streams transported large quantities of this size fraction, so presumably vast quantities of clay were carried along the trough and deposited west or northwest of Lake Athabasca in a geosynclinal basin. The Athabasca Formation at the present time occupies a volume of 10,000 cubic miles. If the shale to sandstone ratio was three to one, then the preserved Athabasca reflects the probable deposition of at least 30,000 cubic miles of shale in a geosynclinal basin to the northwest.

Paleocurrents

Festoon crossbedding, ripple marks of various types, planar crossbedding and current lineations have been observed in surface exposure and drill-cores of Athabasca Formation. Festoon crossbedding is the most obvious directional feature in outcrop and the axes of 1,200 festoon-crossbed troughs have been measured on bedding planes in surface exposures throughout the basin (Fahrig, 1961). Although these measurements are not evenly spaced over the basin, they probably provide an adequate regional indication of outcrop paleocurrent directions. Regional tilting of the western part of the Athabasca Basin has provided access to beds in a section transverse to the direction of transport. Because measurements in this section are concordant with measurements elsewhere in the basin the paleocurrent directions at depth in the basin likely are similar to the directions observed in surface rocks.

The paleocurrent data (Fig. 9) indicate that the detritus composing the Athabasca Formation was carried into the basin chiefly from the east and southeast. These directions converge towards the northwest, suggesting a comparable convergence of meandering tributaries. The crossbed directions exhibit a unimodal distribution with increased deviation from the mean in the western, downstream part of the basin.

Structure

The local deep regoliths at the base of the Athabasca Formation, overlain by rapidly deposited coarse clastics, suggests that sedimentation in the Athabasca Trough was initiated by a sudden change in conditions. This change was probably tectonic, involving depression of the area occupied by the Athabasca Formation relative to the surrounding Churchill craton. The tectonic mechanism is, however unknown. The proximity of old grabens filled with redbeds along the northern rim of the basin might lead to a suggestion that renewed movement on some of the faults that flank the grabens, or on the southwesterly extensions of these faults might have initiated the Athabasca cycle of sedimentation. On the other hand, the redbed basins may themselves have been preserved from erosion simply because they were depressed by independent regional tectonic movements that developed the Athabasca Trough. Filling along the axis of the trough by a thick sequence of cyclical fluvial sedimentary layers with no obvious large scale lithologic variation, implies that the trough was depressed at a fairly constant rate during this period of sedimentation.

The second phase in the filling of the Athabasca Trough consisted of marine transgression from the northwest and deposition of the platform carbonates of the Carswell Formation. Because only a small remnant of this formation has escaped erosion we can say very little about this phase of sedimentation. We can however say something about the timing of later uplift west of Lake Athabasca. The absence of the Athabasca Basin northwest of Lake

Athabasca indicates that there has been several thousand feet of erosion in this area relative to the interior of the Athabasca Basin. Because of the apparent lack of Helikian sedimentary rocks beneath the Devonian cover northwest of Lake Athabasca this uplift and erosion must have taken place between the time of formation of the Athabasca Basin in the Helikian and the deposition of the Devonian strata.

The probable overlap of Devonian and younger strata on the Athabasca is of particular importance if the Athabasca has contained conduits capable of transporting hydrocarbons from Phanerozoic or Proterozoic strata into the Athabasca Formation, because the presence of hydrocarbons in the Athabasca would be effective in the precipitation of uranium from percolating ground waters. One such area of broken Athabasca Formation is centered around the Carswell astrobleme. In this area the Athabasca was overlain by the Proterozoic Carswell Formation which contains evidence of the presence of organisms. Faults such as the Black Lake Fault have been active since deposition of the Athabasca and may have been active while Proterozoic and Phanerozoic strata containing biogenic material covered this area. For these reasons Phanerozoic faulting or disruption of the Athabasca Basin is of particular economic interest.

Age Relationships

The Athabasca Formation occupies an intracratonic basin on the Hudsonian orogen and is apparently cut by 1,200 m.y. diabase dykes of the Mackenzie swarm (Fahrig and Jones, 1969). This indicates that the formation is Helikian and was laid down between 1,200 and 1,700 m.y. ago. A closer estimate of the age of sedimentation may be obtained from Rb/Sr whole-rock work on the shaly intervals of the formation.

BAKER LAKE-THELON REGION

Stratigraphy

In the central part of the Churchill Province west of Hudson Bay, an area of about 35,000 square miles (Fig. 1) is underlain by unmetamorphosed sedimentary and volcanic rocks of the Dubawnt Group. Lying unconformably on gneiss, granite, amphibolite, phyllite and quartzite of the Hudsonian orogen, the Dubawnt Group comprises three stratigraphic sequences: a lower sequence of redbeds, a middle sequence of alkalic and acid volcanic rocks, and an upper sequence of quartz sandstones, pebbly sandstones, conglomerates and dolomites.

The redbed succession, confined largely to an area south of Baker Lake, consists of a lower, discontinuous conglomeratic unit, the South Channel Formation, and an upper unit of arkosic sandstones and mudstones, the Kazan Formation. These units, having an aggregate thickness that may exceed 17,000 feet, are characterized by pink to maroon colours, and an abundance of sedimentary structures: channels in the basal conglomerates; crossbedding, mud-chip conglomerates and primary current lineation in the sandstones; and ripple marks, dessication cracks, and clastic dykes in the mudstones.

Volcanic rocks compose a unit, probably less than 500 feet in maximum thickness, that lies with slight unconformity on the redbeds in the vicinity of Baker Lake, and westward overlaps the basement granites and gneisses. Biotite trachytes and quartz-feldspar latites are predominant rock types, but dacites, rhyolites, andesites, and interflow volcanoclastic units are locally abundant. Dull reddish brown, brick-red, pink, orange, and mauve colorations predominate, and many distinctive units can be traced widely on the basis of colour alone. Two formations have been mapped in the volcanic rocks near

Baker Lake. The Christopher Island Formation includes the trachytes and andesites; the Pitz Formation consists mainly of acidic feldspar porphyries. Chemical and mineralogical compositions as well as field relationships indicate that syenite of the Martell Formation is the probable intrusive equivalent of the Christopher Island Formation. Relationships of the syenite bodies (most abundant in the redbed basin south of Baker Lake) to wallrocks suggest that at least some are laccoliths.

The volcanic rocks are disconformably overlain by quartzose conglomerates and sandstones of the Thelon Formation, the areally most extensive Dubawnt unit. The Thelon Formation is flat-lying, probably less than 1,000 feet in maximum thickness, and shows a general upward transition from coarse conglomerates at the base to fine-grained sandstones and shales at the top.

A pre-Thelon regolith is best developed beneath the Thelon Formation on granitic basement rocks in areas where older Dubawnt units are missing because of erosion of non-deposition. However, regolithic patches also are locally preserved along contacts between Thelon strata and underlying volcanic rocks, demonstrating a post-volcanic age for the regolith.

Dolomite conformably overlies the Thelon Formation. Although predominantly fine-grained to lithographic, this unit also contains clastic zones that show crossbedding, ripple marks, desiccation cracks, and edgewise conglomerates. Stromatolites, oncolites, and oolites occur in distinctive units, and cherty interlayers commonly are stratigraphically controlled.

Diabase dykes of the northwesterly trending Mackenzie swarm intrude the Thelon Formation and older Dubawnt units, but are not visibly in contact with the dolomites. Small erosional remnants of basalt that rest on the Thelon Formation may be the extrusive equivalent of the diabase.

Paleocurrents

Crossbeds are very abundant in both the Kazan and Thelon Formations. A systematic regional study of crossbedding attitudes has revealed remarkably consistent paleocurrent patterns for both stratigraphic units (Donaldson, 1967). A northwest trending average direction of transport is indicated for the redbeds, and a westward direction of transport can be inferred for the Thelon Formation (Fig. 9).

Regional plots of station data for the Thelon Formation indicate that paleocurrents, as recorded by crossbeds, are not significantly related to present configuration of the Thelon Basin. This, coupled with compositional maturity of the sandstones and conglomerates (presupposing tectonic stability and relatively low relief), supports the conclusion that Thelon deposition extended well beyond the mapped formational boundaries.

The redbed basin, on the other hand, need not have been significantly larger than the presently preserved basin south of Baker Lake. The character of the redbeds is such that deposition is likely to have occurred during a period of great tectonic instability that involved contemporaneous faulting. Unconformable relationship of the overlying volcanic rocks indicates erosion, but such relationship could readily derive from a fault-controlled redbed basin not much larger than presently preserved.

Basin Geometry and Structure

The term 'basin' is here used to describe the Dubawnt Group in a structural sense only. The west edge of the Thelon Formation is abruptly truncated by faults, and several additional faults of significance occur within this unit, indicating that present distribution of this blanket-like unit probably is but an accident of preservation. Paleocurrent data support this interpretation (*see* previous section).

In contrast, the redbed basin, largely confined to the area south of Baker Lake, may well approximate configuration of the basin of deposition. Extreme compositional immaturity of the South Channel and Kazan Formations indicates a nearby source area of rugged topography. Post-depositional faults cut the basin, and one that follows the north shore of Baker Lake probably is responsible for the truncation of the basin along its northern margin. However, the tremendous thickness of the redbed sequence together with the requirement for maintenance of rugged topography suggests, in contrast to conditions for deposition of the comparatively thin but extensive and compositionally mature Thelon Formation, that faults were active during deposition of the redbeds. Such tectonic control of sedimentation has been established for other Proterozoic, as well as for more intensively studied Phanerozoic, redbed basins. The Baker Lake redbed basin represents an intracratonic basin of molasse deposition, with the South Channel Formation representing the near-source edge of basin fill.

Age Relationships

The Dubawnt Group rests unconformably on metamorphic and igneous rocks of the Hudsonian orogen, and the Thelon Formation is cut by northwesterly trending diabase dykes of the Mackenzie swarm that have yielded radiometric dates in a range from about 1,100 m.y. to 1,500 m.y. demonstrating that most of the Dubawnt Group is Paleohelikian. Although the dolomite unit of the Dubawnt Group is in apparent conformity with the Thelon Formation, it is nowhere visibly intersected by the Mackenzie dykes, and therefore is not necessarily of Paleohelikian age (a small outlier of fossiliferous Ordovician limestone similarly rests on Thelon strata in the northern part of the region).

Samples of Dubawnt volcanic rocks have yielded a range of K-Ar ages that overlaps the mean age of the Hudsonian orogeny as calculated (Stockwell, 1964) on the basis of regionally distributed samples. To explore this overlap problem by a second radiometric method, a suite of samples has been dated by the Rb/Sr method. Five of fourteen samples of fresh, aphanitic to fine-grained flows were suitable for analysis. One sample is a trachyte from the Christopher Island Formation, and the other four are from the less common rhyolitic flows of the Pitz Formation.

The data for the five rock samples define a very good whole-rock Rb/Sr isochron (Fig. 5), the slope of which indicates an age of 1,732 ± 9 m.y. Because of the high degree of colinearity shown by the values, the Dubawnt volcanic rocks can be interpreted as a system that has remained closed since crystallization. The isochron thus provides a reliable age of extrusion for the Dubawnt lavas and indicates a maximum age for the Thelon Formation as well as a minimum age for the Kazan and South Channel Formations.

TABLE I

K/Ar Ages of Dubawnt Igneous Rocks

Sample No.	Dated Mineral	Lithology	K/Ar Age
GSC59-35	Biotite	Porphyritic flow	(1515 m.y.)*
GSC60-60	Biotite	Trachytic dyke	1720 m.y.
GSC61-100	Biotite	Porphyritic flow	1770 m.y.
GSC64-74	Phlogopite	Trachyte	1685 m.y.
GSC65-73	Biotite	Trachyte dyke	1690 m.y.
GSC65-74	Biotite	Syenite	1715 m.y.

Mean K/Ar Age = 1716 m.y.

*The age of sample GSC59-35 is regarded as anomalous in comparison with the other five ages, in view of the Rb/Sr isochron age. It therefore was not used in calculating the mean age.

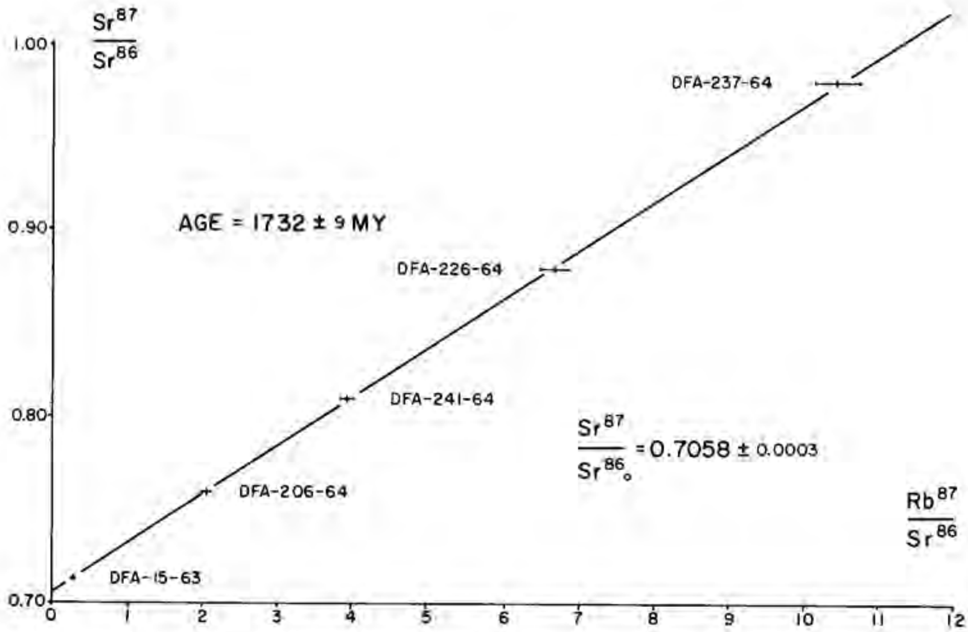

Figure 5. Rb/Sr isochron of samples from the Christopher Island Formation and Pitz Formation volcanics, Baker Lake region.

The K-Ar ages for biotites in samples of Christopher Island Formation and Martell Syenite (Table I) are in good agreement with the Rb/Sr age. Concordance of the ages obtained by these two independent radiometric methods of dating mutually supports the validity of the results. Of particular interest is the indication, implicit in the concordance, that biotites in the Dubawnt volcanic rocks have not lost significant amounts of argon.

BATHURST INLET REGION

Basin Geometry and Stratigraphy

Strata of Helikian age are exposed on many of the islands in Bathurst Inlet, on the mainland south and west of the Inlet and on Kent Peninsula (Fig. 6). The area underlain by these rocks, here termed the Bathurst Basin, covers about 5,000 square miles. The 'basin' is bounded on the east by granitic and metasedimentary rocks of Archean age, and by sediments of Aphebian age and on the west by the Bathurst Trench (Wright, 1957; Tremblay, 1968), a prominent linear topographic feature and graben-like structure that extends 200 miles southeastward from the Arctic coast. The lowermost Helikian strata occur in and along the Trench as far as 80 miles from the Inlet.

Helikian strata in the Bathurst Basin comprise four unmetamorphosed and only mildly deformed formations having an aggregate thickness of more than 15,000 feet.

