

Iron-oxide-Cu-Au-(U-REE) deposits – A practical guide and  
update of recent developments

**Part 1. Introduction**

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## Introduction

Iron oxide copper-gold (IOCG) deposits are a comparatively recently recognized class of copper-gold deposits that are the focus of major exploration activity on several continents. This exploration interest has been prompted by the recognition that IOCG deposits have tonnages and grades similar to those found in porphyry Cu-Au systems (Fig. 1a), with major examples including Olympic Dam in South Australia, Candelaria in Chile and Salobo in Brazil.

The aim of this short course is to familiarize participants with the geological settings and characteristics of IOCG deposits, to examine some of the factors that control mineralization in these systems, and to discuss exploration for this highly variable style of mineralization. Suites of hand samples are available so that participants can become familiar with alteration styles and mineralogical characteristics of a range of different deposits. In some cases, these specimens also provide valuable insights into possible metal precipitation mechanisms and the likely geophysical characteristics of the deposits.

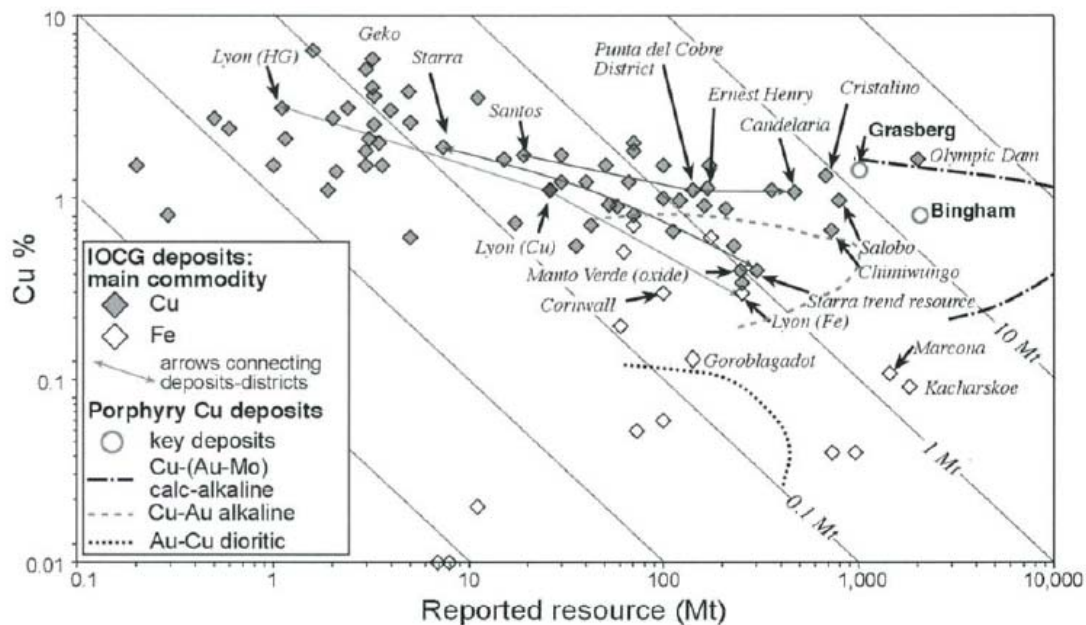
The Fe-oxide-Cu-U-Au-REE class of ore deposits was recognized by Hitzman et al. (1992), who proposed that Cu-Au deposits such as Olympic Dam and magnetite-apatite deposits such as Kiruna form part of an ore deposit class they designated as Proterozoic iron-oxide (Cu-U-Au-REE) deposits. Hitzman et al. (1992) inferred that the deposits formed primarily in shallow crustal environments (<4-6 km) as expressions of deeper-seated, volatile-rich igneous-hydrothermal systems. The deposits were noted as being predominantly mid-Proterozoic in age, located in cratonic or continental margin environments, and in many cases linked to extensional tectonic regimes. The deposits were regarded as being rich in Fe-oxides (magnetite or hematite), commonly with associated CO<sub>3</sub>, Ba, P, or F minerals. Associated alteration was envisaged to be sodic at deep levels, potassic at intermediate to shallow levels and sericitic/siliceous at very shallow levels (Hitzman et al., 1992).

Since that time there has been a surge in exploration interest in IOCG deposits prompted by discoveries in several districts (eg. Ernest Henry, Australia; Candelaria, Chile; Sossego, Brazil). The age of the deposits has also been extended to include Phanerozoic examples such as those in northern Chile and Peru, as well as Archean examples in the Carajas district of Brazil.

Much has been written in the last couple of years about Fe-oxide-rich deposits and the genetic models that have been proposed to account for their formation. This will be discussed later, but it is worth pointing out that this short course has a primary focus on Cu-Au deposits that are commonly rich in Fe-oxides and does not attempt to cover other types of Fe-oxide accumulations (eg. Kiruna-type magnetite-apatite deposits). Much of the controversy surrounding the origins of the Fe-oxide-rich deposits reflects the differing focus of the various authors upon either the Fe-oxide or the Cu-Au part of the title. Amongst those who have studied Cu-Au deposits containing abundant Fe-oxides there is a strong consensus that these deposits represent part of intrusion-related magmatic-hydrothermal systems. It is these Cu-Au deposits that form the focus of this course.

In contrast to porphyry Cu-Au deposits, the iron-oxide Cu-Au deposits do not show a consistently close relationship to subvolcanic intrusive complexes, but formed over a much larger range of crustal depths than typical porphyry systems. However, the IOCG deposits have several features in common that can serve as a working definition:

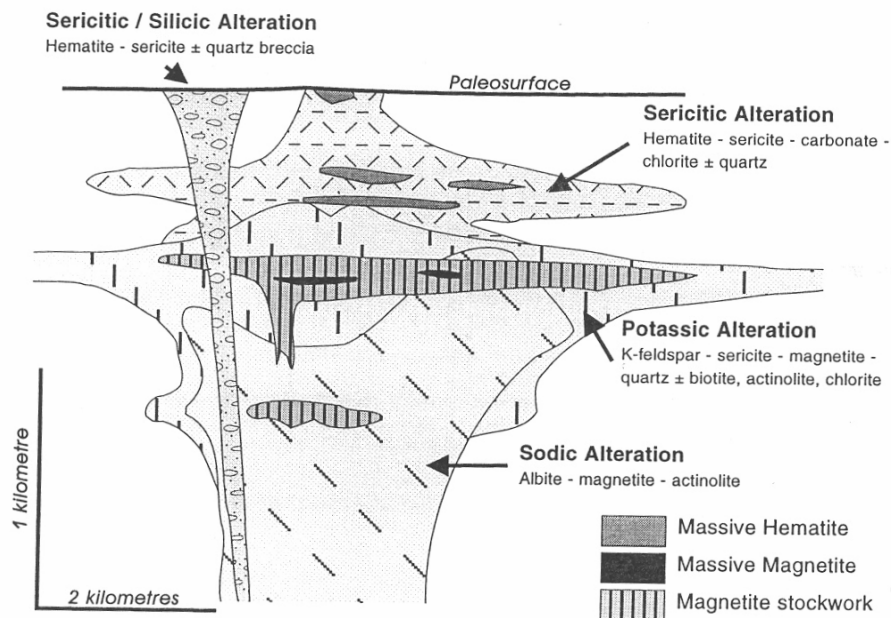
1. mineralization, especially at deeper levels, is controlled largely by reactivation of pre-existing tectonic structures rather than by structures formed during magma crystallization and hydrothermal fluid generation (cf. porphyry Cu-Au systems),
2. deposits commonly have a high-Fe, low-S character such that iron-oxides (magnetite and/or hematite) rather than Fe-sulphides (e.g. pyrite) are abundant constituents of the ores, and
3. in addition to copper and gold, the deposits are commonly, but not universally enriched in Co, Mo, U, REE, Ba and F.



**Figure 1a.** Copper grade – tonnage data for IOCG deposits (from Williams et al., 2005).

Most IOCG districts contain abundant mafic-intermediate to felsic intrusive rocks that were emplaced around the time of mineralization and there is increasing evidence to indicate that the intrusive rocks were major sources of fluids and metals for the deposits (Perring et al., 2000; 2001; Marschik et al., 2003). However, the deposits themselves typically occur peripheral to the intrusive complexes and are commonly hosted by older (meta)-igneous or sedimentary rocks.

Hitzman et al. (1992) presented a general model for IOCG deposits that envisaged a broad depth zonation of alteration types (Fig. 1b), with sodic alteration (albite-magnetite-actinolite) at deeper levels, potassic alteration (K-feldspar-sericite-magnetite-quartz±biotite, actinolite, and chlorite) at intermediate levels and sericitic alteration (hematite-sericite-carbonate-chlorite±quartz) at shallow levels. Sodic-calcic alteration dominated by assemblages of albite, scapolite, actinolite and clinopyroxene is developed on a huge scale in some IOCG districts and typically precedes Cu-Au mineralization. Albitization of a wide variety of host rocks has been shown to cause addition of Na and mobilization of K, Fe, Rb and Ba, elements that are commonly enriched in the copper-gold deposits.



**Figure 1b.** Schematic cross section illustrating alteration zoning in iron oxide (Cu-U-REE-Au) deposits. The section extends from the near-surface to several kilometres below surface (from Hitzman et al., 1992).

## Geological setting and age of IOCG deposits

IOCG deposits form in several geological environments (Fig. 2) and span most of earth history. Part of the diversity of these deposits is demonstrated by four of the largest known examples which, despite a number of general similarities, formed in geological environments ranging from mid-crustal (Ernest Henry) to shallow-level (Olympic Dam; Reeve et al., 1990), in the Archean (Salobo; Requia et al., 2003), Proterozoic (Olympic Dam, Ernest Henry; Ryan, 1998) and Phanerozoic (Candelaria; Marschik and Fontboté, 2001).

IOCG deposits are known from the Archean to the Pliocene. Recent geochronological study in the Carajas district, Brazil has indicated Re-Os ages for molybdenite of 2.57 Ga for Salobo (Requia et al. 2003) and 2.609 Ga for Serra Verde (Marschik et al., 2001), a U/Pb age of monazite of 2.57 Ga for Igarapé Bahia (Tallarico et al., 2005), and a minimum  $^{40}\text{Ar}/^{39}\text{Ar}$  age for amphibole from Sossego of 2.2-2.3 Ga (Marschik and Leveille, 2001). The majority of known IOCG districts (Fig. 2) are Proterozoic in age (e.g. Cloncurry, Tennant Creek, and Gawler Craton, Australia; Norbotten, Sweden), but many deposits are also known from the Phanerozoic of northern Chile and southern Peru (Late Jurassic-Early Cretaceous)(see review by Sillitoe, 2003). In addition, Fe-rich deposits in the Durango district of Mexico (~30 Ma, Oligocene; Lyons, 1988) and the Miocene age Fe deposits of the Iron Springs district of southwestern Utah (Barker, 1995) are commonly included within this group of deposits. The El Laco magnetite-apatite deposits in the high Andes of Chile are amongst the youngest known Fe-oxide deposits that may form part of this class, and these occur in Pliocene-Pleistocene age host rocks (see Rhodes and Oreskes, 1999; Rhodes et al., 1999).