Sandstone and conglomerate of the Tinney Cove Formation (Tremblay, 1968) outcrop only in the southern part of the Bathurst Basin where they lie unconformably on folded Aphebian strata and locally, on Archean basement

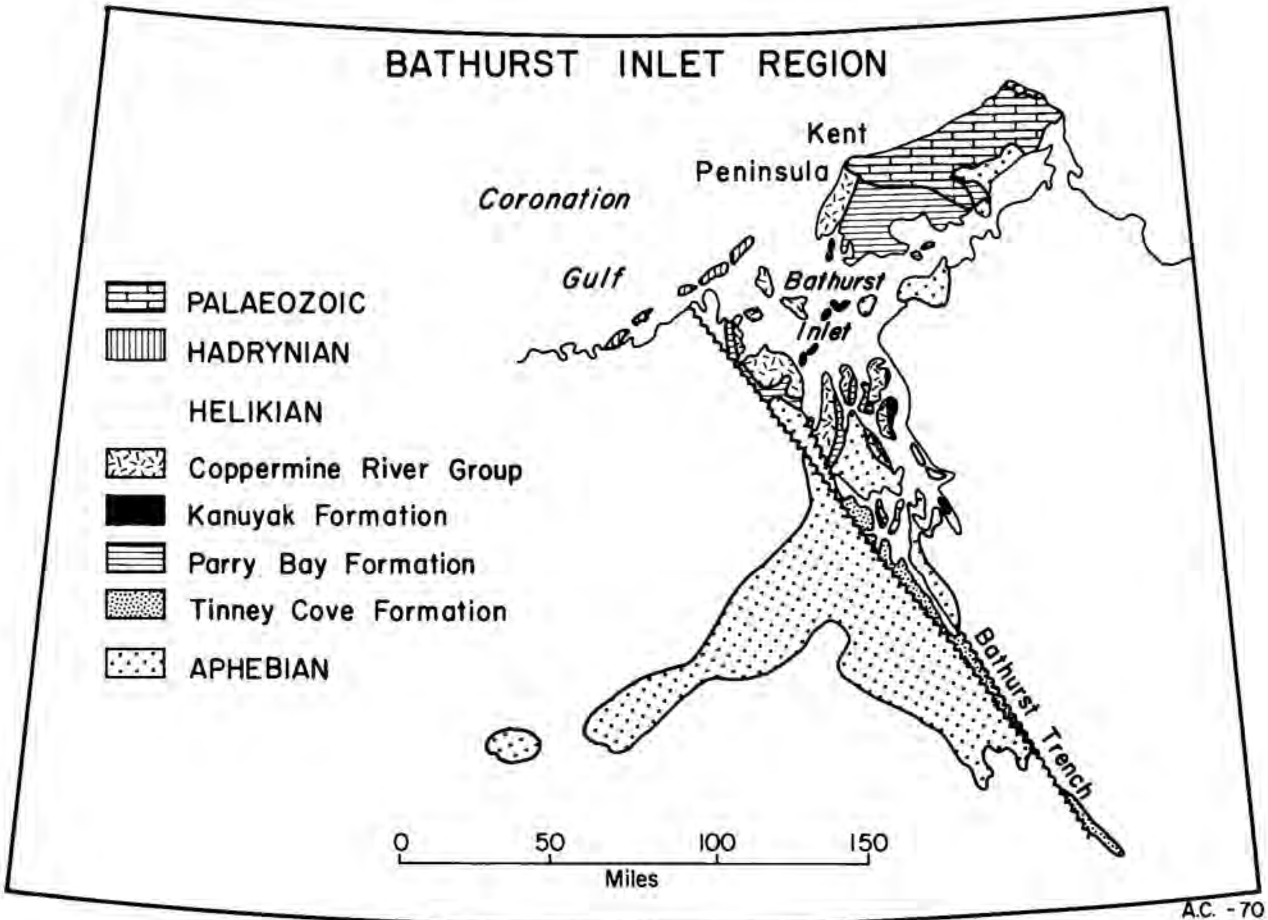

Figure 6. Geology of Bathurst Inlet region.

rocks. The formation is at least 6,000 feet thick at the extreme southern end of the Bathurst Trench where it is known as the Ellice Sandstone (Tremblay, 1968). The formation is at least 2,000 feet thick in the southern part of the Inlet but is missing a few miles farther north. The conglomerate locally attains a thickness of 1,100 feet but is absent from some sections. It is composed of unsorted rounded to angular pebbles, cobbles, and a few boulders of pink quartzite, quartz and granitic gneiss and some argillite and carbonate in an arkosic matrix. The kinds of clasts and their relative proportions in strata immediately above the unconformity vary from place to place, closely reflecting the lithologies of the subjacent rocks.

The conglomerate grades upward into medium to coarse-grained grey, cream, and pink to red crossbedded orthoquartzite characterized by an ubiquitous content of clay, as matrix and as discrete clots, in amounts up to 10%.

Below the Ellice Sandstone is a sequence of red interlaminated carbonate, sandstone, and shale, 150 feet thick, that overlies 150 feet of coarse-grained grey calcareous sandstone and conglomerate containing clasts of grey quartzite and quartz (Tremblay, 1968; in press). These rocks probably underlie the conglomerate and sandstone unconformably. Their relation to the Aphebian strata of the Goulburn Group exposed in this region is unknown. They are tentatively correlated with the basal redbed successions in the other Helikian basins.

The Parry Bay Formation (Fraser, 1964), comprising dolomite with minor intercalated sandstone and shale, is exposed chiefly in the central and northern parts of Bathurst Inlet where it is preserved in downthrown fault blocks capped by basalt. The formation probably overlies the Tinney Cove Formation although contacts between these formations have not been observed. Parry Bay strata are overlain unconformably by the Kanuyak Formation (O'Neill, 1924) which in turn is overlain conformably by Coppermine River basalt. On Kent Peninsula, however, basalt rests conformably on Parry Bay strata. Dolomite of the Parry Bay Formation exposed on the islands in Bathurst Inlet is at least 540 feet thick. Gently inclined to flat-lying dolomite and interlayered sediments on Kent Peninsula are possibly as much as 7,000 feet thick.

The basal section of the Parry Bay Formation includes white and buff sandstone, reddish brown dolomite and shaly dolomite, but the bulk of the formation consists of thin-to thick-bedded, pale grey to pink fine-grained dolomite and limestone. Thin layers and lenses of chert, and arenaceous and oolitic beds are numerous as are beds of domical and conical stromatolites. The latter are associated locally with dolomite containing interbeds of black shale and with black, bituminous material that occurs as vug fillings and as disseminations. At the top of the Parry Bay Formation the dolomite is interbedded with red and green shale.

Unconformably overlying the dolomite of the Parry Bay Formation is a calcareous redbed succession known as the Kanuyak Formation (O'Neill, 1924). The thickness of this formation attains a maximum of about 200 feet but thins rapidly to the west and to the north and is absent from Helikian sections on the mainland. A conglomerate at the base of the formation is composed of angular fragments of chert and carbonate in a fine-grained matrix of calcareous sandstone. The middle and upper parts of the formation consist of fissile grey shale and arenaceous limestone overlain by red calcareous shales, siltstones, and tuff.

Kanuyak strata are overlain conformably by 2,500 feet of dark grey to green, fine-grained to aphanitic, massive, amygdaloidal, and pillowed basaltic flows of the Coppermine River Group (O'Neill, 1924). The basalt is capped by conformable black hematitic argillite succeeded by 100 feet of massive, pink and white, crossbedded quartzite.

Helikian sediments and flows of the Bathurst Basin are succeeded unconformably by sandstone and shale belonging to the upper part of the Coppermine River Group (O'Neill, 1924) of probable Hadrynian age, and by flat-lying Paleozoic sandstone and dolomite. All rocks in the basin except those of Paleozoic age are intruded by dykes and sheets of diabase and gabbro.

Paleocurrents

Paleocurrent directions inferred from crossbed attitudes indicate that the sediments of the Tinney Cove Formation were carried from a source lying to the east of the Bathurst Basin (Fig. 9).

The data consist of 45 measurements, 35 of which were obtained from the Ellice Sandstone at the southeastern end of the Bathurst Trench (Tremblay, in press), and 10 from sandstone in the central part of Bathurst Inlet. The first group yields a mean azimuth of 235 degrees and the second group, a mean of 310 degrees. The mean of all measurements is 243 degrees, a direction transverse to the elongation of the Bathurst Basin, which suggests that the sandstone unit once extended beyond present basin margins.

Because the mean direction (310 degrees) of the second group is based on so few measurements, it cannot be regarded as an established paleocurrent trend for the central part of the Bathurst Basin. The wide dispersion in the data from which this mean was calculated possibly reflects local variations in transport direction produced by initiation of subsidence along the Bathurst Trench during deposition of the Ellice and Tinney Cove sandstones.

Structure

The region extending from the southern end of the Bathurst Trench north to Kent Peninsula contains five successively younger deformational and depositional basins that record sedimentation from Archean to Paleozoic time. The Bathurst Basin truncates the northeastern margin of the Aphebian basin occupied by strata of the Goulburn Group (Fig. 6). Both basins are flanked by Archean metasediments that can be traced 30 miles beyond the southern limit of Goulburn outcrop. Late Precambrian and Paleozoic sediments exposed along the Arctic coast belong to younger basins that are marginal to and encroach upon, the Bathurst Basin. The distribution of these basins suggests a regional tilt northward in Hadrynian or post-Hadrynian time, an interpretation supported also by a north-south zonation in the distribution of diabase sheets. Those sheets of Hadrynian age are found near or along the Arctic coast, whereas older sheets occur farther south. The present margins of the basins are partly determined by post-depositional deformation, but in at least one basin, the Aphebian, a great thickening of sediments in and near the Bathurst Trench identifies this area as a centre of deposition.

Aphebian strata in this region have been folded around north-trending axes and near the Bathurst Trench the folds are much tighter. Helikian and younger rocks, on the other hand, are very little deformed which suggests that folding of Aphebian strata may have been associated with strike-slip movement along the Trench and that this phase of faulting ended before Helikian time.

Displacement along the Trench, estimated from the horizontal separation of Aphebian formations, is about 20 miles and the sense is sinistral. Displacement of this order can be inferred also from published aeromagnetic maps of this region. The observed separation could also have resulted from simple uplift or from rotational faulting, but neither of these alternatives adequately explains the close folding of Aphebian strata east and west of the Trench.

A tensional phase of fault movement followed the compressional (strike-slip) phase and is reflected in the redbed sequence preserved at the south end of the Trench. These sediments are inferred to have been deposited in a fault-basin in response to rapid and continued uplift in adjacent source areas. An unconformity at the top of the redbed sequence marks cessation of the conditions of redbed deposition, but the basal conglomerate of the overlying Ellice Sandstone and the equivalent Tinney Cove Formation indicate some renewal of block faulting, which was subsequently followed by deposition under more stable conditions of fluvial sandstone. Persistence of stable conditions is recorded in the central and northern parts of the Bathurst Basin by the shallow-water marine carbonates of the Parry Bay Formation.

Renewed tensional stress with concomitant block faulting resulted in depression and preservation of strata especially evident in the central part of the basin where Helikian strata, capped by basalt, are downthrown in a series of steps to the east along assumed northerly and northwesterly trending faults. At least 3,000 feet of section are preserved in this way. Faulting in response to tensional stress is compatible with the extrusion of plateau-type basalt through fissures now occupied by the northwesterly trending diabase dykes of the Mackenzie swarm. The tuffaceous Kanuyak sediments that underlie the flows, and the red sediments that overlie them, are part of this volcanic cycle. Local unconformities such as the one separating the Parry Bay and Kanuyak Formations record tilting, probably associated with the normal faulting.

Age Relationships

The upper and lower age limits for Helikian strata in the Bathurst Inlet region are determined by the age of the underlying Goulburn Group sediments and by the age of the Coppermine River basalt.

Sediments of the Goulburn Group have been correlated with those of the Epworth Group (Fraser and Tremblay, 1969) known to be intruded and metamorphosed by granite that dates at 1,760 m.y. This age sets a lower limit to the age of the Goulburn Group and a probable upper limit to the age of the Helikian rocks. A whole-rock K-Ar age of Coppermine River basalt from Bathurst Inlet of 915 m.y. lies within the range of ages determined for basalt flows of the Coppermine River Group southwest of Coronation Gulf. The best estimate of age for these flows, 1,200 m.y., is comparable to that of the related diabase dykes of the Mackenzie swarm (Fahrig and Jones, 1969). This age (1,200 m.y.) represents an approximate lower limit for the Helikian succession in the Bathurst Basin. The age of this succession is therefore considered to lie between 1,760 m.y. and 1,200 m.y.

The most recent movement along the Bathurst Trench followed the extrusion of the Coppermine River basalt (1,200 m.y.) and preceded the emplacement of diabase sheets (about 600 m.y.), the youngest igneous event represented in the Bathurst Inlet region.

GREAT BEAR LAKE REGION

Stratigraphy

Helikian and Hadrynian lithologies between Great Bear Lake and the Arctic coast (Fig. 7) compose a stratigraphic record thicker and more varied than those in the three regions previously discussed. A number of similarities are apparent, however, such as a characteristic lack of significant metamorphic effects, maturity contrasts in adjacent sedimentary strata, volcanic associations, and a regolith in a stratigraphic position comparable to that of the Athabasca and Baker Lake regions.

The oldest rocks considered here are redbeds of the Cameron Bay and Echo Bay Groups, confined largely to the eastern side of McTavish Arm, Great Bear Lake. The relationship between these two groups is not definitely established, although Mursky (1963), who estimated the thickness of the Echo Bay Group to be at least 4,300 feet, has presented evidence that the Cameron Bay is younger. Both groups contain abundant arkose, conglomerate, and argillite. Volcanic rocks, predominantly andesites and dacites, compose the upper part of the Echo Bay Group. Many of these volcanics are porphyritic, closely resembling flows of the Dubawnt Group near Baker Lake. Marked initial relief is recorded in a few places by configuration of the unconformity beneath the Echo Bay Group. Many redbed occurrences are flanked by faults, indicating that present preservation may be largely a function of structure. Compositional immaturity of the sediments suggests the possibility that these and related faults were active at the time of deposition.

The redbeds rest on granites and gneisses, and are cut by younger granites. The more extensive Hornby Bay Group rests unconformably on all these rocks (as well as on folded Aphebian strata of the Epworth Group). A

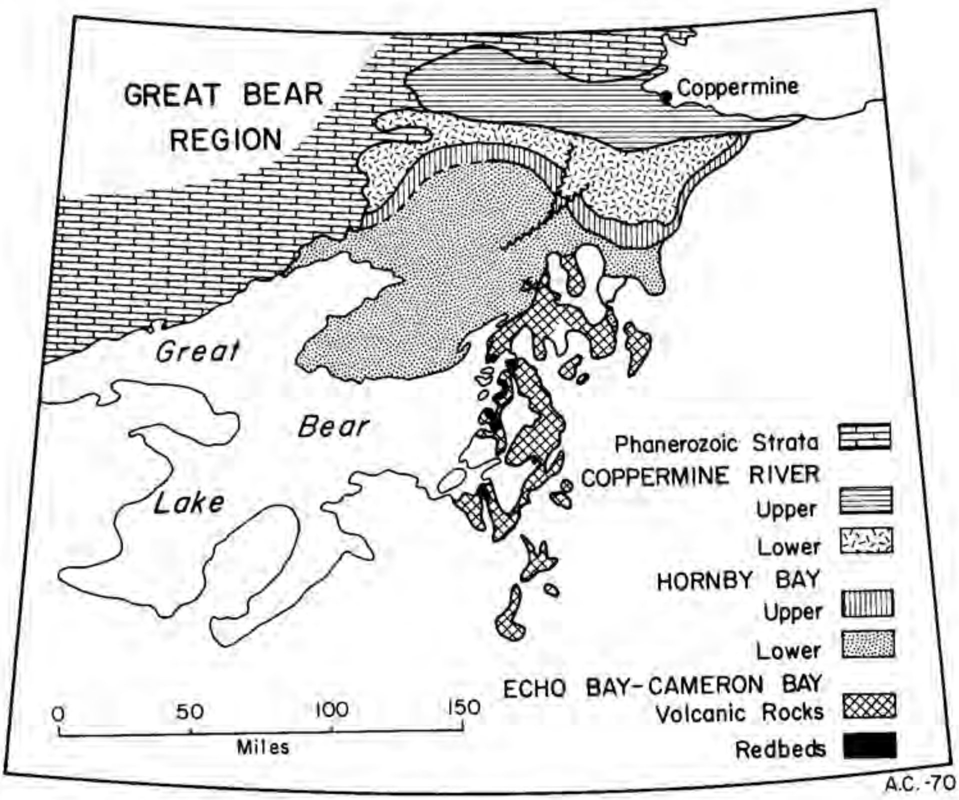

Figure 7. Geology of Great Bear Lake region.

regolith is locally preserved beneath the Hornby Bay Group, particularly where this unit rests on gneiss or granite. This and younger groups of the region are currently being investigated on a scale more detailed than previously attempted; recognition of additional stratigraphic units as well as intra-group unconformities (Baragar and Donaldson, 1970) will require some revision of stratigraphic nomenclature. Existing nomenclature is used herein pending completion of the stratigraphic studies. One of the recently recognized unconformities occurs within the Hornby Bay Group, and components of this unit will be informally designated 'lower' and 'upper' for purposes of this report.

In ascending order, the 'lower' Hornby Bay Group consists of a unit of sandstone and minor conglomerate with a maximum total thickness of as much as 3,000 feet, a unit of stromatolitic dolomite up to 2,000 feet thick, and a succession of mudstones, siltstones, and sandstones, as much as 2,000 feet thick. The 'upper' Hornby Bay Group, resting unconformably on these strata, consists of sandstone, siltstone and mudstone having a maximum aggregate thickness of less than 1,000 feet, and a remarkably continuous unit of dolomite in which distinctive members can be traced for many miles.

The 'upper' Hornby Bay strata are conformably overlain by the Coppermine River Group which, as presently defined, also contains a major unconformity. The 'lower' Coppermine River Group consists of a succession of basaltic flows with an aggregate thickness possibly in excess of 10,000 feet, overlain by a second unit, up to 4,000 feet thick of red sandstones, shales, and intercalated basaltic flows. The 'upper' Coppermine succession is an assemblage of sandstones, shales, argillites, dolomites, and limestones having a total thickness of about 4,000 feet. These strata, intruded by several thick diabase sills, form a monoclinal plate that dips gently towards the Arctic coast.

Paleocurrents

A few measurements of crossbedding inclinations in the Cameron Bay-Echo Bay redbeds along the east shore of Great Bear Lake suggest westward transport, but an intensive study of current indicators in these rocks will be required to fully describe the paleocurrent patterns and relate them to basin geometry.

Crossbedding attitudes indicate a marked westward paleocurrent trend for sandstones of the 'lower' Hornby Bay Group (Fig. 9), the mean azimuth of 622 measurements being 271 degrees. Most of these data are from the central part of the area; additional measurements will be necessary to detect any regional variations such as have been demonstrated for the Athabasca and Thelon Formations.

Limited data for the 'upper' Hornby Bay sandstones suggest a northwesterly trend of crossbed inclinations, and small-scale crossbeds together with stromatolite elongations reflect some bimodal transport for the upper dolomite of the Hornby Bay Group. The latter trend possibly reflects onshore and offshore tidal currents in a shallow-water environment.

Complex transport patterns are indicated for the overlying Hadrynian strata ('upper' Coppermine River Group). Although based on scanty data, north to northwesterly directions of transport appear to be most common.

Basin Geometry and Structure

Four significant episodes of sedimentation can be recognized in the Great Bear region, and termination of each episode can be related to tectonism.

The first episode (possibly late Aphebian rather than Helikian; *see* next section) led to rapid deposition of the Echo Bay and Cameron Bay redbeds. The basins containing these sediments, although not studied in detail, appear to be similar to the basins that contain the lithologically similar Kazan and Martin Formations. Considerable thicknesses are indicated; in one area the lower clastic unit of the Echo Bay Formation is more than 4,300 feet thick (Mursky, 1963).

The redbed basins were deformed prior to the second sedimentary episode, which led to deposition of the extensive sandstones and conglomerates of the 'lower' Hornby Bay Group. This clastic unit and overlying dolomite, mudstone, and sandstone, form a platform sequence that possibly thickens westward (down the paleoslope indicated by crossbeds), and in gross geometry constitutes a tabular blanket that has been folded and faulted by deformation which intensifies westward.

The third episode led to deposition of the 'upper' Hornby Bay dolomite and related strata. The northerly swing of crossbedding azimuths may reflect shift and rotation of the axis of sedimentation. Deep-water equivalents may have been deposited northwest of the exposed edge of the shallow-water dolomites. Flows of the 'lower' Coppermine River Group and overlying redbeds form a succession conformable with the 'upper' Hornby Bay Group, and thus the redbeds can be regarded as part of the same cycle, but the abundance of feldspar and rock fragments in these redbeds clearly indicates a less stable depositional environment.

The fourth episode of sedimentation involves predominantly shallow-water deposition of the Hadrynian strata ('upper' Coppermine River Group). Although the underlying basalts and redbeds are but gently folded, they have been significantly truncated by erosion, and thus a long interval may be recorded by the unconformity. Similar Hadrynian strata are exposed to the northwest in the Hornaday River Basin (Cook and Aitken, 1969), and on Victoria Island where they are known as the Shaler Group (Thorsteinsson and Tozer, 1962). Significant thicknesses of these rocks may lie beneath the intervening areas now covered by Paleozoic strata.

Age Relationships

The Echo Bay and Cameron Bay rocks, like the redbeds and volcanic rocks of the Dubawnt Group, are close in age to the Aphebian-Helikian boundary. An intrusive porphyry that 'possibly' cuts Cameron Bay strata has been dated at 1,765 m.y. by the K-Ar method. The fact that, in addition to the redbed associations, many of the porphyritic flows of the Echo Bay Group are strikingly similar to some of the Dubawnt Group volcanic rocks provides justification for mention in this report. These similarities merit careful evaluation before the possibility of correlation is rejected on the basis of K-Ar dates.

The Hornby Bay Group in places rests directly on basement gneisses and granites dated at 1,745 and 1,765 m.y. by the K-Ar method. In addition, the 'lower' Hornby Bay rocks are cut by the Muskox intrusion which has a probable age of more than 1,100 m.y. Basalts near the top of the Coppermine River flow sequence have yielded ages as young as 735 m.y. and diabase dykes that cut the associated redbeds have provided similar ages, but these ages are generally regarded as dates on rocks that have lost argon.

REGIONAL CORRELATIONS

Helikian basins in the Lake Athabasca, Baker Lake, Bathurst Inlet, and Great Bear Lake regions, although separated by hundreds of miles, contain strata that are remarkably similar in lithology and sequence. These similarities, coupled with radiometric ages of critically related igneous and metamorphic rocks support the proposition that the strata in these basins are correlative as shown in Figure 8.

Composite Helikian sections typically consist of a basal redbed and volcanic succession overlain unconformably by a blanket sandstone, which in turn is overlain by carbonate capped conformably by basalt and associated red sediments. Exceptions to this simple sequence occur in basins subjected to local crustal instability and in basins where part or all of the uppermost stratigraphic unit is missing, possibly as a result of erosion.

The basal redbed and volcanic sequences of the four Helikian basins shown in Figure 1 comprise (1) the Martin Formation; (2) the South Channel, Kazan, Christopher Island, and Pitz Formations; (3) redbeds at the base of the Ellice Sandstone; (4) the Echo Bay and Cameron Bay Groups. The Et-Then Group exposed in the East Arm, Great Slave Lake (Hoffman, 1969), exhibits all the features that characterize the redbeds in the other basins but is not at present capped by younger strata. In each area the red sediments compose thick sequences of arkosic sandstones and conglomerates that contain abundant structures indicative of fluvial transport and rapid deposition in fault-controlled basins. The sediments rest unconformably on folded Aphebian strata or on granitic terrane of Aphebian or Archean age and all, with the exception of the Et-Then, occur near, or are overlain by, deposits of blanket sandstone of distinctive character. Echo Bay rocks are intruded by Aphebian granite but locally, porphyry dykes similar in lithology to Echo Bay volcanics intrude the granite. The age of these volcanics and related redbeds may therefore be considered to straddle the Aphebian-Helikian boundary. Radiometric ages from the volcanic rocks in the Lake Athabasca and Baker Lake-Thelon regions support this conclusion and substantiate the correlation of the redbed successions.

Succeeding the redbeds are the flat-lying to gently inclined sandstones of the Athabasca Formation (1), the Thelon Formation (2), the Tinney Cove Formation (3), and the lower Hornby Bay Group (4) (Fig. 1). Each is a coarse to fine-grained orthoquartzite from 1,000 to 6,000 feet thick. Conglomerate occurs at the base and as interbeds in each succession. Clay in amounts up to 10% is a ubiquitous component of each sandstone, occurring both as matrix and as small dispersed aggregates. Each sandstone displays structures characteristic of fluvial transport and deposition and each is succeeded by carbonate considered to have formed in a shallow-water marine environment.

The carbonate units rest conformably on the sandstones and include the Carswell Formation (1), an unnamed dolomite overlying the Thelon Formation (2), the Parry Bay Formation (3), and dolomite of the Hornby Bay Group (4) (Fig. 1). In each basin the carbonate unit is predominantly a stromatolitic dolomite containing oolitic and arenaceous interbeds and, in some places, interlaminated chert.

Carbonate units in the Bathurst Inlet-Great Bear Lake region are overlain conformably by a thick succession of massive, amygdaloidal, and pillowed basalt and associated red sediments. In the Baker Lake-Thelon area amygdaloidal basalt is stratigraphically above the sandstone of the Thelon Formation but its position relative to the dolomite unit is uncertain. In this paper the basalt is tentatively assumed to overlie the carbonate (Fig. 8). In the Athabasca basin the carbonate unit is the youngest preserved.

O'Neill (1924) correlated the Coppermine River basalts at Bathurst Inlet and Great Bear Lake on the basis of lithologic similarity. This correlation is supported also by the stratigraphic position of these rocks relative to the underlying units and to the unconformably overlying Hadrynian strata as well as by radiometric dates on the basalts of both regions.

Diabase dykes similar in age to the basalt are known to cut sandstone of the Athabasca and Thelon Formations and confirm the age of these formations as pre-Coppermine River.

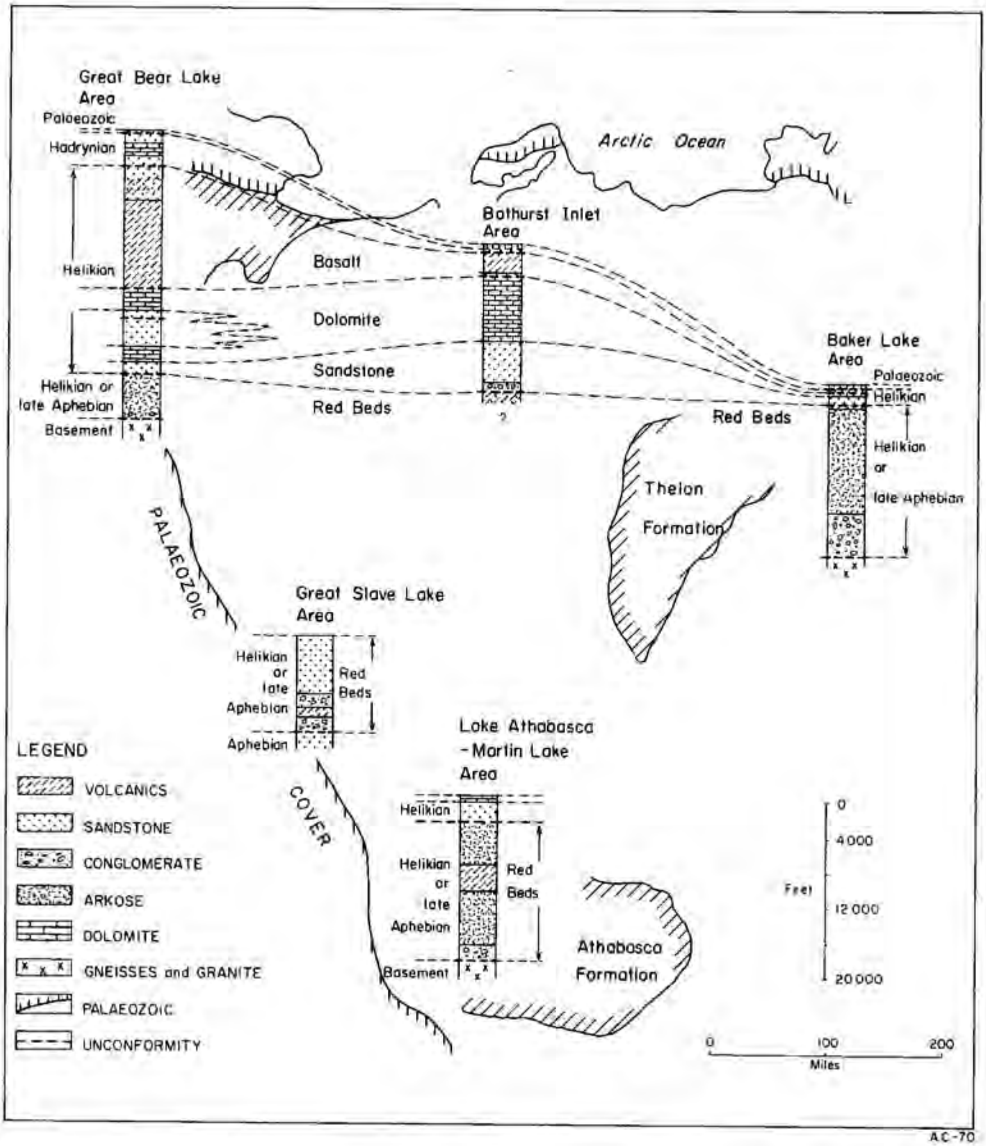

Figure 8. Composite sections of redbeds and cover rocks in the northwestern Shield, showing correlation of Helikian Formations at Baker Lake, Bathurst Inlet and Great Bear Lake.

SUMMARY OF PALEOCURRENT DATA

Paleocurrent directions inferred from attitudes of crossbeds in redbed and younger Helikian basins of the northwestern Shield are shown in Figure 9. Comparison of these data reveal significant differences in the evolution of the redbed and the sandstone basins.

In the redbed basins the dominant direction of transport is to the west but individual transport directions range from south-southwest at Lake

Athabasca to northwest at Baker Lake. Within any one basin, however, the paleocurrents show transport from many directions, converging towards basin interiors. There is also a crude correlation between basin elongation and current direction, which further supports the interpretation that the present redbed basins are remnants of true depositional basins.

Paleocurrent means for the sandstone that overlies the redbeds are west to northwest and have a narrower dispersion than do the averages for the redbeds. Within basins the paleocurrents exhibit a scatter that is less than that in the redbeds. In many cases the paleocurrents are subparallel with basin margins, indicating that transport took place along as well as across the basins. The data suggest that the resulting depositional cover at one time extended well beyond present areas of preservation and may well have been continuous between basins.

At Lake Athabasca, crossbed measurements made over a thick stratigraphic section of the sandstone in the western part of the basin indicate no appreciable departure from a northwesterly trend in transport direction, whereas crossbeds in the lower and upper sandstone units of the Great Bear Lake region record a change in current direction from west to northwest.

The dominant paleocurrent trend in the sandstone at Bathurst Inlet is westerly but a few measurements from the northernmost exposures of sandstone show considerable scatter with a northerly trending mean. The variation in dispersion of paleocurrents in this region may reflect a local influence of Bathurst Trench topography.