In Proterozoic environments IOCG deposits are commonly believed to occur in intracratonic settings (e.g. Midwest U.S.A., Arctic Sweden/Finland), and are sometimes proposed to have formed during rifting (e.g. Hitzman et al., 1992; Fig. 3b). However, a closer examination in several regions has shown that granites and mineralization were emplaced during compressional or transpressional deformation syn- to post-peak regional metamorphism (Ettner et al., 1993; Frietsch et al., 1997; Pollard et al., 1998; Ferris et al., 2002) and the tectonic setting is in some cases interpreted in terms of distal continental volcanic arc environments rather than intracontinental anorogenic settings (e.g. Hildebrand, 1986; Ferris et al., 2002). IOCG deposits of Phanerozoic age occur in the eroded parts of continental margin volcanic arcs in the Coastal Belt of Chile and Peru (Sillitoe, 2003).

The Cloncurry district in northwest Queensland, Australia is one of the best studied IOCG provinces of Proterozoic age. This district forms the eastern part of the Proterozoic Mount Isa Inlier and contains two main sequences of supracrustal rocks which appear to have formed in a large intracratonic basin. The earlier sequence (Cover Sequence 2; 1780-1720 Ma; Blake, 1987) is composed of mafic metavolcanic rocks and a clastic-carbonate-evaporite metasedimentary sequence. The younger sequence (Cover Sequence 3, <1670; Blake, 1987; Page and Sun, 1998) is dominantly siliciclastic in character and is composed of mafic volcanic and intrusive rocks, sandstone, arkose and shale. Initial deformation and peak metamorphism occurred around 1590 Ma, and a second major period of metamorphism occurred during the Isan Orogeny (1550-1500 Ma) and was accompanied by E-W shortening.

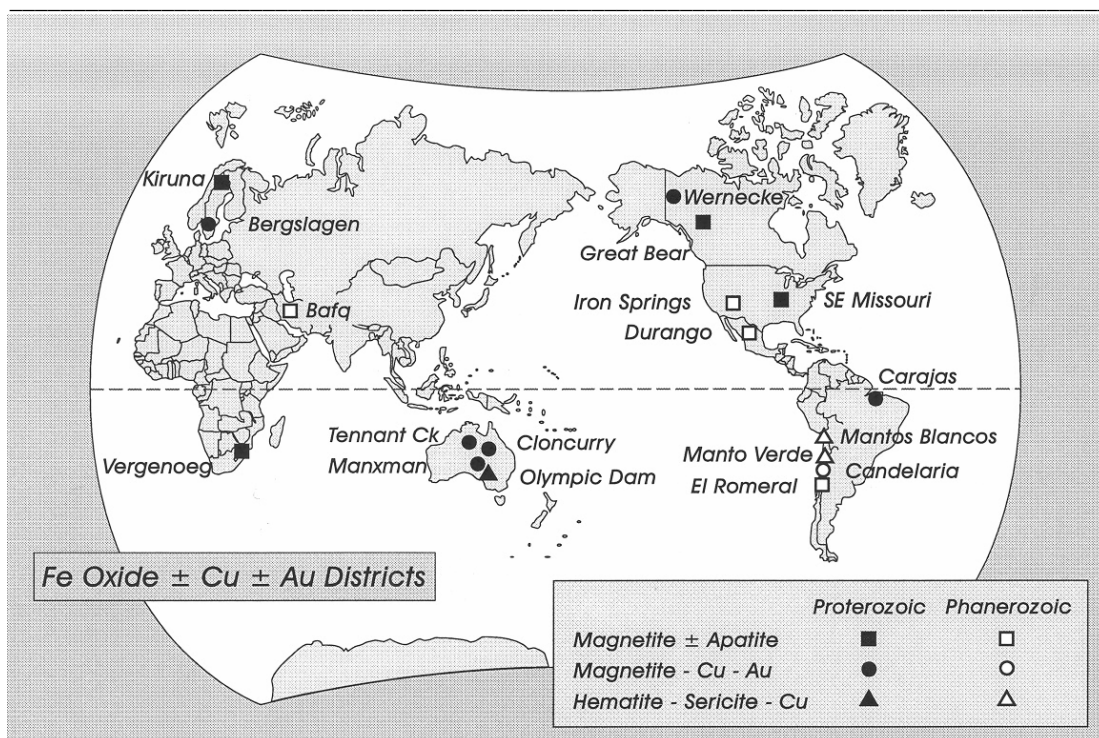
The low pressure, high temperature metamorphism in the Cloncurry district ranged from greenschist to upper amphibolite facies and was most likely produced by underplating and intrusion of mantle-derived basic magmas into the lower crust. During the Isan Orogeny this resulted in partial melting in the lower crust that formed voluminous granitoid magmas which transferred heat to the middle and upper crust and formed major granitoid batholiths (see Foster and Rubenach, 2001). Trondhjemite and granodiorite plutons were intruded syn the major D<sub>2</sub> deformation and accompanying metamorphism. However, most IOCG mineralization formed during emplacement of voluminous intrusions of the post-peak metamorphic Williams and Naraku batholiths (ca. 1530-1505 Ma; see Page and Sun, 1998; Perkins and Wyborn, 1998; Pollard et al., 1998). Intrusion of these granitoids was accompanied by episodic sodic-calcic alteration that is localized within and around the intrusions and along regional fault systems. IOCG mineralization occurs within the sodic-calcic altered rocks and is commonly, but not universally associated with potassium-iron metasomatism (biotite and/or K-feldspar, magnetite and/or hematite).

The Gawler volcanic province is another Proterozoic IOCG district that has been a focus of attention recently (Ferris et al., 2002; Skirrow et al., 2002; Fig. 3b). The province consists of early Proterozoic deformed granitic and metasedimentary rocks that are intruded by quartz monzodiorite to granite plutons of the Hiltaba Suite (1585-1600 Ma; Mortimer et al., 1988; Creaser and Cooper, 1993; Johnson and Cross, 1991). The plutons intrude a consanguineous group of volcanic rocks known as the Gawler Range Volcanics which is dominated by felsic rocks, but also includes a minor mafic component. The Gawler Province contains areas of sodic-calcic alteration and magnetite-rich Cu-Au deposits such as Manxman and Joe's Dam as well as hematite-sericite-rich deposits such as Olympic Dam (Reeve et al., 1990; Williams and Pollard, 2001; Skirrow et al., 2002). The tectonics of the Gawler Craton was previously interpreted largely in terms of intracontinental processes and anorogenic magmatism, but Ferris et al. (2002) have suggested that the Proterozoic intrusive rocks and volcanics were accreted to an Archean core in a series of

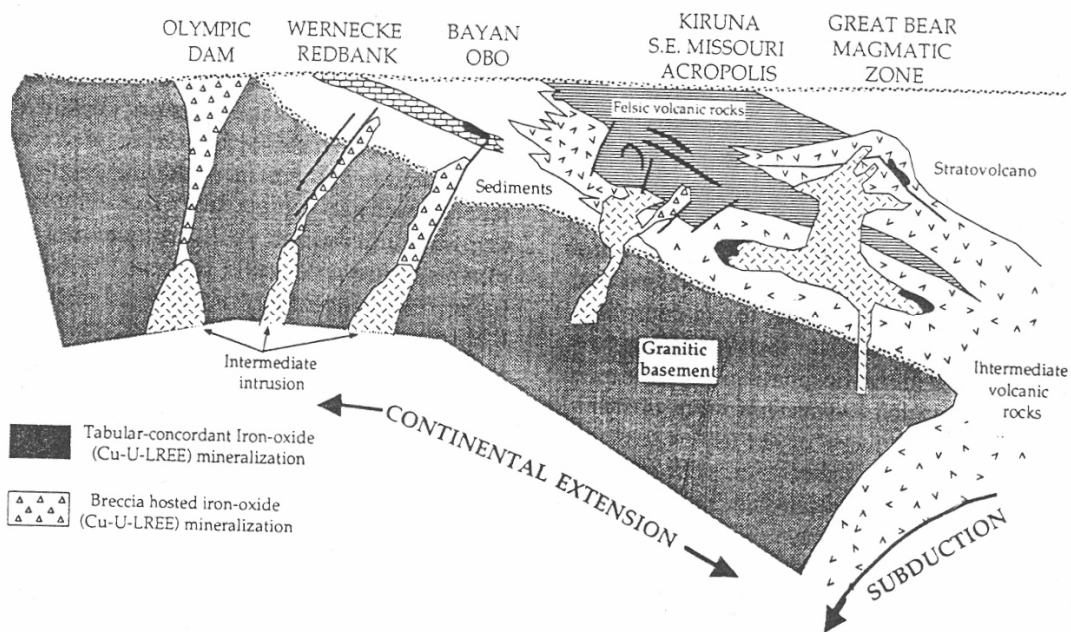
volcanic arcs. The Hiltaba Suite granites are now believed to be contemporaneous with the activation of crustal scale shear zones and related to a northeast dipping subduction zone located to the southwest of the craton (Ferris et al., 2002). In this view, the tectonic setting of the Gawler Craton during intrusion and mineralization has some similarities with that of the Phanerozoic Coastal belt in Chile.

	Carajás	Eastern Gawler Craton	Cloncurry	Northern Chile
Age Cu–Au mineralization	Archean	Proterozoic	Proterozoic	Mesozoic
Tectonic setting	Continental margin arc?	Intracontinental arc	Intracontinental arc	Continental margin arc
Major transcurrent fault system	Carajás fault	WNW faults at Olympic Dam	Mt Dore fault	Atacama fault
Intrusive rocks possibly linked to mineralization	Estrela and Old Salobo granites	Hiltaba Suite (Roxby Downs subsuite)	Eureka Supersuite	Coastal Batholith
Other intrusive types	Mafic and ultramafic	Mafic and ultramafic	Mafic	Mafic
Regional alteration	Sodic–calcic restricted	Sodic–calcic restricted	Sodic–calcic widespread	Sodic–calcic widespread
Evaporites	Not reported	Not reported	Mary Kathleen Group (metamorphosed)	Chañarcillo Group
Other mineralization styles	Ni–Cu and platinoids in mafic/ultramafic rocks	Au, Pb–Zn–Ag	Broken Hill type Pb–Zn–Ag, Co(–Cu) and Au-only deposits, Ni–Cu in mafic rocks	Magnetite–apatite, Porphyry Cu–Au, epithermal Au, Manto-type Cu–Ag
References	Holdsworth and Pinheiro (2000) Dardenne et al. (1988) Tallarico et al. (2005)	Ferris et al. (2002) Budd et al. (1998) Reeve et al. (1990) McLean and Betts (2003)	Williams and Pollard (2003) Pollard et al. (1998) Betts et al. (2002) P. J. Pollard (unpublished data)	Dallmeyer et al. (1996) Sillitoe (2003)