The few paleocurrent data now available from Hadrynian sediments of the Great Bear Lake region indicate a complex pattern of transport in which a northwesterly to northerly trend is apparent.

PALEOGEOGRAPHY AND TECTONISM

Near the end of Aphebian time great thicknesses of red sediments of molasse type were deposited in fault-controlled basins scattered across the northwestern part of the Shield (Fig. 9). The extreme compositional and textural immaturity of these sediments reflects a continuously rising rugged source terrain, and centripetal paleocurrent patterns suggest preservation in depositional basins not significantly greater in areal extent than they are today. The basins are partly filled by acidic to basic lavas, the extrusion of which probably was triggered by tensional faulting. Near Baker Lake the flows spread beyond basin margins and represent the last stage of basin development. The fault-basin cycle was followed by a long period of quiescence during which a thick regolith formed.

Succeeding Helikian sedimentation was induced by tilting of the crust towards the west, possibly attended by the development of gentle depressions in the vicinity of redbed basins. These Helikian sediments, now preserved as erosional remnants ('basins') on the craton, are inferred from shape, lithology, and paleocurrent patterns to have been at one time much more extensive and may have covered much or all of the area lying between preserved Helikian basins. The present shape and size of these basins is considered to be in part depositional and in part deformational.

The probable extension of a homogeneous clastic and carbonate couple across much of the northwestern Shield implies the existence of a broad stable platform. In Bathurst Inlet, however, clastic deposition began with the accumulation of boulders and coarse arkosic sands indicating some rejuvenation of older faults. Much clastic material of the platform consists of supermature sandstone probably derived mainly by recycling of pre-existing sandstone and quartzite. The shallow water carbonates that overlie the sandstones reflect subsequent downwarping of the platform accompanied by marine transgression possibly as far east as Baker Lake (Fig. 9).

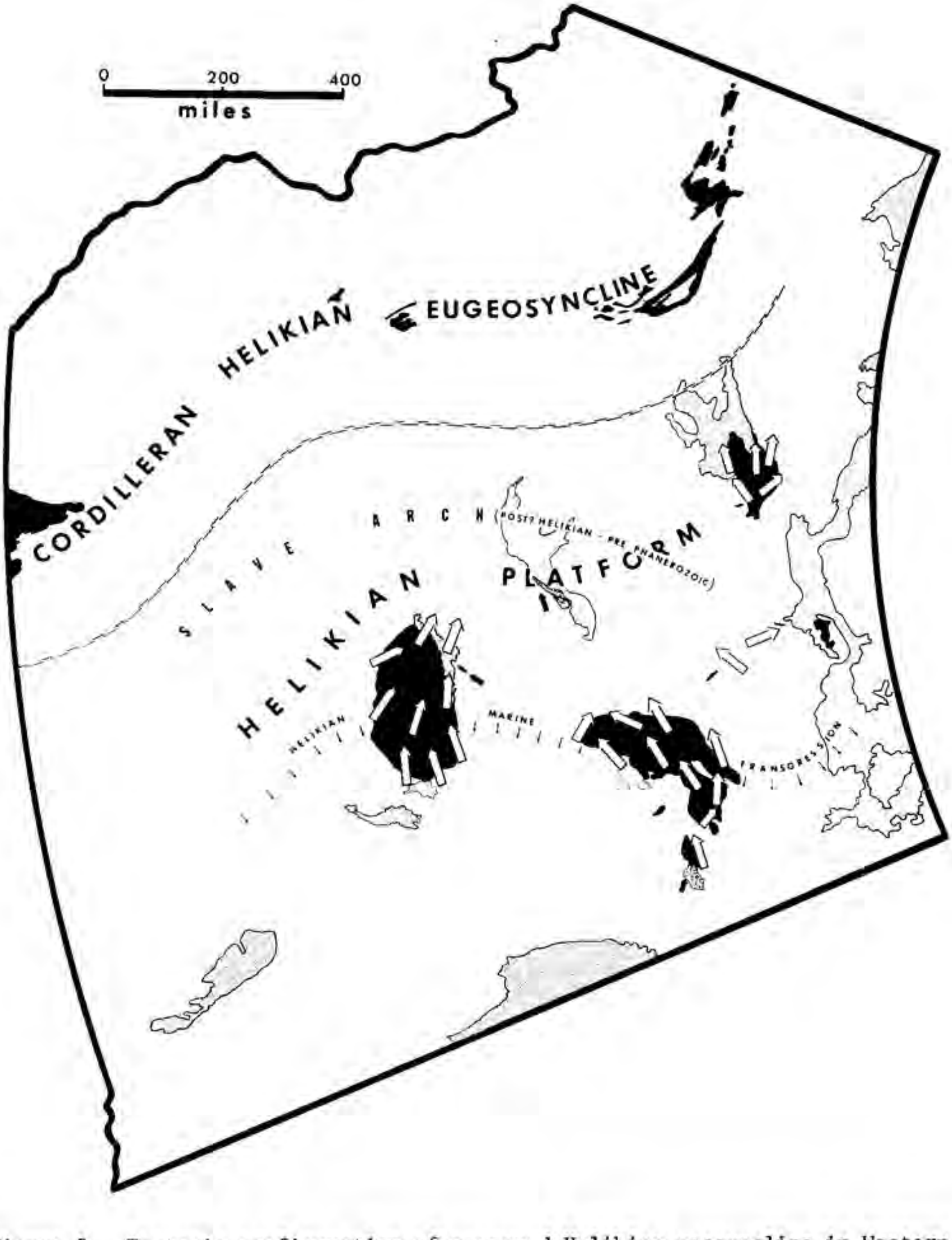

Figure 9. Tectonic configuration of proposed Helikian geosyncline in Western Canada. Solid arrows show directions of paleocurrents in redbeds and open arrows, paleocurrents in sandstone cover rocks. Paleocurrent trend for Et-Then Group, Great Slave Lake, is from Hoffman (1969).

To the west the clastic and carbonate sediments thicken and contain an increasingly greater amount of interbedded shale. More than 200 miles west of the Great Bear Lake basin, eugeosynclinal sediments of Helikian age exposed in the mobile belt of the Cordilleran Province delineate the axis of a Helikian ortho-geosyncline, constituting the deep water complement of the Helikian platform sediments of the Shield. The hinge-line that separates the eugeosynclinal sediments from the miogeosynclinal sediments of the platform probably lies near Great Bear Lake (Fig. 9). Lithologic variations coupled with an unconformity within the platform sequence at Great Bear Lake record uplift possibly associated with instability in the hinge zone.

Paleocurrent and stratigraphic data document a heavy flow of terrigenous material westward, but no Helikian strata are preserved along the Precambrian-Phanerozoic contact in the sector between Lake Athabasca and Great Bear Lake. This reveals the existence in this region of a broad arch (Fig. 9) here termed the 'Slave Arch' that formed during the Paleohelikian-Cambrian interval.

Development of the Helikian orthogeosyncline may reflect compression of the craton from the west and intrusion of the vast swarm of Mackenzie diabase dykes that fed the Coppermine River basalts may mark the ending of this phase of compression and the beginning of a new phase characterized by crustal extension. This was coupled with epeirogenic uplift and a resumption of continental sedimentation as shown by the association of redbeds with basalt in the Bathurst Inlet-Great Bear Lake region.

In the Coronation Gulf area, Helikian strata were mildly deformed, partly eroded, and covered by platform clastics and carbonates of the newly developing Hadrynian cycle of deposition. Hadrynian sediments are absent in other parts of the western Shield which may suggest that Hadrynian deposition was less widespread than that of the preceding cycle of deposition.

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THE CANADIAN SHIELD - A STATUS REPORT, 1970

F.J. Pettijohn,
The Johns Hopkins University,
Baltimore, Maryland, U.S.A.

INTRODUCTION

I want first to express my thanks for the invitation to attend this Conference and for the privilege of presenting a resumé and a commentary on the proceedings of the past two days.

The geology of the Canadian Shield has been of continuing interest to me since my student days at the University of Minnesota. My first field experience anywhere was as an assistant to Professor Frank Grout of the University of Minnesota who was then working on the Coutchiching problem in the Rainy Lake district on the international border. For these services I received the munificent sum of $75 per month from the Minnesota Geological Survey. It was my privilege also to study for a time with Professor A.C. Lawson, then at the University of California at Berkeley, who was on the other side of the Coutchiching controversy. Dr. Lawson was, as perhaps you know, one of Canada's most distinguished geologists - one-time president of the Geological Society of America and Penrose medalist. It was he, who in the years 1882-83, mapped the geology of Lake of the Woods (1885) - a formidable task for a youth of 22, for there was no base-map of this lake with its 2,000 square miles of water and 16,000 islands. In two seasons Lawson prepared a base-map and plotted the geology - a feat of no mean achievement if you recall this was all done under paddle and with no air support. It was from the Lake of the Woods study that Lawson proposed the term "Keewatin" which has since found its way into the literature on the Precambrian. The following season Lawson mapped the Rainy Lake area (1888), the type locality of the Coutchiching at Coutchiching Rapids on the Rainy River betwwen Minnesota and Ontario. Lawson's hand-written doctoral dissertation on the petrology of the Rainly Lake rocks is in the Hopkins library. You can see, therefore, what my ties to the Canadian Precambrian are and where my interest in its problem began.

I would like also to pay tribute to Dr. W.H. Collins, former Director of the Geological Survey of Canada - and also a past-president of the Geological Society of America - for his encouragement of my work in the Sioux Lookout area - work that was my doctoral dissertation at the University of Minnesota - work that followed Collins' own reconnaissance studies in this region, at the time of construction of the Canadian National Railway - then known as the Grand Trunk Pacific.

THE ARCHEAN

I mention this bit of personal history because last summer, in the company of Dr. Roger Walker of McMaster University, I revisited the Sioux Lookout area after an absence of nearly 40 years. I had not worked in Archean terranes since the thirties, having turned to the Animikie (Aphebian) sections of northern Michigan. The re-visit was made with some trepidation as I wondered whether what I had written as a graduate student about the Sioux Lookout area was, in fact, really so. What I found was that the rocks had not changed one iota but that our ideas about them had changed a great deal.

What was the nature of this change? Let's review how things stood 40 years ago. Lawson demonstrated, in his Lake of the Woods study, that the granites (Laurentian) heretofore presumed to be basement were in fact irruptive and hence younger than the greenstones with which they were in contact.

In the Rainy Lake area Lawson discovered a series of mica schists to which he gave the name Coutchiching, which were presumed to underlie the greenstones to which, in the Lake of the Woods area, he had applied the term Keewatin. Moreover he found a conglomerate formation (Seine series) which contains cobbles of both greenstone and granite and is therefore presumably younger than both but which was, in turn, cut by granite to which Lawson applied the term "Algoman" (Lawson, 1913). Thus we had two granites in the Archean, the older one (Laurentian) post-dating the greenstones and predating the conglomerates and thus recording an orogeny which divided the Archean. Another granite (Algoman) cut all Archean rocks and marked an orogeny which closed the Archean era. This is the way things stood 40 years ago. Where do we stand today?

Table 1. Lawson's Rainy Lake Stratigraphy (1913)

Algoman granites
intrusive contact
Seine series (called Timiskaming by most workers)
unconformity
Laurentian granites
intrusive contact
Keewatin greenstones
Coutchiching mica schists

The Laurentian and Coutchiching have vanished, the pebbles in the conglomerates have been down-graded, there is no unconformity of note within the Archean. Our whole stratigraphy seems to have collapsed.

The biggest difference, perhaps, in our interpretation of the Archean now and 40 years ago, is the concept of the Archean as an integrated volcanic-sedimentologic entity. The sediments are considered intravolcanic whereas 40 years ago there was a major unconformity in the sequence, recording an orogenic episode - the "Laurentian". Looking at Stockwell's subdivisions of Precambrian time (1964) one finds the Archean as a single time-stratigrahic unit closed by the Kenoran orogeny about 2,500 million years ago. No other orogenic event - no "Laurentian" - appears in the Stockwell's scheme. Is the Archean in fact indivisible? Or is it that the geochronologist has so far been unable to penetrate the 2,500 million year barrier and make effective separation of earlier orogenies? What is the status of this problem?

The Laurentian was supposed to intrude a predominantly volcanic greenstone assemblage and to be unconformably overlain by a predominantly sedimentary assemblage. Although conglomerates of the latter contain granitic cobbles, geologists had great difficulty finding the Laurentian *in situ*. Where the granites and the sediments were in contact, the former intruded the latter. Even Lawson had some difficulty discriminating between the older Laurentian and the younger Algoman granites. Doubt began to be expressed about the existence of the Laurentian (*see* Thomson, 1936, for example). Granites demonstrably older than the conglomerates were rare indeed if they existed at all. But what about the granite pebbles?

A careful petrographic study of the clasts in the Archean conglomerates in eastern Ontario and Quebec, led Bass (1961) to conclude that the Laurentian was non-existent, that the pebbles reported as "granite" by the field geologist were, in fact, felsites, mainly acid porphyries, or at best hypabyssal rocks probably derived from shallow intrusives penecontemporaneous with and related to the volcanism of the greenstone belts. No true granite, i.e. potassic granite is present. Hence the conglomerates were believed to record only local interludes of erosion, the sediments being in fact intravolcanic deposits, products of minor episodes of erosion and sedimentation. There is, therefore, neither a significant unconformity nor a period of granite magmatism and orogeny prior to the Kenoran (or "Algoman" to use Lawson's term). This view now seems to be a general one. Ayers (1969) presented a paper at Atlantic City in which he stated that the sediments of the Lake Superior Park area are nothing more than volcanoclastic (pyroclastic) deposits - products of the late acid phase of the major volcanism which, in its earlier stages, generated the basic eruptive lavas. Several papers at this Conference (McGlynn and Henderson and also Dimroth *et al.*) refer to the gradational boundary between the earlier volcanics and the later sediments without orogenic episode between them.

The view promulgated by Bass presumes little or no sialic crust, considers the Laurentian to be an outworn myth to be discarded. This may be so but the evidence which gave rise to the older concepts in the first place should not be so lightly dismissed. The conglomerate on the west shore and islands of Yellowknife Bay truncates the thick sequence of greenstone flows mapped by J.F. Henderson and Brown (1966, p. 31). The angular discordance between these flows and the conglomerate is considerable and can hardly record an insignificant interval of erosion. No less than 8,000 feet, and maybe much more, is cut out between Shot and Jackson Lake. Moreover some "granitic" debris is present in the conglomerate although it is much subordinate to the greenstone rubble despite its conspicuous nature. Re-examination of the conglomerates of Abram Lake, near Sioux Lookout, this past summer with Roger Walker, confirmed the abundance of "granitic" material in these conglomerates. Actual point-count (rather than pebble count) on some outcrops showed that over fifty per cent, by volume, of the clasts one centimetre or larger were tallied as "granite". Thin-section studies of these by Dr. George Fisher of Johns Hopkins showed them to range from leucocratic quartz diorite to granites, about half of which have been crushed and partially recrystallized under greenschist facies conditions. Modal analyses (Table 2) of the coarse meta-arkose associated with these conglomerates on Little Vermilion Lake showed 41 per cent detrital quartz and 27 per cent feldspar (which includes 2 per cent K-feldspar). Inasmuch as the "granites" had 32 per cent quartz and 63 per cent feldspar, the arkoses record a marked quartz enrichment relative to feldspar. Since the arkoses are over 5,000 feet thick and contain at least 40 per cent quartz, this is equivalent to a pure quartz sand 2,000 feet thick. I do not see this volume of quartz being derived from quartz porphyries (which in this region have 2 to 4 per cent quartz and 40 to 45 per cent feldspar) or being acid pyroclastics. The quartz must come from granitic rocks similar to the pebbles in the associated conglomerates. Consideration of the quartz budget problem in the North Spirit Lake area led Donaldson and Jackson (1965) to a similar conclusion.