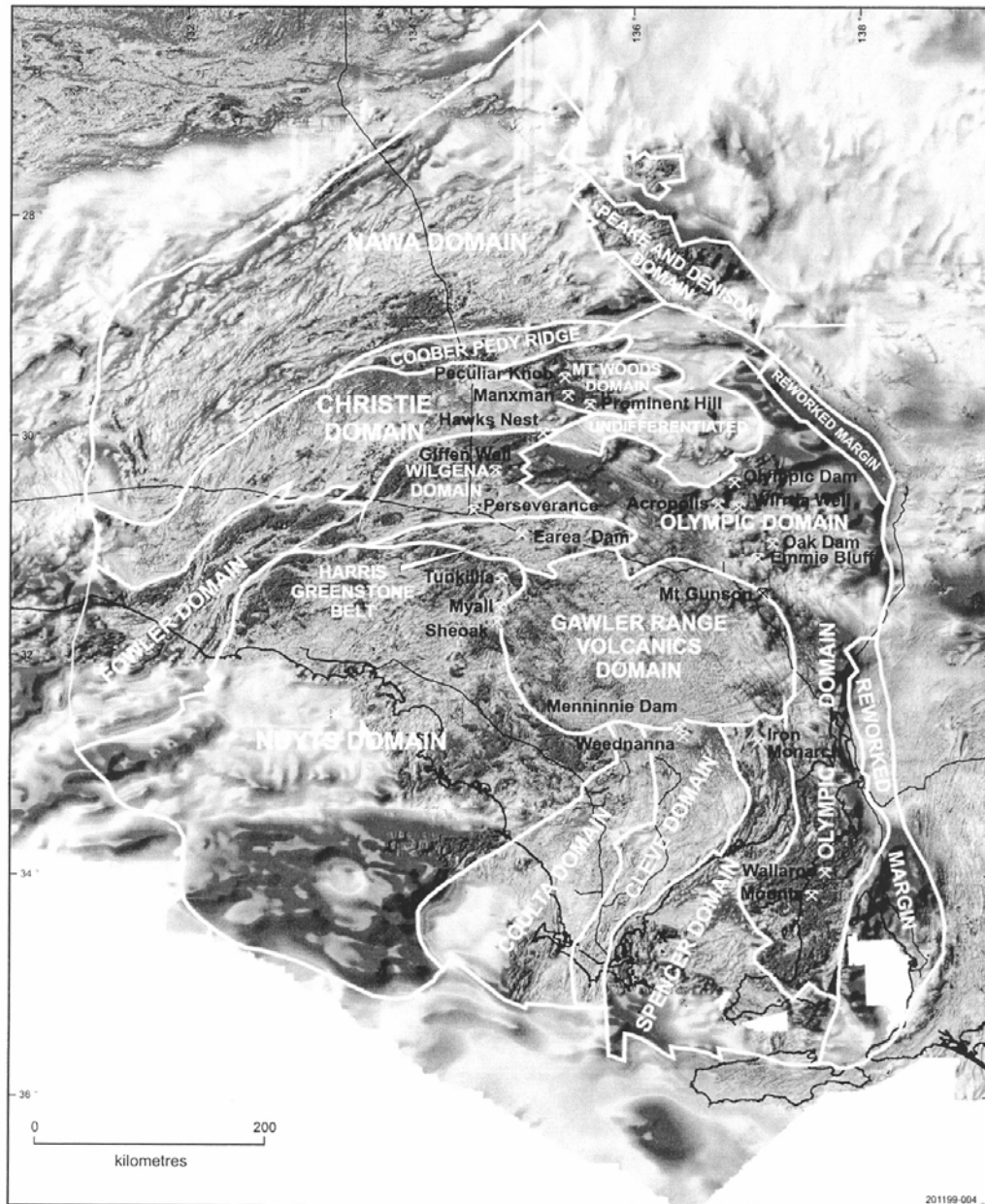
Table 1. Characteristics of IOCG provinces that contain major Cu–Au deposits (from Pollard, 2006).



**Figure 2.** Location of examples of Fe-oxide±Cu±Au districts and deposits (from Williams and Pollard, 2001).



**Figure 3a.** Schematic representation of the tectonic setting and host rock sequence for iron oxide (Cu-U-REE-Au) deposits (from Hitzman et al., 1992).



**Figure 3b.** Domains of the Gawler Craton superimposed on total magnetic intensity (from Ferris et al., 2002).

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## Nature of associated intrusive rocks

The intrusive rocks that are spatially and temporally associated with IOCG mineralization are magnetite-series granitoids that range from diorite to granite in composition. These commonly occur as extensive granitoid batholiths composed of multiple intrusions, in some cases in association with basaltic and/or ultramafic rocks. In many Proterozoic environments the magmas appear to have formed by partial melting of older igneous material in the middle to lower crust with a variable mantle input (Ramo and Haapala, 1995; Creaser, 1996; Pollard et al., 1998). This tends to produce magmas with low S and high F and CO<sub>2</sub>/H<sub>2</sub>O (Harris, 1982). The high temperatures required to promote partial melting via biotite and amphibole breakdown in such environments promotes higher solubilities of refractory minerals such as Fe-Ti phases and zircon, leading to enrichment of the magmas in components concentrated in these minerals, including zirconium, titanium, uranium and rare earth elements (e.g. Keppler, 1993). Uranium and rare earth elements partition strongly into fluorine-rich hydrothermal fluids (Keppler and Wyllie, 1991) and this may account for their enrichment in many IOCG deposits of Proterozoic age.

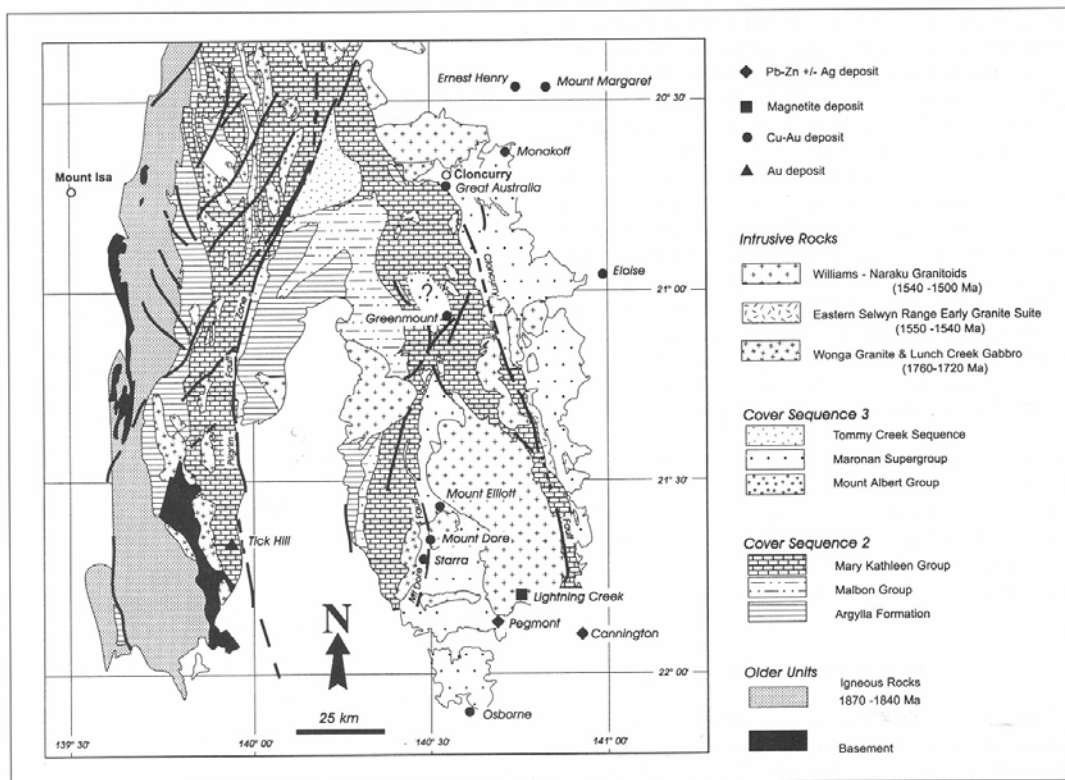
Detailed studies in the Cloncurry district (Pollard et al., 1998) indicate that the magmatism was long lived (ca. 40 Ma) and involved multiple magma sources. The granitoids linked to mineralization are alkaline and K-rich, and exhibit continuous fractionation trends from pyroxene-bearing diorites to hornblende-biotite granites (Figs. 4-6). Some of the geochemical diversity of the deposits may reflect fluids sourced from different parts of this compositional spectrum. At Mt Elliott there is evidence for intrusion of intermediate dykes during alteration and mineralization (Wang and Williams, 2001) and Ni and Co enrichment could reflect derivation of the hydrothermal fluids from such intermediate magmas (Fig. 1), and/or leaching of nearby, older basic rocks. In contrast, Ernest Henry is enriched in F, REE and Mo which could represent derivation of the fluids from more fractionated magmas of the Naraku granite.

The magmas linked to IOCG deposits exhibit a range of trace element characteristics reflecting their sources, which may include components of the down going slab, overlying mantle wedge and/or continental crust, including older metaigneous rocks. Mantle-derived mafic/ultramafic magmas are also commonly associated with the granitoids (e.g. Arancibia and Clark, 1996; Creaser, 1996; Pollard et al., 1998), and these may also provide components to the Cu-Au deposits (e.g. Olympic Dam). In some Proterozoic provinces the granitoids appear to be derived largely from older crustal materials, commonly with extensive evidence of mixing and mingling with more basic magmas (Creaser, 1996; Perring et al., 2001; Fig. 6).

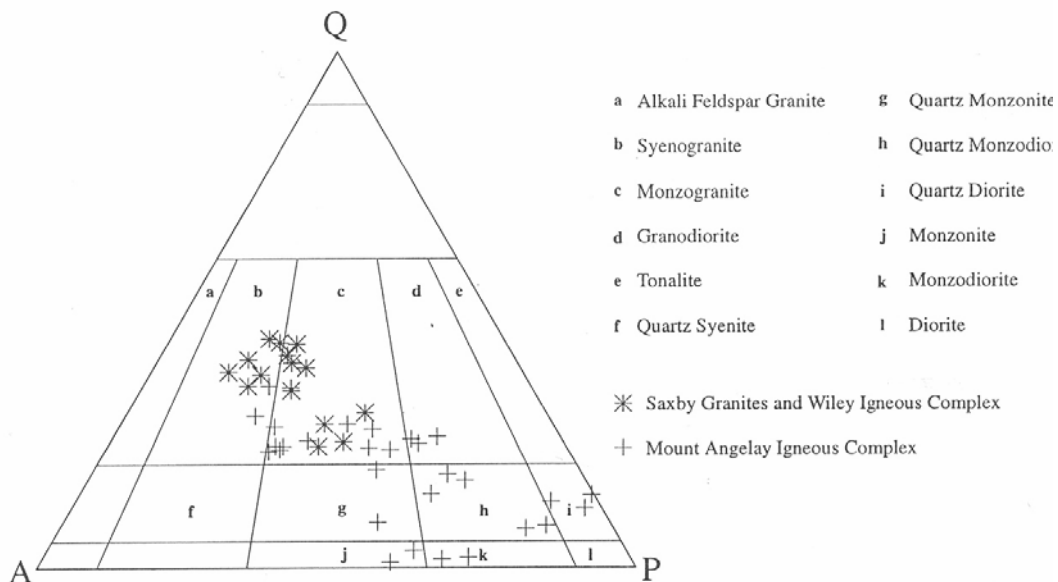
Experience in porphyry Cu-Au environments suggests that there is unlikely to be a unique mineralogical or chemical signature that would allow prospective plutons to be identified. Rather, it appears that the mineralizing potential of an individual intrusion may be a complex function of water (and other volatile) contents, degree of fractionation, nature of the liquidus minerals, and the timing and mechanism of fluid release.

Province	Carajás	Eastern Gawler Craton	Cloncurry	Northern Chile
Name	Estrela granite Old Salobo granite	Hiltaba Suite–Roxby Downs subsuite	Eureka Supersuite	Copiapó pluconic complex Sierra Dieciocho plutonic complex
Composition range	Tonalite to monzogranite (Estrela granite) Monzogranite (Old Salobo granite)	Quartz monzodiorite to granite	Diorite to monzogranite	Diorite to quartz monite and aplite
Oxidation state	Magnetite series (Old Salobo granite) Magnetite series (Estrela granite)	Magnetite series	Magnetite series	Magnetite series
Alkali–silica relationships	Subalkaline (Estrela granite)	Alkaline to subalkaline	Alkaline to subalkaline	Alkaline to subalkaline
Depth of emplacement	~15 km (Estrela granite)	6–8 km	10–15 km	2–3 km
Age	2,573±2 Ma (Old Salobo granite) 2,763±7 Ma (Estrela granite)	1,588±4 Ma (Roxby Downs Granite)	1,500–1,530 Ma	Copiapó plutonic complex—97–119 Ma Sierra Dieciocho plutonic complex—120–127 Ma
Major deposits	Salobo, Sossego, Alemão, Cristalino	Olympic Dam, Prominent Hill	Ernest Henry	La Candelaria, Manto Verde
Age of major deposits	Salobo—2,576±8 Ma Re–Os molybdenite Sossego—>2.2–2.3 Ga Ar–Ar amphibole Alemão—2,575±12 Ma, Pb–Pb monazite Cristalino—2,719±36 Ma, Pb–Pb sulphides	Olympic Dam—~1,590 Ma U/Pb zircon	Ernest Henry—1,514±24 and 1,529±11 U/Pb titanite ages	Candelaria—114.9±1.0 Ma Ar–Ar biotite, 114.2±0.6 and 115.2±0.6 Ma Re–Os molybdenite Manto Verde—117±3 and 121±3 Ma K–Ar sericite
References	Lindenmayer et al. (1994)	Creaser (1996)	Pollard et al. (1998)	Marschik et al. (1997, 2003a,b)
	Machado et al. (1991)	Creaser and Cooper (1993)	Page and Sun (1998)	Mathur et al. (2002)
	Barros et al. (1997, 2001)	Johnson and Cross (1995)	Mark et al. (2006)	Vila et al. (1996)
	Requia et al. (2003) Marschik et al. (2003c) Tallarico et al. (2005) Soares et al. (2001)	Haynes et al. (2005)		Sillitoe (2003)

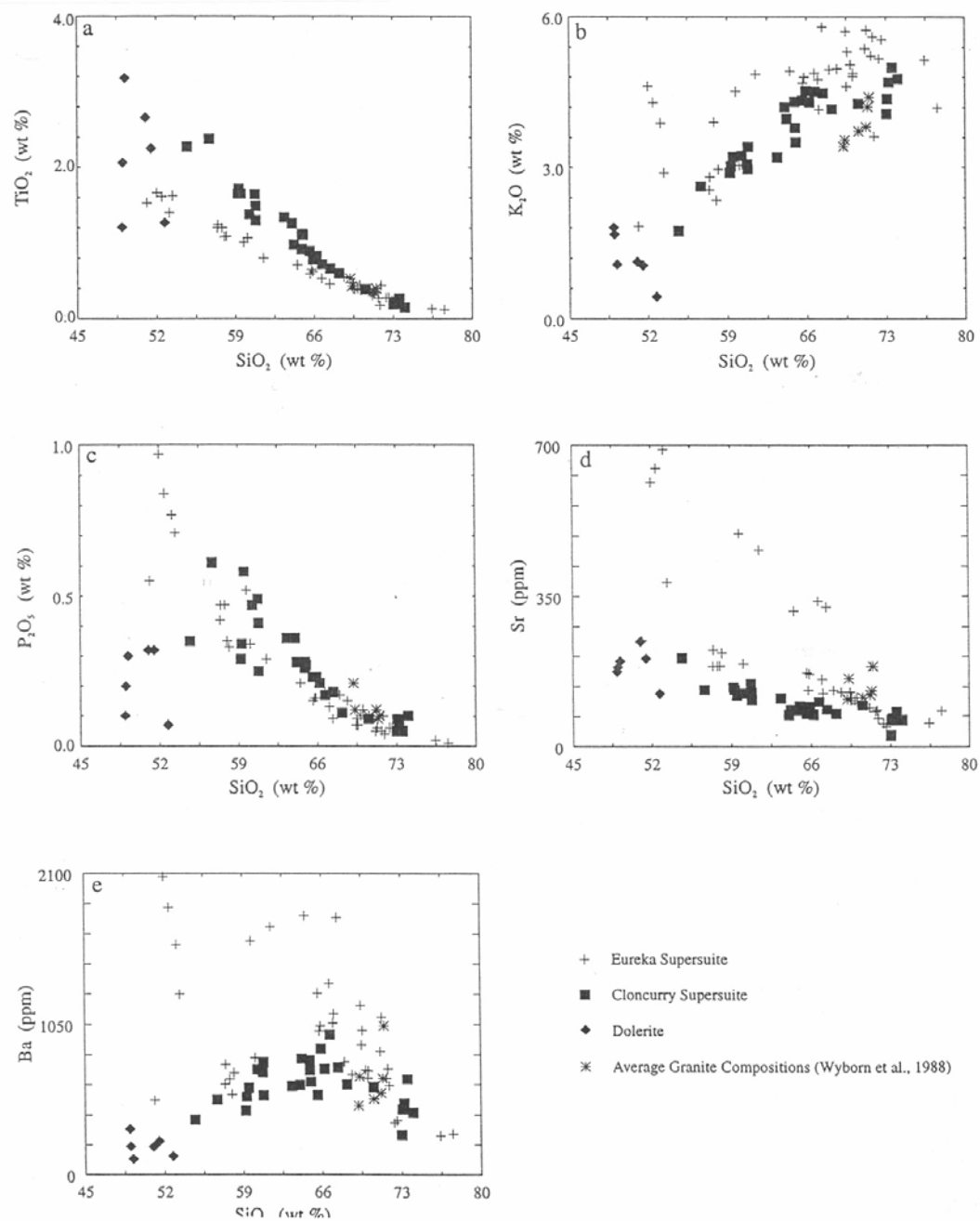
**Table 2.** Characteristics of granitoids and age of granitoids and Cu–Au deposits in major IOCG provinces (from Pollard, 2006).



**Figure 4.** Distribution of granitic rocks in the Cloncurry district of northwest Queensland. From Williams and Pollard, 2001...

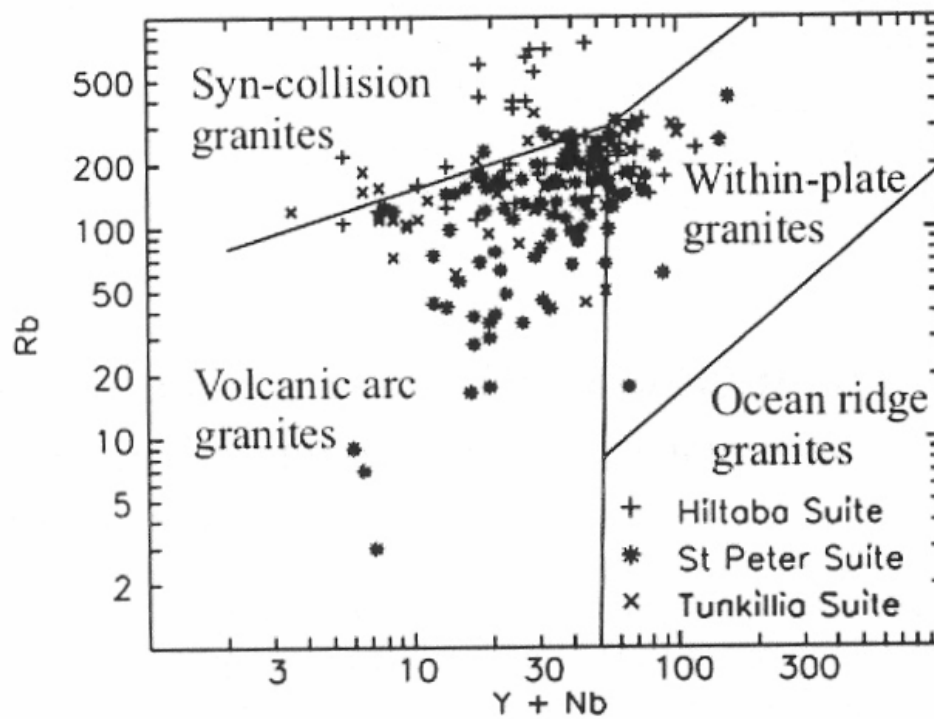


**Figure 5.** Quartz-alkali-feldspar-plagioclase triangle showing compositions of intrusive rocks from the Saxby Granite and the Mount Angelay Igneous Complex. From Pollard et al., 1998.

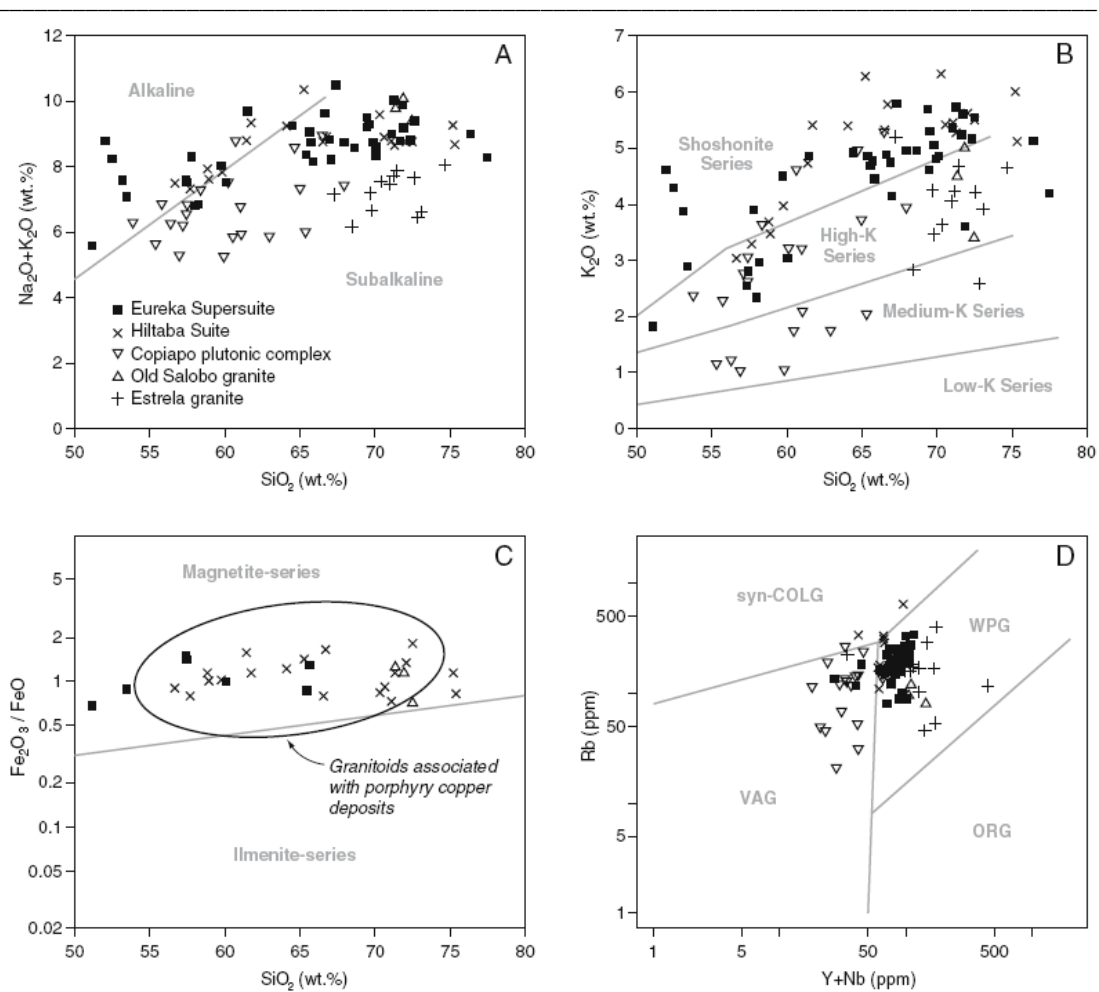


**Figure 6a.** Plots of whole-rock geochemical data for intrusive rocks from the Cloncurry district. Eureka Supersuite granitoids (which includes the Mount Angelay Igneous Complex) are thought to be related to IOCG mineralization. From Pollard et al., 1998.