The paucity of K-feldspar, in many cases its complete absence, from most Archean greywackes and the general sodic character of the "granites" is an intriguing problem. The greywackes of East Bay of Minnitaki Lake examined by me contain no K-feldspar. Those in the Yellowknife area are likewise devoid of K-feldspar and are characterized by an excess of Na_2O over K_2O (J.B. Henderson, personal communication, 1969). Parenthetically, it is of interest to note that the associated and interbedded slates have an excess of K_2O over Na_2O. But the scarcity or absence of K-feldspar in greywackes is not a uniquely Archean phenomenon. The greywackes of the Rove Formation (Aphebian) of northern Minnesota contain little or no K-feldspar (Morey, 1967). Of twenty samples analyzed, only two had K-feldspar and these had only 2 to 3 per cent whereas the plagioclase content generally ran 15 to 30 per cent. Likewise the feldspar in the Franciscan (Jurassic ?) of northern California is primarily sodic. Seventy-eight per cent of all samples examined by Bailey and Irwin (1959) showed no K-feldspar; of the remainder, 6 per cent had only a trace and the balance generally less than 5 per cent. I missed a discussion of this problem in any of the papers of the Conference. This is particularly pertinent when, as in the Yellowknife area, all the granitic bodies contain microcline. If some are indeed remobilized basement, as McGlynn and Henderson suggest, they must have undergone K-metasomatism.

Table 2. Modal Analyses Greywackes and "Granite" Cobbles

	East Bay, Minnitaki Lake, Ontario		Little Vermilion Lake Ontario	
	"Granite"	Greywacke	"Granite"	Greywacke
Matrix	—	45	—	26
Feldspar	57[1]	17	63[2]	27[3]
Quartz	30	25	32	41
Rock fragments	—	11	—	5
Other	13[4]	1	5[4]	—

1) 39 per cent plagioclase; 18 per cent alkali feldspar mainly albite

2) 48 per cent plagioclase; 15 per cent microcline

3) 25 per cent plagioclase; 2 per cent K-feldspar

4) Mainly chloritized biotite

If we can, then, entertain the notion of a Laurentian orogeny (as I think we must if we interpret the granite debris and the quartz budget rightly, Bass notwithstanding) then there were orogenies prior to the Kenoran that we are unable to resolve by present dating techniques. I might add, parenthetically, that I retain some skepticism about the reliance to be placed on the alleged dates. This skepticism arises from several well-known examples where ages have been assigned to pegmatite dikes, based on K/Ar ratios in muscovites, which make them older than the rocks they cut (Tilton *et al.*, 1959, p. 196). There is no denying the abundance of volcanogenic sediment interbedded with the greenstones but neither can one deny that sialic crust existed and was the principal source of some of the younger sediments which are separated by a significant unconformity from the subjacent volcanic complex.

McGlynn and Henderson are, perhaps, somewhat inconsistent in downgrading the unconformity in the Yellowknife area, believing "it does not necessarily require an orogeny" because, there are just " .. a few granitic boulders in the conglomerate". Yet the greywackes are considered, correctly I believe, to have been derived from a source of sialic composition. To expose such sources, covered in places by 30,000 or more feet of a volcanic assemblage, requires a major orogeny of some sort. This orogeny many not be quite what most of us think - severe Alpine-type folding, with large lateral translation and thrusting, and accompanying invasion of magmatic granites. It may be as Anhaeusser *et al.* (1969) suggest; namely, that the Archean granites we see today were the basement on which the thick greenstone assemblage was deposited and that these granites have risen primarily as crystalline solids, rising in diapiric fashion, penetrating and deforming the greenstones as they do so, in the fashion of salt domes. Outpouring of a thick volcanic sequence of the density of andesite or basalt (near 3.0) on to granites of lower density (near 2.65) creates a gravitational instability which permits the creation of a kind of convective system and diapiric rise of granite flowing as a very viscous, though crystalline solid, as does glacial ice or salt. This hypothesis explains the generally conformal boundaries between the granite plutons and the host rock, the fact that the plutons occupy anticlinal areas and that the greenstones "face away" from the granite, the minimal metamorphic effects (leaving most of the country rock in the greenschist facies) and the observed cataclasis in the border rocks of these plutons and in the pebbles of the conglomerates. Only locally are the granites fluid enough to display injection phenomena. This concept also explains the clustering or gregarious habit of the Archean plutons. This idea is not new. J.B. Thompson, Jr. suggested in 1950 that the gneiss domes in west-central New England, being of quartzo-feldspathic material were less dense than the basic Ordovician volcanics and hence may have risen and been emplaced in the same fashion as salt domes (Thompson *et al.*, 1968, p. 216).

It is interesting, therefore, that we now seem to have reverted to the earlier view and that the large granitic batholiths are basement. This hypothesis appears in the paper by McGlynn and Henderson who recognize the intrusive character of the grancdioritic bodies that flank the Archean basin but consider this to indicate a remobilization phenomenon and think that the extensive bordering granodiorites may indeed be basement and the source of the quartzo-feldspathic sands of the sedimentary sequence - a notion also expressed by Goodwin and Ridler in their discussion of the granitic massifs flanking the Abitibi volcanic belt. The notion that these bodies are the floor on which the volcanics were deposited brings us full circle in our thinking. In terms of the older terminology, the "Laurentian" is once more basement for the "Keewatin" instead of separating it from "Timiskaming". In my judgment this is a plausible hypothesis but we need more than plausibility to establish it. Definite proof of granite subjacent to and older than the greenstones is very meager indeed.

To sum up this discussion, we can say that the evidence seems to indicate a derivation of the quartzo-feldspathic sediments - as opposed to the volcanogenic sediments - from sialic source rocks. On this point most of the contributors to this Conference agree. Whether this source was a sialic basement on which the Archean greenstones ("Keewatin") were presumed to have been deposited or whether it was a truly intrusive granite ("Laurentian") associated with a post-volcanic and pre-sedimentary orogeny is less clear. Even in the latter case, there may have been no single Laurentian orogeny. There may indeed have been several, temporaly and geographically separated orogenic events. Perhaps the answer is more complex, as some contributors have suggested and the granitic debris did indeed come from a sialic basement which, however, now appears as piercement domes in the volcanic complex.

Some of the Archean problems that plagued us 40 years ago are still with us. Although the Archean has been defined as "...the basement complex... in which ordinary stratigraphic methods are not applicable" (Leith, Lund and Leith, 1935, p. 10). H.C. Cooke and Morley Wilson long ago dispelled the notion that the Archean would not yield to stratigraphic analysis. The utilization of air photographs, the construction of better base maps, large-scale mapping, close attention to internal structures in both sediments and lavas to determine stratigraphic sequence, have shown that it is possible to work out Archean stratigraphy at least locally. It remains difficult, however, to carry the stratigraphy across to adjacent areas separated by major faults and still more difficult to go from one greenstone belt to another across intervening granitic terrane.

One of the troublesome questions concerns the nature of the Archean basins. They contain a rock assemblage much like that of younger eugeosynclines - greenstones and greywackes as noted by Tyrrell in 1933 - the latter graded and interbedded with pelitic rocks. But were they basically different from these younger geosynclines - different for example from the trough in which the Franciscan of California, with its pillowed greenstones, graded greywackes, and bedded cherts accumulated? Many authors, the speaker included (Pettijohn, 1943a) considered the Archean basins to be eugeosynclines. Anhaeusser and his colleagues think the term "geosyncline" inappropriate for these basins and regard Archean volcanism and sedimentation as a thing apart from later times. Before we take this non-uniformitarian approach we should take a closer look. The map-patterns showing relations between the greenstones, volcaniclastic sediments, and the plutons in some parts of the Sierras are reminiscent of those of Archean terranes in the Shield, though the California example is late Paleozoic or Mesozoic in age (*see*, for example, the U.S. Geol. Surv. Folio 43, Turner, 1898). The Calvareras Formation, primarily lavas and pyroclastics, is over 15,000 feet thick, on top of which is the dominantly sedimentary Mariposa Formation, some additional 10,000 feet, some of which is also volcanogenic. It seems probable that this pattern of granite to host rock will develop anywhere and any time a thick sequence of denser lava rests on lighter sialic material. A kind of convective readjustment will take place, with rising diapiric granitic stocks and downsinking of greenstone basins. It does not depend on any hypothetical thinner crust and has been reproduced in the ingenious experiments of Ramberg (1967).

What model shall we use for the Archean sedimentary record? Various suggestions have been made. E.L. Bruce (1927) once thought that the Archean metasediments scattered over a large part of the Superior Province were parts of a vast delta - the "Coutchiching Delta". Inspired, perhaps, by Eskola's view that the graded beds in the Archean of Finland were varves (1932) a view later restated by Simonen and Kouvo (1951) and once invoked to explain the graded beds of the Sudbury series by Coleman (1926, p. 233), I interpreted the graded phyllites of Minnitaki Lake in the same manner. Re-examination by Walker and myself this year convinced us that these, like the graded sediments east of Yellowknife Bay described by Henderson, are turbidites comparable in all particulars to those found in younger flysch sequences. Even some of the associated conglomerates on East Bay of Minnitaki Lake are also turbidites similar to the Cretaceous conglomerates of Wheeler Gorge, California, which are interbedded with deep-water siltstones and shales (Rust, 1966).

But were the Archean basins typical flysch basins? Were the various now scattered and isolated Archean sediments deposited in one or many basins? McGlynn and Henderson advance the provocative idea that the present basins are very nearly of the same size and configuration as the original basins and that what we see is not the deeper infolded remnants or "keels" of mostly eroded structures. These belts are not the infrastructure of a vastly greater geosyncline. This view is somewhat akin to that presented some years ago in which "... the conglomerates are apparently wedge-shaped, basin-margin type of deposits and very probably mark, in a rough way, the original margins of the basins of sedimentation in which these various series accumulated" (Pettijohn, 1937, p. 191). These gravels were thought to be aprons of waste derived from the positive areas adjacent to the basins. The long, nearly continuous bands of conglomerate are not the courses of Archean rivers as Lawson once presumed (1913, p. 62) but are instead the marginal facies of Archean geosynclines.

McGlynn and Henderson regard the volcanics as a marginal facies - one that does not extend across the basin beneath the sedimentary fill (*see* their Fig. 2). Goodwin and Ridler believe the volcanics to be present throughout the basin and not simply a marginal development. How can we reconcile these views and decide which is the appropriate one?

The lineal or sublineal map-patterns and the related distribution of conglomerates in Ontario led me to delineate what seemed to be major sedimentary belts, 50 to 100 miles wide, separated by predominantly volcanic belts of comparable width. The sedimentary belts were given appropriate names and each could be traced several hundred miles (Pettijohn, 1937, 1943). Four such sedimentary belts were recognized in northwestern Ontario and interpreted as geosynclinal residuals. More recently Goodwin (1968a, 1968b) following somewhat the same line of thought, has extended this approach to a larger part of the Superior Province, has emphasized and named the volcanic belts, some six in all - separated by intervening sedimentary belts. It is one of these, the Abitibi greenstone belt, perhaps the best known, that Goodwin and Ridler have reported on at this Conference.

Whether this approach is valid or not and whether meaningful correlations between isolated greenstone relicts can be made is still uncertain. Even if so, we are still confronted with the problem of whether these various "geosynclinal" tracts existed and were filled concurrently or whether they were filled and deformed in sequence.

As noted by many, the Archean basins seem not to show the asymmetry believed to be characteristic of true geosynclines, do not show a sequence of filling found in Alpine-type geosynclines, have an absence of the more mature sands and carbonate rocks. The lack of asymmetry is not due to the apparent intracontinental rather than continent-margin character of the Archean troughs, as the Ural geosyncline, also intracontinental, is markedly asymmetrical. It may well be, however, as Anhaeusser and his associates suggest, that the Alpine model is inappropriate for Archean basins, a conclusion also reached by Dimroth and his colleagues. This does not mean that the Archean is exceptional as Dimroth implies but rather that we have the wrong model. As suggested earlier we need a model that will provide the complementary piercement granite plutons and greenstone basins. It should be pointed out, however, that the dome and basin tectonics is not the whole story. This pattern is superimposed on a lineal or sublineal control as noted above.

I might add, in concluding my remarks on the Archean, that if the hypothesis which seems currently gaining ground; namely, that there are two kinds of granites - three kinds according to Anhaeusser - the remobilized basement emplaced by piercement salt-dome tectonics and the pegmatitic and magmatic granites which cut and granitize the sediments, this concept may have some practical implications. The former are "dry" bodies with minimal fluids and no superheat producing limited metamorphic effects on the host rocks; the latter are "wet" granites generating a host of pegmatites with their rare elements. The mineralization associated with the two may be significantly different.

THE PROTEROZOIC

If the Alpine model is inappropriate for the Archean, the Conference seems to have demonstrated that it is a viable model for much of the Proterozoic of the Shield. This is most dramatically shown for the Coronation geosyncline and the associated sediments of the East Arm of Great Slave Lake, the Epworth and the Goulburn areas by Hoffman, Fraser and McGlynn. This geosyncline displays the characteristic elements of the Alpine model (Aubouin, 1965): the cratonic foreland, the asymmetrical fill and, by inference, the tectonic source-land. The former provides only a small part of the fill - mainly mature cratogenic sediments; the latter furnishes the thick clastic wedges of immature sediments - greywacke turbidites of the flysch facies in the earlier stages and red clastic molasse in the final stages. Of special interest are the thick stromatolitic carbonates on the stable cratonic shelf - co-existing in part with the orogenic sediments in the geosyncline itself. It is possible that the Coronation geosyncline is the precursor of the Cordilleran geosyncline of Paleozoic and later times. The zone of the geosynclinal downwarp and sedimentation may have shifted westward during Paleozoic times.

The Labrador trough is a well-organized geosyncline with many similarities to the Alpine model as Dimroth and his colleagues point out. As in the Appalachian geosyncline, there are marked facies changes across the trough, in an east-west direction in central Labrador, with sediment input mainly from the west or cratonic side during the early stages and from the east or orogenic belt during the later history. Like the New England Appalachians, the western part of the trough is of miogeosynclinal aspect, characterized by cleaner quartzites and carbonates whereas the eastern part is marked by argillaceous and coarser, ill-sorted sediments and by a thick sequence of volcanics. The formation of a medial geanticline dividing the trough is matched by a similar structure in New England.

The Labrador trough displays a cyclic arrangement of deposits - some two major cycles with a possible third. In this respect the Labrador trough is also like the central Appalachians (Meckel, 1970, pp. 49-67) in which the Paleozoic is comprised of two major cycles and an abortive third. The Appalachian cycles begin with orthoquartzitic sands and carbonates which are followed by an euxinic phase, a flysch and a molasse. In the Labrador trough, the carbonate-orthoquartzitic phase is present, as is an euxinic stage - the graphitic slates (Hautes Chutes and Ruth slate), which is followed by greywackes and slates, which if not typical flysch, have some flysch-like affiliations.

Another aspect to be kept in mind is that the sequence need not be the same throughout the length of the geosyncline. The centers of maximum accumulation shift from time to time. The center of maximum accumulation in Cambrian times in the Appalachians was in Virginia, the Devonian maximum in southeastern Pennsylvania, and the Pennsylvanian maximum in Alabama. The shifting depo-centers are reflected in significant changes in thickness and character of the deposits. There were also dramatic shifts in the loci of maximum deposition in the Labrador trough. The lowest sequence (Cycle I) is absent in the north whereas the second sequence (Cycle II) is thin in the south and very thick in the north. The complexity of geosynclinal filling is similar to that in the Penokean fold belt which extends across Minnesota, Wisconsin, and Michigan (James, 1958). Iron-bearing formations appear at four stratigraphic levels, major greenstone volcanics at two levels, all within the Animikie series (Supergroup). The sequences in the several iron-mining districts are markedly different from one another in many respects. The Labrador sequence includes two iron-bearing formations and three of volcanism.