**Figure 6b.** Rb vs. (Y + Nb) plot of Palaeoproterozoic and Mesoproterozoic intrusive rocks from the western Gawler Craton (from Ferris et al., 2002).



**Figure 6c.** Variation diagrams for granitoids temporally and spatially related to major Cu-Au deposits in IOCG provinces (from Pollard, 2006).

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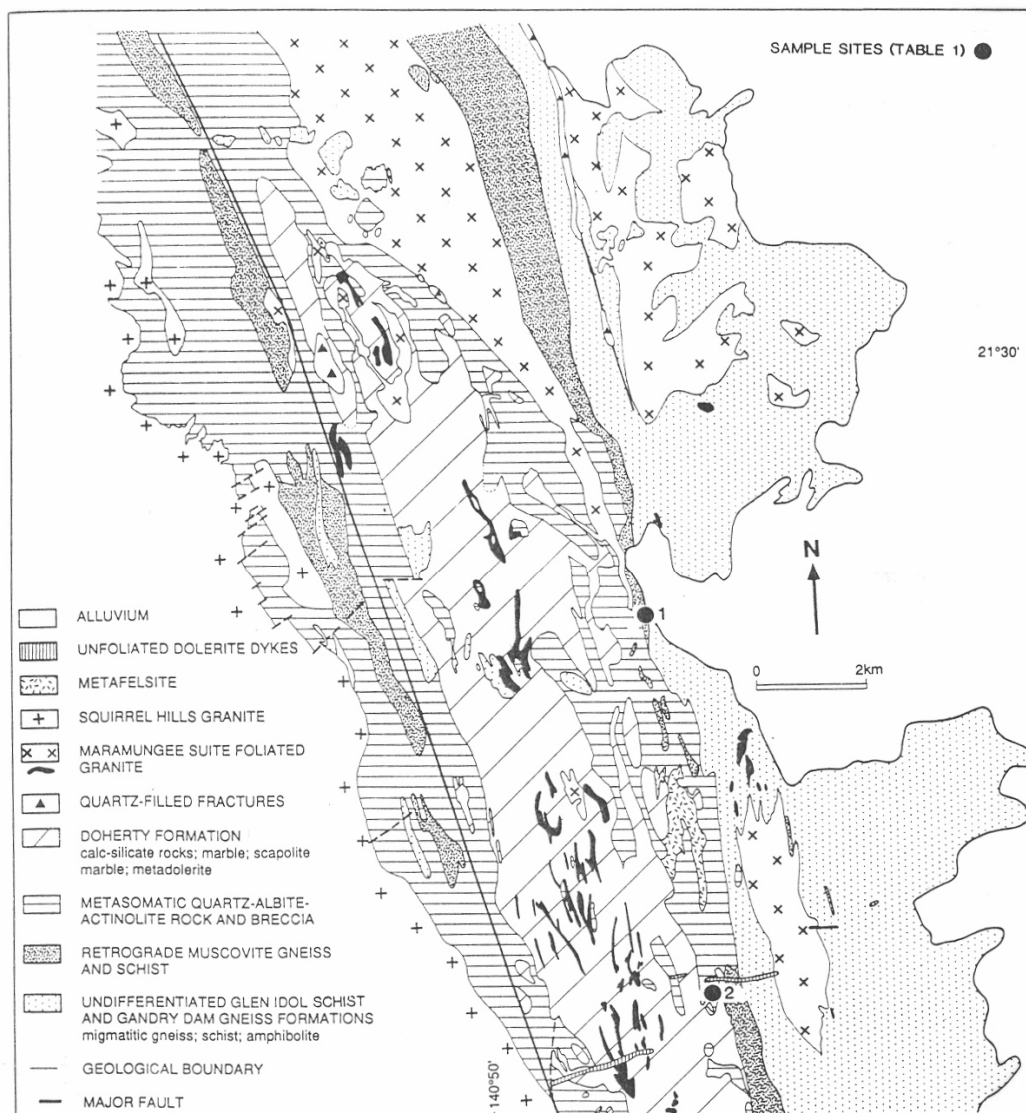
## Regional Na-Ca-Cl alteration

Many iron-oxide Cu-Au districts are characterized by the development of regional scale sodic-calcic alteration (see Barton and Johnson, 1986; Frietsch et al., 1997) that is dominated by Na-, Ca- and Cl-rich minerals including albite, actinolite, clinopyroxene and scapolite, with variable amounts of apatite, titanite, epidote, magnetite, hematite and sulphides. Sodic(-calcic) alteration occurs in association with a wide variety of brittle and ductile structures, and in some cases there is evidence for multiple, overprinting stages of alteration. The most common types of structural controls on alteration include: brittle-ductile shear zones (Ettner et al. 1993), ductile fabrics (cleavages) linked to regional deformation (De Jong and Williams 1995; Oliver 1995; Rubenach and Barker 1998), breccia complexes (De Jong and Williams 1995; Mark 1998), and brittle fracture systems (veins).

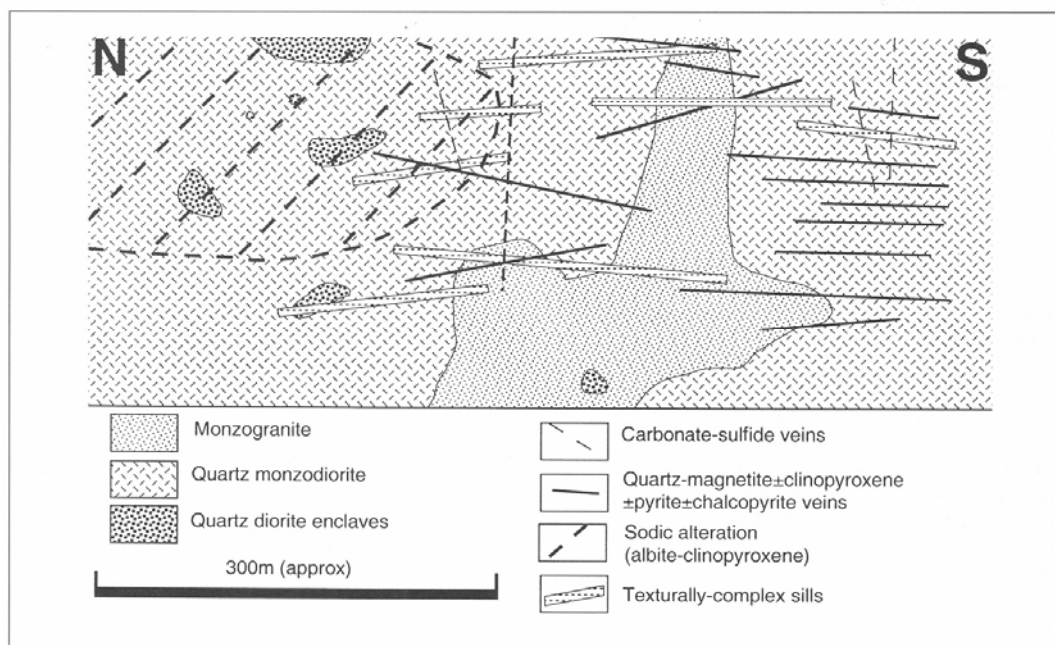
In the Cloncurry district, large volumes of sodic(-calcic) alteration are associated with the formation of breccia complexes that are particularly well-exposed along the Cloncurry fault (e.g. De Jong and Williams 1995; Mark 1998; Fig. 7). These breccias occur predominantly in the roof zones and margins of the Williams and Naraku Batholiths, and at least in part, were developed in multiple episodes in response to successive granite intrusions (Mark 1998). Breccia zones from tens to hundreds of metres across commonly consist of clasts of albitized host rocks (metasediment, calc-silicate, granite, and amphibolite) in an altered matrix composed of sodic(-calcic) minerals. The clast size varies considerably in the breccias, with large clasts (metres across) being mainly sub-angular to angular, while the smaller clasts are commonly more rounded. The presence of rounded clasts of mixed rock types in an altered matrix suggests that the sodic(-calcic) breccias are in part, hydrothermal intrusive breccias. Brecciation of this type is commonly formed by upward escape of a magmatic vapour phase that promotes fracturing and milling of the roof rocks of the intrusion, with subsequent alteration of clasts and matrix due to later ingress of salt-rich magmatic-hydrothermal fluids (e.g. Grant et al. 1980).

Pervasive clinopyroxene-albite±titanite alteration overprints monzodiorite and monzogranite at the Lightning Creek prospect in the Cloncurry district (Perring et al. 2000). No channelways for infiltration of fluids have been observed associated with this alteration, and fluid movement appears to have been controlled predominantly by grain boundaries and microfractures. Quartz and feldspar from the sodic(-calcic) altered granites have essentially identical oxygen isotope compositions to those in the unaltered granites, suggesting that the fluid responsible for alteration was hot (550-650°C) and of similar isotopic composition to primary magmatic fluids (Perring et al. 2000). Subhorizontal granitic sills which crosscut the pervasive sodic(-calcic) alteration at Lightning Creek display a transition from magmatic to hydrothermal crystallization that resulted in albitization of the sills and development of magnetite-clinopyroxene-albite-quartz-apatite-titanite veins containing minor amounts of pyrite (Perring et al. 2000).

Timing evidence from the relationships between alteration, metamorphism, deformation and granites in the Cloncurry district indicates that alteration was developed over approximately the same 40 million year period as the granites. For example, Rubenach and Barker (1998) suggest that sodic alteration commenced during peak metamorphism during D<sub>2</sub> which was at approximately 1545 Ma, while Page and Sun (1998) report the presence of ca.1500 Ma hydrothermal zircons in the Malakoff and Levian Granites near Cloncurry, which they interpret to indicate the time of regional sodic-calcic alteration in that area.



**Figure 7a.** Geological map of part of the Cloncurry district showing the distribution of rocks with intense alkali-metasomatism. From Williams, 1994.



**Figure 7b.** Schematic cross section of the Lightning Creek prospect (see Williams and Pollard, 2001).

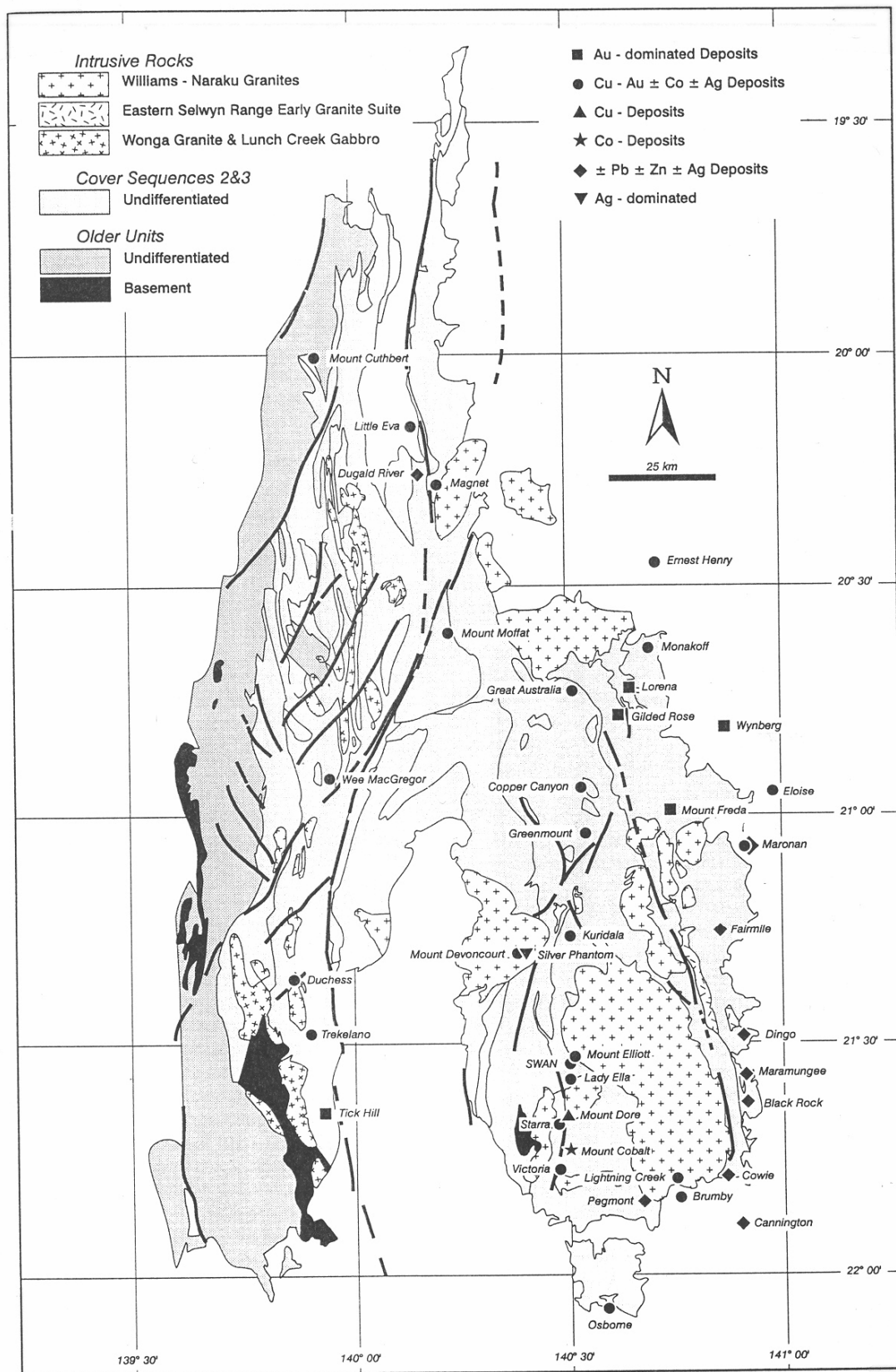
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## Structural and host rock controls on mineralization

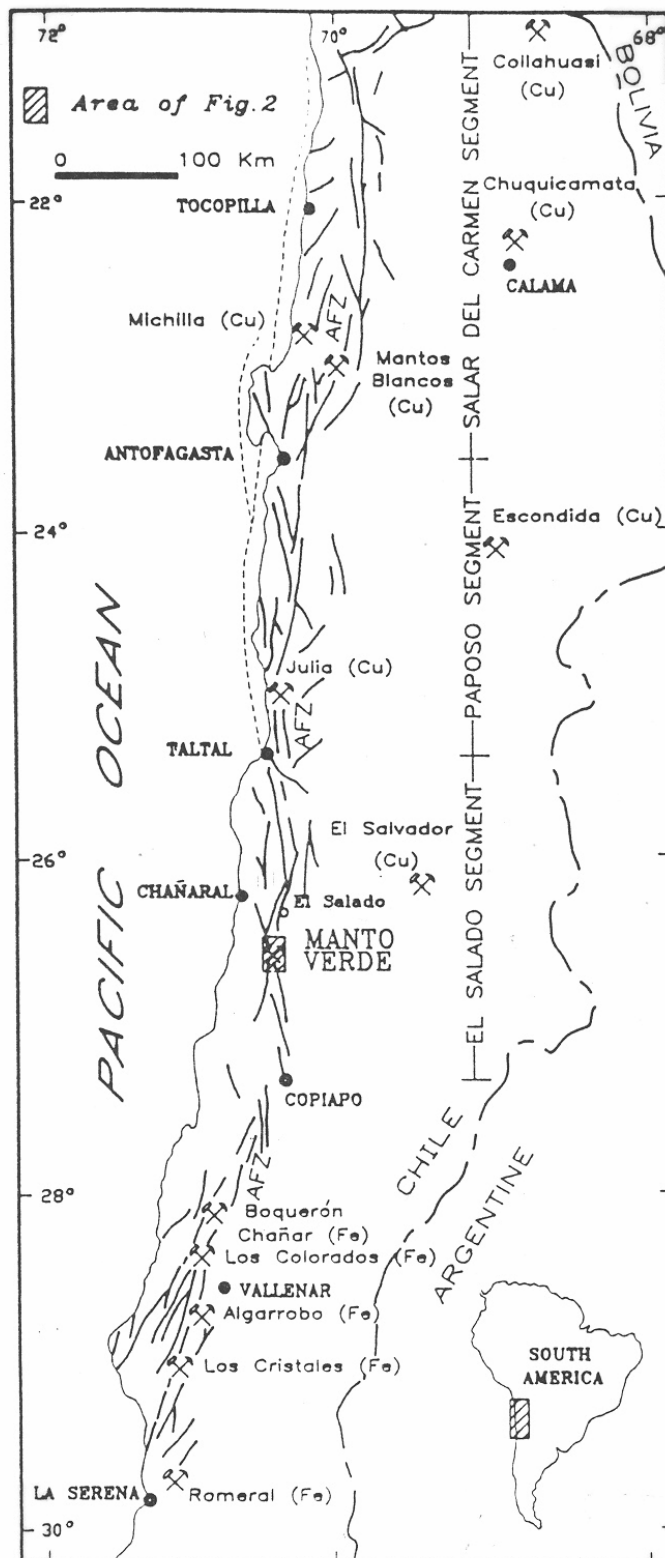
Mineralization in IOCG deposits is commonly controlled by regional scale structures which were active during mineralization. In deeper level environments these structures commonly had both brittle and ductile components, and in some cases represent reactivation of older ductile shear zones. Mineralization typically occurs in second order structures near the main regional structures. An excellent example of this is the Mount Dore fault system in the Cloncurry district which extends N-S for over 100 km in a corridor of mineralization that includes a number of medium- to small-scale deposits (e.g. Starra, Mount Dore, Lady Ella, Mount Elliott, Kuridala, Greenmount, Great Australia, Fig. 8). Similarly, the Atacama Fault Zone in northern Chile is a major focus for IOCG and magnetite-apatite mineralization over a distance of approximately 1000 km from La Serena to Tocopilla (Fig. 9), the deposits themselves are commonly localized by second order structures.

At the local scale, mineralization occurs in a wide variety of structural settings that include features such as fault jogs, fault intersections, intersections of faults and lithological contacts or brittle lithologies, and subsidiary structures linking strands of major fault systems (Fig. 10). Hydrothermal intrusive breccias, break-up breccias, crackle zones and veins are the most common local fluid channelways.

In most cases it appears that the fluids responsible for mineralization originally consisted of H<sub>2</sub>O-CO<sub>2</sub>-salt mixtures with very high salinities. Copper mineralization may be linked to mixing of this fluid with a fluid containing sulphur, or interaction with sulphur-bearing rocks (Oliver et al., 2004). In some deposits there is good evidence to suggest that certain types of host rocks may have exerted a chemical control on ore formation. Deposits at Tennant Creek, Northern Territory (Huston et al., 1993) and the Starra deposits in the Cloncurry district of northwest Queensland (Rotherham, 1997) appear to have formed where metasomatic magnetite ironstones caused reduction of oxidized magmatic fluids resulting in copper and gold precipitation. Interaction of hydrothermal fluids with graphitic metasediments may similarly result in fluid reduction and metal precipitation (e.g. Ettner et al., 1993; Frietsch et al., 1997).

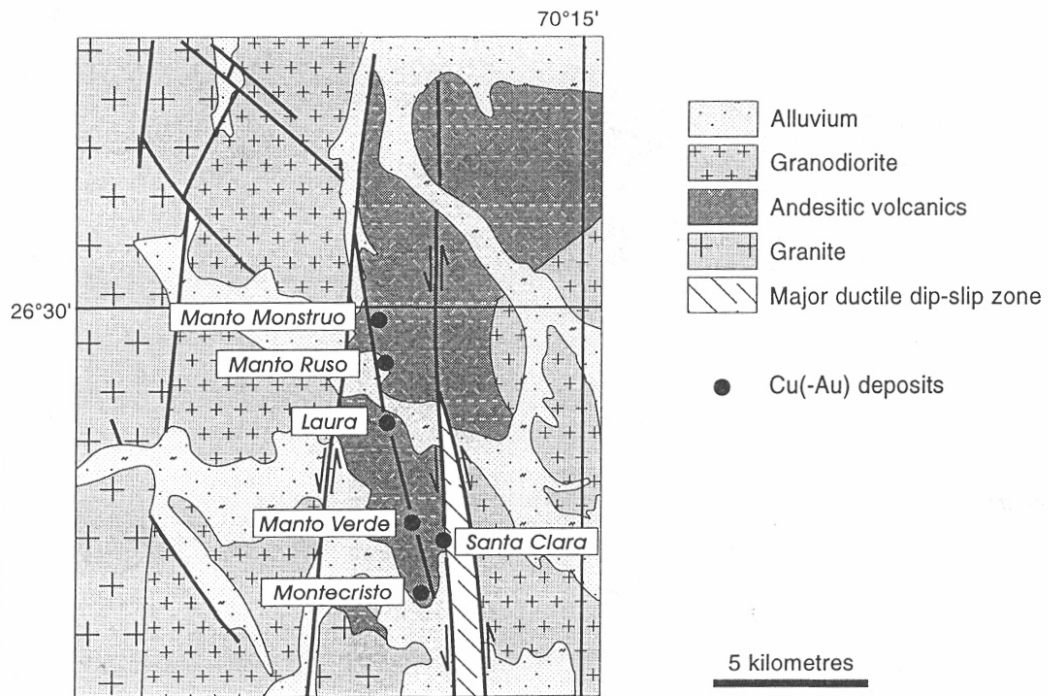


**Figure 8.** Map of the Cloncurry district of northwest Queensland showing the distribution of major faults and mineral deposits. From Williams, 1998.



**Figure 9.** Map of part of northern Chile showing the Atacama Fault Zone and the distribution of some of the copper and iron deposits (from Vila et al., 1996).





**Figure 10.** Map of the Manto Verde area illustrating the major faults and the location of Cu-Au deposits (from Vila et al., 1996).