Unlike the Labrador and Appalachian geosynclines and also unlike the Coronation geosyncline, the Huronian trough seems to have received all its fill from the cratonic (north) side. It has the wrong polarity and it does not, therefore, fit the Alpine model (Aubouin, 1965). This cratonic provenance and absence of sediments from a tectonic source-land probably explains the character of Huronian sedimentation - the abundance of cratogenic sediments, primarily thick quartzites and feldspathic quartzites, and the absence of orogenic and post-orogenic sediments - the greywackes and lithic arenites which characterize flysch and molasse. Nor does the Huronian have the characteristic volcanics of the Alpine geosyncline - the ophiolites.

The problem of the nature of the Huronian basin cannot be fully analyzed until the relations between the Huronian and Sudbury Series is clarified. This has been one of the long-standing problems of the North Shore-Sudbury area. Many, including Frarey and Roscoe, regard the Sudbury series - renamed and redefined as the Elliot Lake Group - as the lowest part of the Huronian. Whether a part of the Huronian or not, it records a volcano-sedimentary history unlike the remainder of the Huronian. It is a eugeosynclinal assemblage with a great thickness of regularly graded sandstones (McKim) associated with flows and pyroclastics (Stobie and Copper Cliff). The McKim sediments thicken greatly southward (Roscoe, 1968, p. 49). In contrast, the Huronian proper contains limestones, many clean quartzites and some terrestrial, including glacial, shallow-water sediments lacking graded bedding, dark colour and showing instead ripple marks and large-scale crossbedding. The contrast between the tectonic and depositional environment of the Sudbury and the Huronian could not be greater.

The only other volcano-sedimentary sequence in the region is the Whitewater series - a series which is as difficult to place in the stratigraphy as the Sudbury series, having been considered Animikian by some, Keweenawan by others. The Chelmsford sandstone has more sedimentologic affinities with the McKim Formation than any other. The Chelmsford too is basically a turbidite deposits consisting of interbedded graded greywackes and dark slates. The only significant bodies of argillaceous sediment in the whole North Shore-Sudbury area are the McKim and the Onwatin-Chelmsford Formations. That the Whitewater series might be the time equivalent of the Sudbury series is not a new idea. Thomson (1957, 1961) and Williams (1957) both come to the same conclusion. If Thomson's identification of the greywackes at the south end of Wanapitei Lake as McKim is correct, then the Sudbury group is most certainly pre-Huronian, as the angular unconformity at this locality between these greywackes and the overlying Mississagi quartzite is as profound as any in the whole of the Shield. The Chelmsford and the McKim were deposited by turbidity currents in relatively deep water, associated with volcano-clastic sediments and eruptive flows. The Chelmsford, however, is not a volcanic sediment, it had a granitic provenance (Williams, 1957, p. 83). If the Chelmsford is not the equivalent of the McKim, it occurrence is anomalous and wholly inexplicable. No hypothesis to explain the Sudbury Basin can be considered adequate unless it accounts for the turbidites of the Chelmsford. Here is an opportunity for a good sedimentological study.

To me, one of the unique features of the contribution by Frarey and Roscoe was the recognition of three major regressive cycles represented by the Hough Lake, Quirke Lake, and Cobalt groups - each beginning with a tillite (a dropstone tillite in fact), followed by fine-textured marine beds, including limestone, and ending with current-bedded fluvial deposits. Unfortunately the authors do not give the evidence for the fluvial interpretation of the uppermost member of each cycle. The formations in question do not show the usual attributes of fluvial beds, namely the fining-upward cycles (Visher, 1965, p. 46) which characterize such well-known fluvial deposits as the Old Red Sandstone of Great Britain (Allen, 1964), the Catskill (Allen and Friend, 1968) of the Appalachians, or the Preble Formation of the East Arm of Great Slave Lake (Hoffman, 1969, p. 455). These fluvial cycles begin with a channeled floor and basal conglomerate containing mudstone intraclasts, which passes upward into crossbedded sands, thence into rippled silts and finally mudstones - each such fining-upward cycle being 10 to 100 feet in thickness. It has been argued that in the absence of land plants the alluvial deposits would be non-cyclic sheets of coarse immature materials (Schumm, 1968, p. 1583). The characteristic fining-upward cycles in the Preble Formation of the East Arm of Great Slave Lake seem to contradict this thesis.

It may be, of course, that the quartzites were not deposited by meandering rivers which produce the fining-upward cycles. Another fluvial model, that of a braided stream, may be more appropriate. I do not think so, for in any case the more mature sands of the Lorrain and the Bar River Formations differ from fluvial sands in their rounding and composition - rivers not being capable of producing either the roundness observed or the high concentration of quartz. The persistent, but thin banks of jasper-bearing gravel found in the Lorrain are more reminiscent of beaches than of river channels. A further observation indicating a marine rather than fluvial environment are the limestones which occur not only in the lower part of the regressive sequence but also in the upper part as in the Serpent and Gordon Lake Formations. In short, the Huronian Groups (excluding the Elliot Lake Group which may not be Huronian at all) seem each to record a regressive marine cycle which only rarely, if ever, becomes an emergent surface of accumulation.

One of the unique features of the Huronian are the tilloid deposits contained in the Ramsay Lake, Bruce and Gowganda Formations. Despite contrary opinions of some, I hold these to be glacial - in part glacio-marine. Their great lateral extent is inexplicable under any other hypothesis. The rafted clasts in the associated varved argillites, so well exposed on the Chapleau road near the Mississagi River crossing, is irrefutable evidence of ice-action. Glaciation records a climatic event of great and probably widespread significance. It should provide a basis for long-range correlation. Tillites outside the original Huronian are scarce and questionable. They do occur in

Michigan where they lie disconformably below the Animikian iron-bearing sequence of that area (Pettijohn, 1943b; Puffett, 1969). It has been suggested that the Fern Creek tillite is of Cobalt age (Young, 1966); others (Frarey, 1966) disagree. While other glacial epochs may have occurred, the only known example is Huronian. Hence I think the presumed correlation made by Young is probable.

Just a word about the so-called anoxic character of the Huronian. Attention has been called to the absence of red colour in the pre-Cobalt Huronian (Roscoe, 1968, p. 80) and to the occurrence of detrital pyrite in the uraniferous conglomerates of the Elliot Lake area. It is suggested that these can be best explained if the Precambrian atmosphere were oxygen-free. It has also been suggested that the time of appearance of free oxygen would be a world-wide event which should affect the geologic record everywhere and thus provide a datum plane for correlation. This concept has, in fact, been utilized by Bell in his report on the Hurwitz Group. The absence of red colour below the Cobalt Group, however, has no more meaning than the absence of red sediments in the Paleozoic of the central Appalachians below the Silurian (or actually below the late Ordovician Juniata). It means only that conditions for redbed development were lacking - the prevalence of marine rather than terrestrial conditions - and not that the atmosphere was anoxic. One does not look for red sediments in a eugeosynclinal sequence (Sudbury) or shallow marine sequence (Bruce Group sensu stricto). The meaning of detrital pyrite is not very clear. The pyrite of mine tailings is known to travel many miles downstream.

The soil profile devoid of a zone of iron accumulation, which is developed on the Archean basement beneath the Huronian is a more compelling argument for an oxygen-deficient atmosphere. It does not, however, constitute proof. There are modern gley soils which form under conditions of poor drainage and water-logging in which there is no horizon of iron enrichment. Frarey and Roscoe exclude organic material as an agent for iron-reduction and removal. The argument could be turned around and the absence of iron accumulation cited as evidence that organic matter was present - not that the atmosphere was reducing.

In summary, the Huronian does not bear much resemblance to the classical geosynclinal fill. It lacks the usual flysch (the Sudbury Series being excluded) and molasse. Except for the Gowganda pebble argillites it lacks any significant body of pelitic sediment. This is surprising in view of the vast quantity of sand, some of it being clean quartz, implying large-scale weathering with concommittant production of clay. Where are the missing shales? Even the limestones are atypical - not being the usual stromatolitic type which characterizes much of the Precambrian. Most of the Huronian, at least that north of the Murray fault zone, was deposited on a surface of low relief on a relatively stable area (the evidence of high relief is not particularly convincing - Frarey's Wawekobi Lake sheet (1961) shows the Gowganda resting on the thin Bruce limestone for 8 miles, his Bruce Mines sheet (1962) shows the Gowganda resting on a uniformly thick Serpent for 9 miles. These observations are consistent with Lindsey's observations in the Whitefish Falls and Timagami areas (1967).

Clearly not all of the Proterozoic of the Shield fits the Alpine model. Not included in our conference program, for example, is the Keweenawan of the Lake Superior region. Although only a small part is Canadian, it constitutes a major Precambrian sequence and the gravity anomaly associated with it is one of the most marked in North America (Woollard, 1964). The Keweenawan forms a mass of continental red clastics and basic amygdaloidal lavas over 50,000 feet thick. These materials fill a trough nearly a hundred miles wide and perhaps a thousand miles long. The sequence is exposed on the shores of Lake Superior, especially on the Keweenaw Peninsula from which it takes its name. Its southwestward extension is concealed by the Paleozoic and younger rocks but is clearly traceable, as the most prominent gravity high in North

America, southward as far as Kansas. The Keweenawan trough is not a geosyncline in the ordinary sense. The strata are tilted in places, bounded by normal faults. The structure is a rift-valley or a graben or half-graben feature, like the smaller Newark basins with their Triassic fill in the eastern seaboard region. There is no folding, no thrusting, no metamorphism and no granitic intrusion, only injection of diabase sills. We are dealing here with a major, late-Precambrian (Helikian) rift-valley fill with marginal fanglomerates, mainly alluvial materials and, because of blocked drainage, some lake deposits. Marine deposits are absent.

Clearly the Athabasca, the Dubawnt Group and the sediments of the Copper Mine River area are not geosynclinal in the usual sense. They are large residuals of once more extensive, mainly continental sediments. The basins of accumulation were relatively shallow, deformation is restricted to faulting; magmatism to lava flows. These deposits have more affinities with the Keweenawan (even to the copper mineralization) than to the Aphebian geosynclines.

Most troublesome to understand and most difficult to unravel are the sediments of the so-called "mobile belts" (Anhaeusser *et al.*, 1969, p. 2194). These are rather wide belts, a hundred or more miles in width, composed of highly metamorphosed sediments and volcanics - now high-grade schists, granulites, and orthogneisses. They have been interpreted as a series of geosynclines which have undergone deep burial, orogenesis, and subsequent erosion to reveal the root zones of the orogens accompanied by widespread granitization. I refer to the Churchill and the Grenville mobile zones. To what extent these mobile belts constitute reworked cratonic materials (granites and greenstone belts) as against infolded supracrustal Proterozoic formations and younger granite is as yet undetermined. As shown by Wynne-Edwards (1964) and others (Baer, 1969) the Grenville belt contains rocks of a diversity of ages including recognizable Archean elements and highly metamorphosed Aphebian rocks of the Labrador trough. Similarly Money and associates in their paper on the Wollaston Lake belt show a presumed mix of older and younger rocks. In a sequence like this, where one has to use bulk chemical composition as a clue to what the original rock was, meaningful analysis of the sedimentary basin is most difficult and results least satisfactory. Where one cannot even be certain of either the geologic age or the original character of the rock, one may be forgiven for being skeptical of the total synthesis.

The Hurwitz Group described by Bell presents a mixed picture. Of special interest are the tilloid conglomerates in the lower part of this group (Padlei Formation), perhaps correlative with similar beds in the Huronian, and the prominence of thick clean quartzites in the middle of the group (the Kinga Formation). One unit (the Whiterock Lake Member) displays ripple marks that have a strikingly uniform orientation over a large region. Bell applies the term "geosynclinal" to the succeeding strata but these are really not very thick, not even much thicker than those designated "platform". They are not true "orogenic sediments", such as the flysch, but are, instead, sediments suggesting restriction of the environment - cherts, black slates, iron formation - and volcanics. The uppermost beds are classed as "molasse". It is not clear from Bell's description whether or not these beds contain the conglomerates and immature sands arranged in fining-upward alluvial cycles which characterize molasse. In summary, it does not seem to me that the terms "geosynclinal" or "molasse" are particularly appropriate and it seems that the Hurwitz Group, like the Huronian, to which it bears some resemblance, does not fit the Alpine-type geosyncline with its flysch and molasse.

As is typical of the rocks of the mobile belts, those of the Wollaston Lake area are a group of rather highly metamorphosed plutonic and supracrustal rocks - the latter including both sediments and volcanics, in some cases so much altered that the original nature is in doubt. In such cases one has to rely on bulk chemical analyses to resolve these doubts. This is obviously an unsatisfactory situation if one wants to reconstruct the environment of deposition and the sedimentological history of the region.

It might be worth noting that the undoubted Aphebian or Helikian rocks, which are found at various places in the Shield all show, at one place or another, a little-deformed phase which lies unconformably on a cratonic block. The Animikie (Aphebian) of the Penokean fold belt extends in little-deformed manner into the Animikie district of Ontario; the Huronian of the North Shore overlaps the craton to form an extensive little-deformed sheet in the Cobalt-Timagami region; the sediments of the Coronation fold belt are found in undeformed platform facies in the eastern part of the Epworth and East Arm areas and in the Goulburn area. At no place are Archean sediments found as undeformed "platform facies" unconformable on older sediments. The same is true of the schists of the Churchill and Grenville fold belts which makes one suggest that most or all of these might be in fact Archean.

RESUMÉ

In summary, we have seen that the Precambrian sedimentary basins display four basic structural styles: (1) the Archean pseudo-eugeosyncline with its flysch-like sediments and volcanics, (2) the well-organized Aphebian Proterozoic geosynclines of Alpine affinities, (3) the shallow to deep half-graben structures, mainly Neohelikian continental sediments and lavas, such as the Keweenawan, and (4) the enigmatic mobile belts with a mix of rocks of diverse ages, now high-grade, partially granitized metamorphics. It is apparent from our previous discussion that the Huronian fits none of these categories. It has the wrong polarity and lacks the characteristic fill of the Alpine model, lacks the red clastic alluvial fill and volcanism of the rift-valley and fault-basin model, and is unlike the Archean in all respects. To what model is it related? Were it not for the folding and great thickness south of the Murray Fault, it could be considered an intracratonic basin - an "autogeosyncline" of Kay. Clearly the question cannot be answered without resolving the problem of the McKim and the Stobie volcanics.

THE FUTURE

We ought perhaps to look ahead and ask ourselves "what next"? I shall attempt to play the role of soothsayer knowing that the life of a practitioner of the world's second oldest profession is a risky one. I shall, however, be audacious and suggest areas or topics which might be profitably explored.

In the last two or three decades our knowledge of the Shield has increased at an accelerated rate due mainly to interest in the far north. Barren ground, air travel, and air support have led to rapid filling-in of the blank spaces on the map. We know pretty well the answer to the question "what is there". And yet if we define the Shield as that stable block of Precambrian rocks little deformed in Cambrian or later times, most of the Shield is unknown, for most of it is concealed beneath the younger rocks of the continental interior. The area of the exposed Shield is roughly 3 million square miles. The area under cover of 5,000 or less feet is almost as large. We might presume that the area covered is as complex and diverse as that exposed. If so, our knowledge of it is precious little, being what has been gleaned from a few small exposed bits and a few deep borings. Can we ever unravel the geology of this vast area?