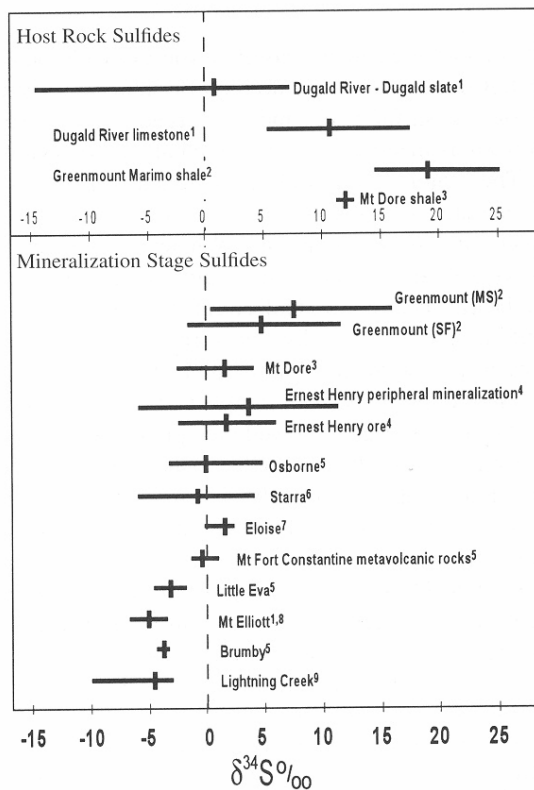
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## Source(s) of fluids and metals in IOCG deposits

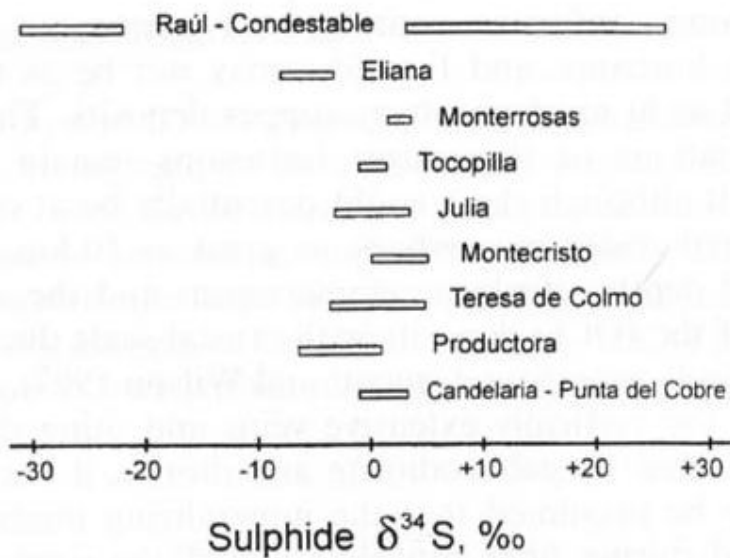
Mineralization in IOCG deposits is commonly linked to potassic styles of alteration represented by biotite and/or K-feldspar. Detailed fluid inclusion and stable isotope studies at several deposits have indicated a dominant magmatic component to the fluids responsible for mineralization (e.g. Adshead, 1996; Lindblom, 1996; Rotherham et al., 1998; Pollard, 2000, 2001). This has been substantiated in the Cloncurry district where fluids derived during crystallization of graphic and spherulitic textured granitic sills associated with magnetite-clinopyroxene-albite-(pyrite-chalcopyrite) veins have been shown to contain greater than 1% Cu (Perring et al., 2000). The huge amount of magnetite contained in the associated vein system is consistent with large volumes of copper-rich fluids having been channelled out of the granites to form mineralization elsewhere.

Another potential source of metals is the regional scale alteration systems associated with IOCG mineralization. Williams (1994), Mark (1998) and Oliver et al (2004) have shown that the fluids associated with sodic(-calcic) alteration leached metals from certain types of host rocks, and would therefore have had the potential to contribute to alteration, Fe-Cu-Au and possibly base metals mineralization at higher crustal levels.

The commonly low-sulphur character of the iron-oxide Cu-Au deposits raises questions about the sulphur contents of the fluids and the source(s) of sulphur in these systems. As noted above, extremely Cu-rich fluids were generated during granite crystallization at Lightning Creek in the Cloncurry district, but an economic copper(-gold) deposit was not formed. Was the fluid simply too hot for crystallization of chalcopyrite, or was there insufficient sulphur in the fluid for crystallization of abundant sulphides? Sulphur isotope data for the ore deposits are generally consistent with a dominantly magmatic source for sulphur (e.g. Starra (Rotherham et al., 1998), Eloise (Baker, 1996), Aitik (Freitsch et al., 1997), Tennant Creek (Huston et al., 1993), Candelaria (Ullrich and Clark, 1999; Marschik and Fontboté, 2001; Sillitoe, 2003, Figs. 11a and 11b). Oliver et al. (2004) suggest that sulphur could be derived from basic magmas which commonly accompany felsic magmatism in IOCG provinces. A role for sulphur from non-magmatic sources has also been proposed for Olympic Dam (Haynes et al., 1995) and late stage fluids at Candelaria (Ullrich and Clark, 1999).



**Figure 11a.** Means and ranges of sulphur isotope data for Cloncurry ore systems and host rocks (from Williams and Pollard, 2001).



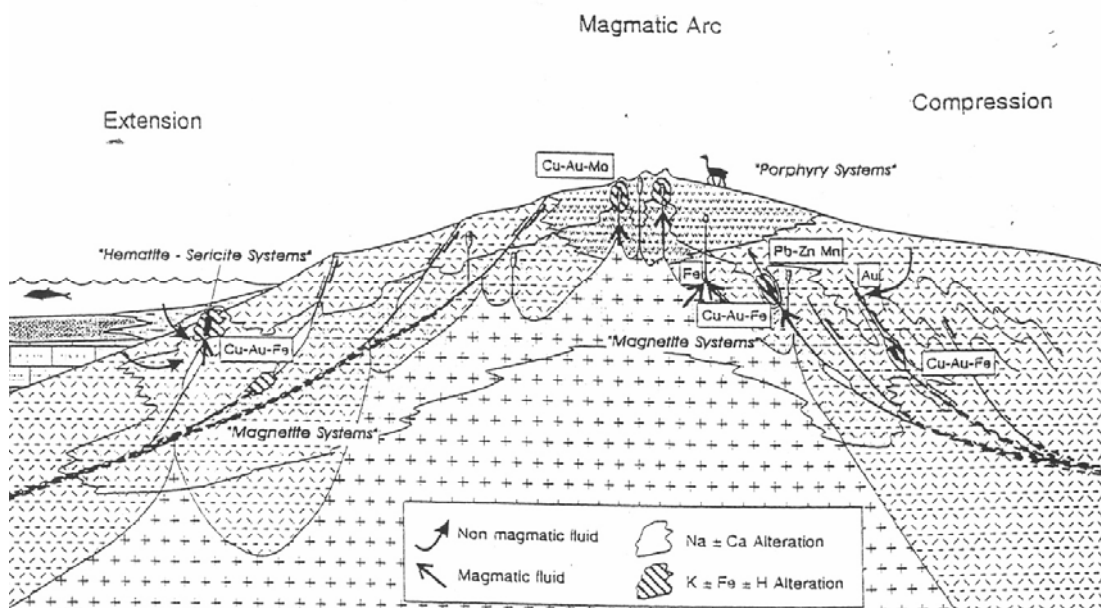
**Figure 11b.** Sulphur isotope data for IOCG deposits in Chile and Peru (from Sillitoe, 2003).

## Links to other mineralization styles

There is strong circumstantial evidence based on spatial and temporal relationships for links between the IOCG deposits and other styles of mineralization (shown schematically in Fig. 12). In some districts there is a close association with magnetite-apatite deposits (e.g. Kiruna district, Sweden and the Coastal Batholith, Chile; Brookstrom, 1977; Hitzman et al., 1992). Indeed, some of the magnetite-apatite deposits contain minor Cu-Au mineralization and have very similar alteration to that associated with the Cu-Au deposits.

In some Proterozoic IOCG districts there is an association with Mn-Pb-Zn-Ag deposits, including “Broken Hill Type” deposits (e.g. Cannington, Australia). It is likely that the formation of the Cannington Ag-Pb-Zn deposit in the Cloncurry district was related to the granites and associated Cu-Au mineralization (e.g. Williams et al., 1996).

Cretaceous IOCG mineralization in the coastal batholith of northern Chile (e.g. Candelaria; Ryan et al., 1995) occurs in the same belt as porphyry-related Cu(-Au) mineralization of similar age (e.g. Andacollo; Reyes, 1991). As yet, the nature of the links between these two styles of mineralization is not known. Given the association of sodic-calcic alteration with some porphyry Cu-Au deposits, and the very magnetite-rich character of some porphyry-related Cu-Au deposits (e.g. Ridgeway, Australia; Grasberg, Indonesia) these associations may be more than coincidental, and possibly reflect similar alteration processes.



**Figure 12.** Schematic section illustrating possible links between alteration types, magnetite- and hematite-rich IOCG deposits, Au-only deposits, Pb-Zn-Ag deposits and porphyry Cu-Au-Mo deposits.

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Iron-oxide-Cu-Au-(U-REE) deposits – A practical guide  
and update of recent developments

**Part 2. Geological aspects of selected IOCG deposits in  
Australia**

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## Introduction

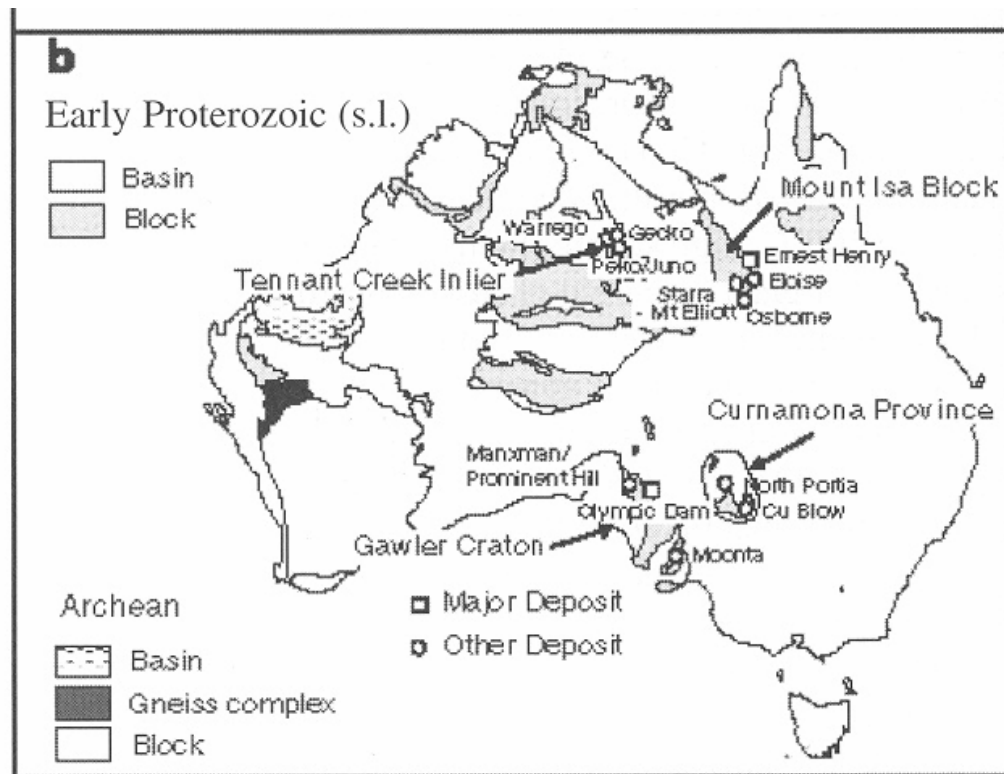
The major known Fe-oxide Cu-Au deposits in Australia are Proterozoic in age, and occur in the Cloncurry district of the Mount Isa Inlier, Queensland, the Gawler and Curnamona Cratons in South Australia and New South Wales, and the Tennant Creek Inlier, in the Northern Territory (Figs. 1-3). The deposits range from the major, breccia-hosted Olympic Dam deposit, to small, high-grade deposits in the Tennant Creek and Cloncurry districts. Examples of both magnetite- and hematite-rich systems are present, and there is a common association of Cu-Au mineralization with F, Ba, LREE and U, as noted elsewhere in Proterozoic deposits.