What happens to the Aphebian and younger fold-belts when they pass under the cover of younger rocks? The Coronation fold-belt is exposed only in a relatively small area in the Bear Province, from which place it passes southward under a cover of Paleozoic and younger rocks. The iron-bearing districts of Minnesota extend westward under a cover of Cretaceous rocks. Is there somewhere concealed another Mesabi Range? another Labrador trough? Iron is now being mined at a depth of 4,000 feet in Wisconsin. Would not a concealed Mesabi be equally profitable?

We have seen how the Keweenawan trough is traceable hundreds of miles southward. Geophysics and deep drilling is the clue to the exploration of the covered Shield. But such data can only be interpreted if we have some idea of how the Shield is organized and put together.

There is, of course, room for a vast amount of field and laboratory work on the rocks of the exposed portion of the Shield. The usefulness of paleocurrent analysis in furthering our understanding of the filling of sedimentary basins has been amply demonstrated by several papers at this meeting. Whether the sediment came from the stable basement or from an orogenic uplift or mid-geosynclinal ridge is an important problem. But paleocurrents alone are not enough. We need to look at the sedimentary rocks themselves. They deserve the same scrutiny that we conventionally reserve for the igneous and metamorphic rocks. Petrographic analysis is made not so much to provide a more detailed description or even to classify or name the rock in question as to unravel the geologic history. The sandstones are especially useful in attaining this goal. Petrography is not particularly helpful in environmental analysis but it is a powerful tool in working out, in conjunction with stratigraphy and paleocurrent data, the provenance of the sediment, that is, the kinds of location of the source rocks and the climate and relief of the source area. We need also to scrutinize more carefully the small-scale stratigraphy - to see how the various lithologies are organized or put together - whether they show graded bedding and turbidite characteristics or show the fining-upward cycles characteristic of the alluvial environment. Foot by foot measurement of Archean sections are almost unheard of; comparable measured sections of Aphebian and later rocks are rare. I commend the examples published in GSC Paper 68-42 on the East Arm rocks of Great Slave Lake by Paul Hoffman.

I have earlier suggested that the kinds of effort represented at this Conference may not be without some economic significance. There has been a recognition of the stratigraphic control of ore deposits. The uranium-bearing conglomerates of Elliot Lake, the gold of the Rand and the South African copper deposits come to mind. A striking example of the stratigraphic control of heavy metal deposits has recently been described by Crowley (1968, p. 74) in the metamorphic terrane of Connecticut. Goodwin and Ridler show that even in the Abitibi orogenic belt there seems to be some relation between stratigraphy and metalliferous deposits. Whether certain stratigraphic horizons are just favorable host rocks or whether there is a genetic connection between the deposit and the host-rock is not clear. Jackson and Beales (1967) in a provocative paper on the lead-zinc deposits of Pine Point on Great Slave Lake suggest that the mineralization of this type of deposit is related to the escape of metal-bearing brines from the sedimentary basin. If such a principle operates on a Phanerozoic basin might not it also apply to a Precambrian basin?

If the concept mentioned earlier of the "wet" and "dry" granites be correct, the mineral deposits associated with the wet exhalations would be related to Archean structure and petrology. We have a long way to go to test these speculations.

EPILOGUE

Let me say in closing that I think this Conference is symptomatic of a maturing of our work on the Precambrian - particularly of the Shield. As the answer to the question "What's there" is attained we turn more to analysis of how things are put together - how the Shield as a whole is organized. This seems to be a banner year for meetings to explore these questions. There are three Canadian conferences: Ottawa, London, and Sudbury - this year and one in the States still in the formative stages. In my judgement the stage is set for organizing a group which will have some continuity, that will address itself to all the problems - structural, sedimentological, magmatic, economic - that relate to the Precambrian - a new society: "Friends of the Precambrian".

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Comments by: S.M. Roscoe and M.J. Frarey

Professor Pettijohn's contribution to the workshop, entitled "The Canadian Shield - A Status Report 1970", includes a number of criticisms of the paper by Frarey and Roscoe on "The Huronian Supergroup North of Lake Huron". He expresses serious reservations concerning some of the correlations accepted by us, he doubts that non-marine sediments are important if present at all in the Huronian succession, and he questions the validity of evidence that a non-oxidizing atmosphere prevailed during deposition of the Lower Huronian Groups. These questions are interrelated in part and are fundamental considerations with respect to the conceptual model that we have proposed for the deposition of the Huronian succession. We want to make it clear that all of these points were carefully considered before we developed our hypothesis, although they obviously could not be treated comprehensively in a summary paper.

The old problem of relationships between those Huronian rocks that include the Ramsay Lake, Mississagi and younger formations and pre-Ramsay Lake strata, formerly termed Sudbury series by some, was not discussed in our paper. Pettijohn's impression, that we dismissed this problem by renaming the latter strata the Elliot Lake Group and arbitrarily classifying this group as the lowest part of the Huronian Supergroup is unfortunate. The name Elliot Lake Group was first applied to undoubted Huronian rocks at Elliot Lake (Roscoe, 1957). Comparisons of stratigraphic sections (Ginn, 1960; Roscoe, 1960) soon made it apparent that the term might be applied further east to include argillaceous rocks that had been mapped as McKim Formation and underlying quartzite which contains radioactive, pyritic quartz pebble conglomerates. These rocks had been considered pre-Huronian (Sudbury series) by some, but not all workers (e.g. Cooke, 1946; Fairbairn, 1944). Subsequently, systematic detailed mapping was completed between the Elliot Lake area and the Sudbury area (Card, 1965; Robertson, 1966) and the relationships were established beyond any reasonable doubt. This has been one of the most important results of the extensive geological work that has been done in the region in the past ten to twenty years.

It has also been established during this period that there are volcanic rocks in the lowermost part of the Huronian succession, not only west of Elliot Lake (Frarey, 1959; Hay, 1961) but also east of Elliot Lake (Robertson, 1966). The latter include some volcanics that had been considered part of the Sudbury series or part of the Archean basement complex by some writers. Roscoe (1969) and Frarey and Roscoe (1970, this workshop) chose to classify all of the volcanics, including the Stobie Formation, which underlie the McKim and Matinenda Formations with apparent conformity as Huronian, in the absence of evidence to the contrary. This may be open to question but it is not a critical factor in our model for the deposition of the Huronian Supergroup.

Perhaps one of the reasons that Pettijohn is loath to accept the correlations that we and other recent workers accept (Robertson, Card, and Frarey, 1969) is because he thinks of the old Sudbury series as "a eugeosynclinal assemblage with a great thickness of regularly graded sandstones (McKim associated with flows and pyroclastics (Stobie and Copper Cliff)". The Stobie volcanics may be basalts comparable to the Thessalon volcanics which are interlayered with Huronian sediments west of Elliot Lake, rather than volcanics with eugeosynclinal affinities. The Copper Cliff Formation is so different from most pyroclastic formations in Archean belts that Phemister (1956) considered it intrusive rather than effusive or pyroclastic. Sorted sediments like those in the Huronian succession, crossbedded quartzite, argillite, quartz pebble conglomerate - rather than sediments like those in Archean belts - greywacke, unsorted volcanic conglomerates, iron formation - are interlayered with the volcanics.

The original character of the McKim Formation near Sudbury is somewhat obscured by metamorphism but most of the sediments do not appear to be turbidites, high rank greywackes, or to differ markedly from other fine-grained argillaceous units in the Huronian succession. Individual layers in the least metamorphosed sections can be described as argillaceous quartzite, siltstone and argillite. The rocks have an intact, sorted fabric rather than the disrupted fabric that some authorities consider diagnostic of greywackes. Clasts are quartz and subordinate feldspar obviously derived from a weathered granitic provenance. No lithic clasts have been reported. Ripple marks and crossbedding are present. All of these features, and the conspicuous small scale layering, or lamination found in some outcrops, are duplicated in the Pecors, Espanola, Gowganda and Gordon Lake Formations. The feldspathic quartzite beneath the McKim Formation in the Espanola area is crossbedded, contains pyritic quartz pebble conglomerate beds and is similar to the Matinenda Formation at Elliot Lake and not very different from parts of higher Huronian quartzite units - Mississagi, Serpent, and Lorrain Formations. There is no justification for Pettijohn's belief that "The contrast between the tectonic and depositional environment of the Sudbury series and Huronian could not have been greater".

The term Sudbury series, much less the problem of relationships between this mythical series and Huronian rocks, should not be resurrected unless new evidence is uncovered to show that conclusions drawn from recent mapping are fallacious and that there is a Huronian-like succession in the Sudbury area which is not correlative with any of the Huronian rocks of the Elliot Lake area.

Pettijohn discusses the possibility that the Whitewater series may be equivalent to rocks that he calls Sudbury series and that we refer to as Elliot Lake Group. We see no reason to depart from the classical concept that the Whitewater rocks are the youngest Precambrian strata in the area and were laid down after older strata including the Huronian rocks were folded, intruded, and metamorphosed. Some structural and geochronological evidence for this was reviewed recently by Roscoe (1969). The Onaping breccia is unlike any volcanic or other formations in Archean or Proterozoic rocks in the region. The overlying strata differ from the McKim Formation and sediments in Archean rocks in many respects - e.g. their slight metamorphism, their highly carbonaceous character, the presence of algal structures and the character of associated mineral deposits.

Evidence for fluvial origins of various Huronian arenites has been presented in a number of substantial publications - e.g. McDowell (1957), Pienaar (1958, 1963). As noted by Pettijohn, these conclusions were not reviewed in our paper. It did not seem essential to do so because the evidence for fluvial origin had not been seriously disputed, nor had contrary evidence, that the arenites were deposited in the sea, been presented. The following points may be useful.

The formations contain crossbedding of both the trough type and the steeply-dipping planar type. These have strongly preferential dip orientations which indicate deposition in water that flowed swiftly in a southerly direction. K.D. Card, commenting on our paper, stated that paleocurrent data indicate marine deposition in the south as opposed to non-marine deposition in the north. This requires documentation. Definitive southward changes in the character of crossbedding within any given lithostratigraphic unit have not been reported. If the suggested changes include a decrease in dominance of the preferred southerly directions of crossbed dips, with increases in frequencies of aberrant dips or with the appearance of secondary modes, this could be due to changes from upland intermittent or braided stream environments to downstream flood plain environments. There is no evidence that crossbedded sands were transported by reversing currents parallel or normal to shorelines as might be the case along beaches and offshore bars and aeolian crossbedding indicative of seashore dunes has not been reported.

Most of the arenaceous sediments have grain size distributions indicative of fluvial rather than littoral deposition. Interstices between sorted granules of quartz and feldspar contain smaller clasts, silt and argillaceous material. The matrix of interlayered conglomeratic beds containing quartz, chert and jasper pebbles consists of arenaceous material similar to that which comprises non-conglomeratic beds. Authigenic mineral cements, quartz or carbonate, are extremely rare.

It is doubtful that the maturity of quartzites and the sphericity of quartz clasts provide any indication of the distance of transport or the depositional environment of Huronian sediments, as suggested by both Card and Pettijohn. Angular and subangular quartz granules are found in some very poorly sorted, gritty, argillaceous feldspathic quartzites, but rounded quartz grains are common in all of the arenaceous formations including sands overlying sub-Huronian paleosols. Even these paleosols themselves do not lack spheroidal quartz grains whose shapes are partly inherited from the anhedral shapes of quartz in igneous rocks and partly developed, perhaps, through corrosion of sharp edges during weathering. In the upper part of such soil profiles, quartz grains are scattered in an argillaceous matrix and it is evident that erosion and sorting of this material would have produced two types of supermature sediments, argillite and orthoquartzite, regardless of the distance of transport, the depositional environment, and the amount of mechanical attrition. Quartz grains in relatively immature, highly feldspathic, Serpent quartzite are as rounded as those in many mature to supermature quartzose quartzites in the Cobalt Group. Arenites most likely to have been deposited under marine and marginal marine conditions are those that are found within or adjacent to silty, argillaceous, or limy formations, such as the Pecors, Espanola and parts of the Serpent Formation. These are far from mineralogically mature quartzites; in fact they are amongst the most immature of Huronian arenites. Pettijohn accepts that the Athabaska and Dubawnt sandstones are non-marine, yet these are generally finer grained and more mature mineralogically and texturally than most Huronian arenites.

The concept stated by Card that the northern part of the Huronian sequence was deposited in shallow epicontinental seas and the southern parts in deeper water within a fault-bordered basin seems a rather static one. It can hardly help explain the many changes in lithology that occur vertically through the succession and the comparatively slight (although definite) lateral variations found within extensive formations. Surely argillites and crossbedded quartzites in the Pananche area, for example, were not both deposited in similar marine settings, similar depths of sea water and at similar great distances seaward. Some variations within individual formations, including zones of rapid thickening, were probably controlled by recurrent movements along important faults, as stated by Card. It is not valid, however, to suppose that such tectonic features controlled or imposed restrictions on coastline positions. A long-lived hinge line does not necessarily mark the position of a static coastline. A great thickness of strata does not require that all, most, or any sediments were deposited at or immediately below sea level. The Cretaceous of the Foothills-Plains belt, for example, includes relatively thin successions deposited in advancing and retreating Arctic and Gulf seas at the time that very thick clastic wedges were deposited, mainly subaerially, in a narrow belt along the uplifted west border of the basin.

Pettijohn pointed out that fining-upward cycles characteristic of fluvial sediments deposited by meandering rivers have not been described in Huronian arenites. He acknowledged, however, that this did not preclude the possibility that they were deposited in other fluvial environments, notably braided streams. Moreover, it has been suggested that absence of land plants would modify the patterns of sedimentation developed in meander belts. In any case, the lack of such cycles is no more significant than the lack of positive evidence for deltaic, littoral or marine depositional environments. Silts and muds form the tops of the cycles developed in meander belts, the bases of cycles in deltas, and they should be prominent in any sequence formed in migrating or oscillating littoral environments. The scarcity or absence of substantial layers of siltstone and shale through hundreds, or even thousands of feet, of arenaceous beds is hardly a reason for believing that these formations were deposited under marine conditions rather than under high energy fluvial conditions wherefrom silt and clay was carried seaward. The question "Where are the missing shales?" was in fact raised by Pettijohn.

Our paper dealt explicitly with this point. The thickest and volumetrically most important units in the Huronian succession are great wedges of coarse, quartzose, crossbedded arenites that thicken southward to as much as 7,000 feet through known widths of 25 to 50 miles for most units, and as much as 70 miles in the case of the Lorrain Formation. The original extension farther to the south has been destroyed by tectonic developments or covered by Paleozoic strata and waters of Lake Huron. The bulk of the sands represent the coarsest fraction of first generation, submature to supermature, weathered detritus derived from soils developed atop Archean igneous and metamorphic rocks. Clay and silt-sized particles comprised a much greater proportion of these soils than the medium to very coarse quartz grains which are the dominant constituents of the arenaceous formations. Thus, fine-grained sediments must have been deposited contemporaneously with the sands and in greater volume than the coarse quartzose sands presently preserved. The greater part of the clastic wedges - including their axes of maximum thickness, their downstream tails of fine sediments and a broad zone containing interlayered shale and sand - are missing. Transitions from marine to non-marine conditions were more likely to have been represented in these missing parts of the wedges than they are to be present in the preserved parts which consist of relatively homogeneous quartzite deposited in a high energy environment.

Unconformities attest that parts of the Huronian belt, if not the entire area, were emergent periodically. Crossbedded sands and gravels of the Matinenda Formation were deposited on a deeply-weathered surface with appreciable relief. The uranium ore deposits are within great elongate lenses of grit and conglomerate which occupy basement valleys and extend downstream from valley mouths. Individual beds abut against valley walls. Thinning of units over grit zones suggests that these zones had elevated central parts so that they resemble alluvial fans (Roscoe, 1959, p. 66). An unconformity marked by scour channels as much as 300 feet deep is present within the Matinenda Formation near Elliot Lake. Sands above this unconformity have similar structures and mineralogy to those below it, but they are better sorted, suggesting that they may be redeposited products of erosion and washing of upstream equivalents of the lower sands (Pienaar, 1963). The formation was extensively eroded in the Quirke Lake area prior to the deposition of the Ramsay Lake Formation. The Mississagi and Serpent Formations were also uplifted and eroded prior to deposition of succeeding units. A bleached, silicic zone at the top of the Mississagi Formation at Quirke Lake was probably produced by subaerial weathering. Emergent conditions may have been the rule rather than the exception during Huronian time. There is no reason to suppose that sediments were not deposited on emergent surfaces or that the preservation of such terrestrial sediments would be unlikely.

Both Pettijohn and Card support the concept that several major periods of glaciation occurred during Huronian time. Their tacit assumption that all Huronian sediments were deposited at, or near, sea level is not wholly consistent with this belief. By analogy with Pleistocene and recent events, we should expect land elevations in the Huronian area and sea levels to have changed by many hundreds of feet during accretion and melting of great ice sheets and glacial loading and unloading of the land surface (Roscoe, 1969, pp. 77, 78). The highest relative positions of the sea would be reached when the ice caps melted and the land was still depressed; thereafter the land would rise and the sea coast migrate rapidly to the south. Fine grained sediments immediately overlying tillites were deposited in shallow water as evidenced by ripple marks and mud-cracks, so it seems reasonable to suppose that the overlying crossbedded arenites were deposited inland above sea level. Some of the coarsest arenites may have been deposited hundreds of feet above sea level, hundreds of miles from the sea, and much closer to the headwaters of river systems than to their mouths. This is particularly likely in the case of relatively poorly-sorted quartz gravels and grit in the Matinenda Formation, in the northernmost sections of the Mississagi Formation and the lower part of the Lorrain Formation. These are highly enriched in heavy minerals and may be piedmont placers. The idea, expressed by Holmes (1956) and by Robertson and Douglas (1970), that they are gravel deltas, is not supported by any evidence and does not seem to be a useful precept to apply as a guide to exploration.

Nothing in our paper requires more careful scrutiny than our acceptance of the hypothesis that the atmosphere was oxygen-free during deposition of most of the Huronian rocks. The suggestion that the sub-Huronian iron-deficient paleosols might have been developed under conditions of poor drainage is interesting and would raise serious doubts if only one or a few isolated occurrences of these unusual weathering profiles had been observed. The known occurrences are very widespread, however, in outcrop, in diamond drill cores, and in mine workings; all lack ferric oxide. They are overlain by coarse, current-bedded clastics and by sorted cobble conglomerates as well as by pebble conglomerates. Soil-mantled ridges evidently stood above the level of deposition of basal members in places. There is no doubt that the soils were developed mainly on sloping ground where the drainage should have been good.

The idea that euxinic conditions produced by biogenic processes, rather than a lack of atmospheric oxygen, prevented oxidation and fixation of iron in soils and in clastic sediments may be examined briefly. Sources of organic reductants would have been much more limited in early Aphebian time than in subsequent geological times. Certainly there could not have been any vegetative mats between the soil and the atmosphere. Plant debris would not have been carried and deposited with the detrital sediments. Some analyses that include carbon contents (non-carbonate) and sulphur contents of various rock samples from the Elliot Lake area have been published by Roscoe (1969, Appendix Table A), and others that give trace element contents and sulphur isotope ratios in sulphides in various rocks are also given in the same publication (Roscoe, 1969, pp. 130, 131). The carbon contents of paleosols and Huronian arenites is very low and isotopic fractionation of sulphur is slight. These data may not, in themselves, disprove the possibility of strong reducing conditions but they certainly do not provide any support for this idea. Argillites and argillaceous conglomerates that we consider to have been deposited in marine or marginal marine environments, incidentally, contain a little more carbon (0.1 to 0.6%) but not the sort of amounts that one would expect to find in euxinic sediments. The sulphides in these argillaceous rocks contain distinctly fractionated sulphur and were certainly biogenic in origin.

Arenites are commonly drab-coloured for a number of reasons in addition to that mentioned by Pettijohn - deposition in unaerated marine environments. They may be orthoquartzites that contain virtually no iron, lithic sandstones with iron locked in unaltered mineral grains, or they may contain abundant dispersed carbonaceous reducing material derived from land plants. None of these causes can apply to all of the pre-Cobalt arenites in the Huronian Succession. Our reasons for doubting that they are marine sediments have been stated but this is not an absolutely critical factor. Marine redbeds are very common. The point is that features such as extreme coarseness, trough crossbedding, ripple marks, dessication cracks on argillaceous partings, and mud chip clasts (Serpent Formation) indicate that the sediments were carried and deposited in aerated waters.

We are not familiar with the drab-coloured sediments in the lower Paleozoic of the Central Appalachians which Pettijohn has mentioned, so we cannot say whether they resemble the Matinenda, Mississagi and Serpent Formations. We do consider, however, that these Huronian formations are very similar to many arenites that contain hematite and some reddish colourations including: the lower part of the Lorrain Formation in the uppermost Huronian Group; the Keweenawan, Athabaska, Dubawnt sandstones; and Paleozoic, Mesozoic and Tertiary non-marine sandstones as in Carboniferous basins of the Appalachian region, in the Colorado plateau and in the Wyoming basins. We know of no major Proterozoic quartzite formations in the Canadian Shield that lack hematite but contain pyrite on a regional scale, outside of the Huronian succession, which is probably older than others; (there is a small patch of such an arenite, resembling the Mississagi Formation and associated with paraconglomerates similar to Huronian paraconglomerates, 230 miles northwest of Churchill, Manitoba).

The suggestion that the occurrence of detrital pyrite in Huronian conglomerates may not be very significant is a surprising one. Pyrite is not merely present in the conglomerates but it is abundant, other iron-minerals -

magnetite, ilmenite and hematite - are scarce, and uranium-rich minerals are present on a regional scale. The situation is an extraordinary one, unknown in modern placers and, as far as we know, unknown in fossil placers younger than Huronian rocks. Radioactive black sand placers with high thorium to uranium ratios, on the other hand, may be found in many sandstones which are similar structurally, mineralogically, and texturally to the Huronian feldspathic quartzites. This includes the Lorrain Formation and other sandstones mentioned above.

The various unusual features of Huronian soils, sands and placers discussed above could be logically explained as related to a critical phase in the evolution of the atmosphere. No other adequate explanation has been advanced as far as we are aware.

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Response by: F.J. Pettijohn

I yield to none in my admiration for the detailed, painstaking work of Roscoe and Frarey in their mapping and restudy of the classic Huronian - work carried out in the best tradition of the Geological Survey of Canada. My remarks on the problems of the Huronian were not intended in any way to detract from or denigrate this work nor to challenge the validity of the mapping. My intent was to call attention to some of the still unsolved problems - some of the contradictions that exist and differing opinions held by other competent geologists relative to the stratigraphic position of the Sudbury series and also to call attention to some of the uncertainties and gaps in our knowledge of the sedimentological history of the Huronian.

The problems touched on in my paper and discussed by Roscoe and Frarey are primarily three: (1) the McKim - both its sedimentological significance and its relations to the Huronian and the Whitewater series, (2) the sedimentological history of the great Huronian quartzites, and (3) the problem of the Huronian atmosphere.

There is little that I can add to the problem of the relations between the McKim and the Huronian. I simply point out again the contradiction between the conclusions set forth by Roscoe and Frarey and those of Thomson based on his mapping in Maclennan Township at the south end of Wanapitei Lake. I regret that Roscoe and Frarey did not address themselves to this problem.

Nor can I add anything further to the problem of the Whitewater series other than to say that the Chelmsford sandstone appears to be a turbidite formation and is, therefore, sedimentologically more like the McKim than any other formation in the region. That the two might actually be equivalent is not a new idea (Thomson, 1956, pp. 25-29). While lithologic similarity is an unreliable guide to correlation, correlation of sedimentologically incompatible formations should be made with caution.

Despite an extensive literature dealing with the stratigraphic problems of the McKim, I could find no adequate description of it and almost nothing in the way of a proper sedimentological analysis. I know of no adequate field description - not even a measured section so that even the bulk lithology is inadequately known. Vague references are made in the literature to greywackes, slates, and quartzites but none are described. Coleman (1914, p. 212), who named the formation, notes that greywacke is the dominant rock of the main belt, that quartzite is present in areas west of Nairn. The published field descriptions do little more than make brief mention of the graded bedding, dark colour, and passing reference to ripple marks and crossbedding. I could find no adequate statement of composition, no modal or chemical analyses, not even a photomicrograph, and no description of the sedimentary structures.

I have seen the McKim at several places - on Highway 68 just north of the Spanish River crossing and on Highway 17 near the Kelley Lake road. The rocks exposed in these places are essentially greywacke and slate, and show the usual characteristics of a normal turbidite sequence: rhythmic bedding, grading, flame structures, climbing ripples of Walker's type 3 (believed by him to be diagnostic of turbidity current deposition), and various structures indicative of soft-sediment deformation. If these exposures are typical, the McKim is certainly a turbidite formation and hence a deep-water sediment - not a fluvial or deltaic deposit.

Our knowledge about the place of the McKim in the scheme of things is even less satisfactory than about its character. We do not know where it came from, no paleocurrent data having been published. What we know about its composition, sedimentary structures, general immaturity and association with volcanics (Stobie) including pillow lavas suggests a different history from that of the Huronian quartzites. The McKim is a part of a 15,000-foot accumulation of subaqueous volcanics and turbidites. It is indeed difficult to conceive of a greater tectonic and depositional contrast between this assemblage and the superjacent Huronian.

The Huronian quartzites are a different problem. We have several excellent sedimentological studies of these by McDowell (1957), Pienaar (1963), and Hadley (1968) so that we know something about them. The most significant character of fluvial sediments is their organization in fining-upward cycles which are characteristic of the best-known alluvial sequences of the geologic record but are lacking in the Huronian. As noted, the absence of such cycles does not vitiate the fluvial hypothesis - it only makes proof of fluvial origin more difficult. But it is not just the absence of these cycles that makes a fluvial origin unlikely. Even if a different stream regimen were conceivable, the petrology of the sands would not be significantly different. All the known alluvial sands are immature - poorly rounded, generally arkosic or lithic sands, loaded with polycrystalline quartz and pelitic particles. The deposits of a braided stream complex differ in organization but they would not differ in composition and texture from those of a meandering stream.

Like the well-known Appalachian Silurian Tuscarora quartzite, in places over 1,000 feet thick, the Huronian quartzites may have accumulated in the shore zone, being in part neritic, in part littoral, and in part alluvial. As Folk (1960, p. 14) noted, the marine facies of the Tuscarora is much better rounded, better sorted, and has the highest proportion of monocrystalline quartz and the lowest content of polycrystalline quartz and rock particles. The fluvial facies, in contrast, tends to be pink in colour, display mud-chip conglomerates and be much less mature in composition and texture.

Some of the most mature and best-rounded sands in the geologic record are Shield quartzites - such as the Sioux and the Palms. Comparable rounding, sorting and compositional maturity also characterize the upper Lorrain, the Bar River, and the Mississagi. It is noteworthy that many of the Shield quartzites are associated with or overlain by dolomite. This is true of the Sturgeon, the Mesnard, the Thelon, the Athabaska and the Wilson Island quartzites to name a few. Some such as the Sioux, the Sturgeon, and Whiterock Lake (Hurwitz Group) display an abundance of oscillation ripples suggestive of a marine rather than a fluvial origin.

The kind and scale of the crossbedding and the grain-size characteristics are no sure guide to origin. Very large-scale crossbedding can be marine. The bimodal distribution of azimuths is good evidence of tidal currents but the absence of this feature is no proof of fluvial origin.

In all respects, except perhaps thickness, these quartzites resemble most closely the Lower Paleozoic cratogenic sands of the Upper Mississippi Valley of marine origin. Certainly the high degree of rounding can only mean eolian or surf action, rivers being incapable of producing such rounding. Although it is possible that the last episode was fluvial, it seems more likely that a considerable part, perhaps most, of these quartzites are marine - not fluvial.

The question of the Huronian atmosphere is even more difficult to answer than the depositional environment and the significance of the McKim or the Huronian quartzites. The problem is compounded by various assumptions as well as by lack of knowledge. We don't know, for example, what detrital pyrite means, having little or no information on its persistence in modern sediments. It can survive transport and has even been used as a tracer mineral in study of shore drift. If brief transport is followed by burial, it could well be present as a detrital component. The presumed absence of land plants in pre-Devonian times also colours our thinking. In fact, we have no assurance that the pre-Devonian lands were without plant cover. Bare rock surfaces today soon acquire a lichen cover and we cannot assume an absence of such cover in the Precambrian.

Photosynthesis implies breakdown of CO_2 and release of oxygen. Once initiated this mechanism will produce an oxidizing atmosphere. Remains of blue-green algae in the Fig Tree Series (Schopf and Barghoorn, 1967) and the stromatolites of the Bulawayo Series of South Africa show that photosynthesis far antedates the Huronian. The geologic record of this activity is the carbon locked up in sediments, the shales in particular. The oxygen equivalent of the carbon locked up in the sediments - even the Archean sediments - is probably many times greater than that in the present atmosphere. It seems most unlikely that the Huronian atmosphere was reducing.

The thrust of my comments on Roscoe's views on the Huronian atmosphere was to point out only that none of the criteria used by him are conclusive and that lack of adequate data make any conclusions at this time suspect.

It seems to me that the areas of our disagreement and uncertainty are largely the result of our lack of knowledge about the rocks themselves. Up to this moment the greatest efforts have been expended in mapping and establishing a stratigraphic and structural framework. This is as it should be and is most necessary. The time has now arrived to look at the rocks themselves. A good beginning has been made in the past decade but many formational units have never had a proper study. The lack of elemental data - such as modal and chemical analyses, measured sections, paleocurrent data - on just "what's there" are lacking. Until such studies are made, our ability to synthesize and interpret Huronian history will remain unsatisfactory and incomplete.

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Comments by: J.A. Fraser, J.A. Donaldson,
W.F. Fahrig, L.P. Tremblay.

Re: The Canadian Shield - A Status Report 1970

Professor Pettijohn has drawn attention to the similarities between the Helikian succession described in our paper and the Keweenawan succession of the Lake Superior region. The Keweenawan rocks he interprets as occupying a rift valley, noting the absence of marine strata in this sequence as well as the lack of folding, thrusting, metamorphism, and granitic intrusion. He considers that deposits of the northwestern Shield are not geosynclinal 'in the usual sense' but are 'large residuals of once more extensive, mainly continental sediments'.

In considering these comments it is important to recall that Helikian strata in the northwestern Shield record deposition in three distinct stages. The first stage is characterized by red sediments and intermediate to basic lavas deposited in fault-controlled basins. The second stage is represented by marine carbonates and fluvial clastics. The third stage is marked by a thick sequence of basaltic flows (Coppermine River) with interbedded red sediments.

We agree with Professor Pettijohn that there are similarities between some of these strata and those of the Lake Superior region but, in our opinion, the similarities apply only to the deposits of the first and third stages. The Coppermine River flows and associated red sediments certainly have many features in common with the Keweenawan succession. Moreover, the approximate equivalence of age (1,000 to 1,200 m.y.) in the two successions has been well documented.

Sediments of the second stage, however, are lithologically mature and include marine carbonate sequences. Futhermore, they exhibit a pattern of transport that indicates that they were not deposited in isolated continental basins, and suggests a significant flow of clastic material beyond the present western limits of the Shield. These features, together with the thickening of clastic and carbonate units westward and the occurrence in the Cordilleran Province of eugeosynclinal sediments of Helikian age, supports our hypothesis that the Helikian sediments of the northwestern Shield represent the eastern margin and platform of a Helikian geosyncline.