## Tennant Creek

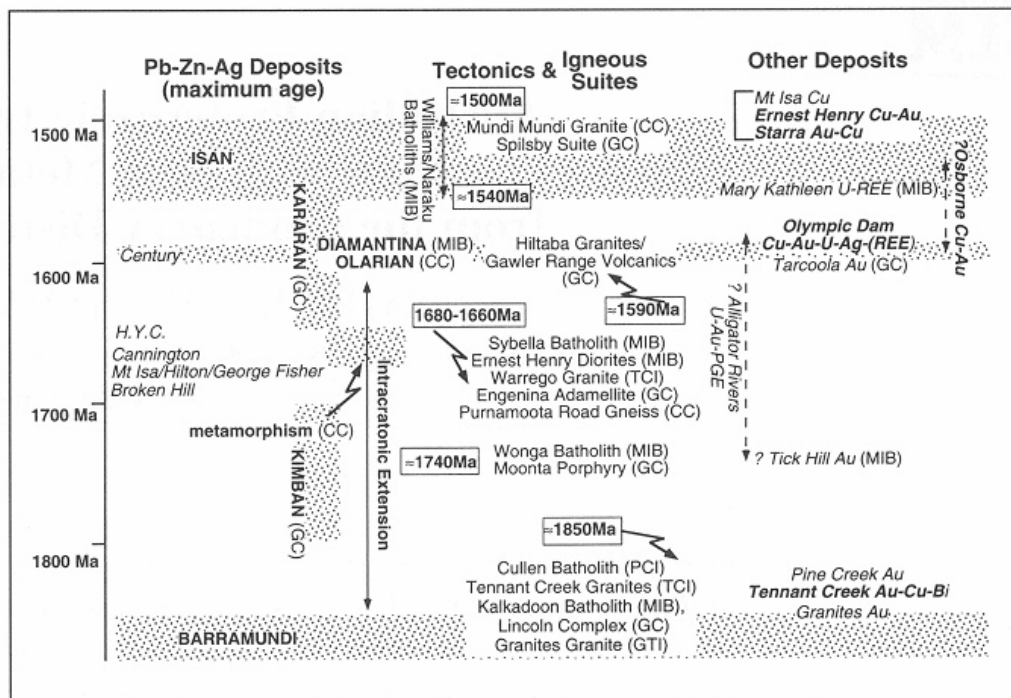
The Tennant Creek block is composed largely of deformed, but rarely metamorphosed Paleoproterozoic (~1860 Ma) greywacke and felsic volcanic rocks. Deformation, mineralization and granitoid emplacement occurred during the Barramundi Orogeny (1870-1840 Ma), shortly after sedimentation. The deformed sequence is unconformably overlain by slightly younger supracrustal rocks which are not significantly affected by the tectonic events that are related to mineralization in the other two Australian Fe-oxide Cu-Au districts.

There are nine significant mines (mostly worked out) in the Tennant Creek Inlier (Fig. 4), and these commonly have Au:Cu ratios higher than deposits in other districts (Huston et al., 1993; Davidson and Large, 1994). About 650 structurally-controlled, metasomatic magnetite-hematite lenses with chloritic alteration halos are concentrated at specific stratigraphic positions (Wedekind et al., 1989; Rattenbury, 1992; Huston et al., 1993). These range in size from a few 100t up to more than 15Mt (Wedekind et al., 1989). Approximately 200 of the known ironstones were selectively affected by a second hydrothermal phase of Cu-Au-Bi mineralization, and in many cases, the ironstones are much larger than the volumes of ore they contain. Compston and McDougall (1994) established that ironstone formation and mineralization occurred over a short time interval, and have a minimum age of 1830-1825 Ma. Coupled with structural constraints, this establishes the close temporal link with the *ca.* 1850 Ma felsic magmatism.

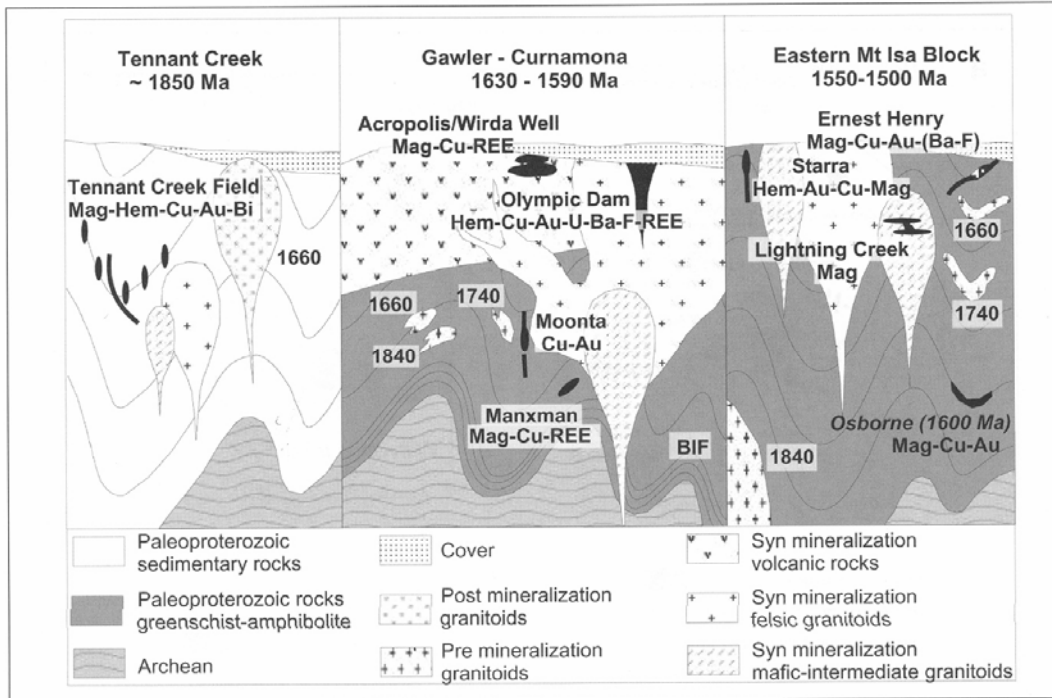
Fluid inclusion studies of the Tennant Creek Cu-Au deposits suggest that the host ironstones formed from low temperature (~250°C), moderate salinity fluids. Later Cu-Au mineralization was formed from hotter (~350°C) and higher salinity fluids with both oxidized and reduced carbonic components (Huston et al., 1993; Khin Zaw et al., 1994). Huston et al. (1993) suggested that variations in fluid composition reflect differing proportions of oxidized magmatic fluids and reduced basinal brines. Skirrow (1999) presented evidence for a mantle-derived component in the ores, implying a connection to primitive magmas as at Olympic Dam. He also suggested that mixing of oxidized and reduced fluids may have operated in conjunction with fluid-ironstone reactions to precipitate ore metals. Skirrow and Walshe (2002) interpreted the reduced deposits to have formed from reduced, sulphur-rich and N<sub>2</sub>-CH<sub>4</sub>-rich low salinity fluids and the more oxidized deposits by mixing of an oxidizing Ca-Na-Cl brine with an intermediate *f*O<sub>2</sub>, low- to moderate salinity ore fluid during reaction with the ironstones.



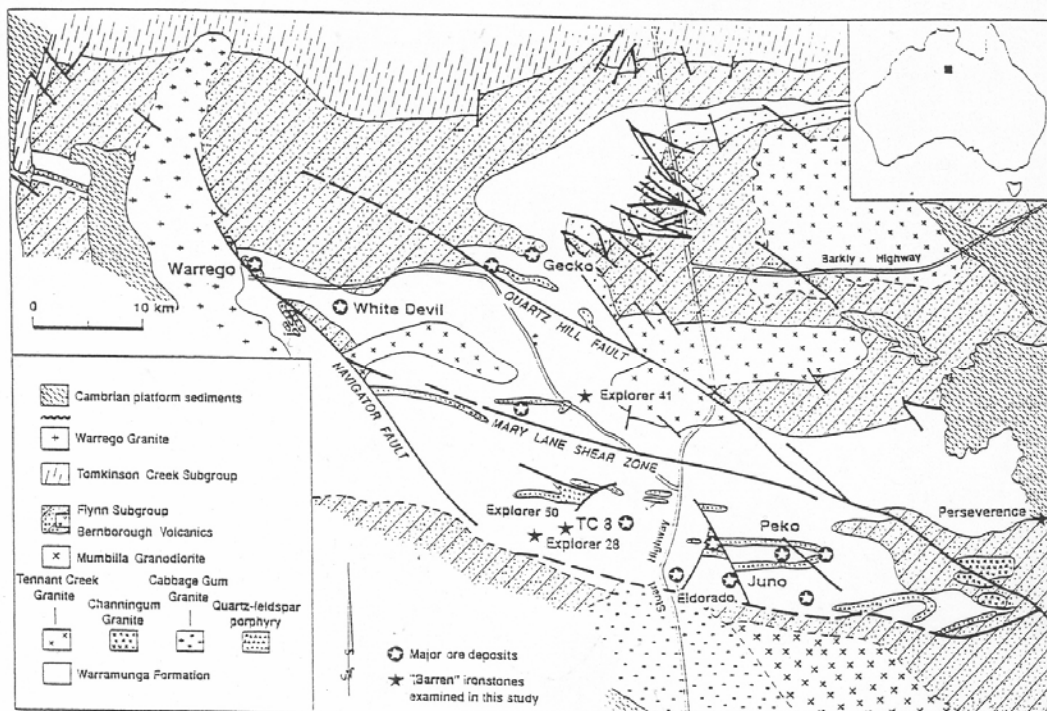
**Figure 1.** Proterozoic blocks with significant Fe-oxide-Cu-Au deposits in Australia (from Williams and Pollard, 2001).



**Figure 2.** Major late Paleoproterozoic and early Mesoproterozoic tectonic, magmatic and metallogenic history of Australia. PCI – Pine Creek Inlier, GTI – Granites-Tanami Inlier, TCI – Tennant Creek Inlier, GC – Gawler Craton, CC – Curnamona Craton, MIB – Mount Isa Block.



**Figure 3.** Cartoon illustrating geological environments of Australian Proterozoic Fe-oxide-Cu-Au deposits (from Williams and Pollard, 2001).



**Figure 4.** Regional geology of the Tennant Creek goldfield and location of major mines and ironstone bodies (from Khin Zaw et al., 1994).

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## Gawler Craton and Stuart Shelf

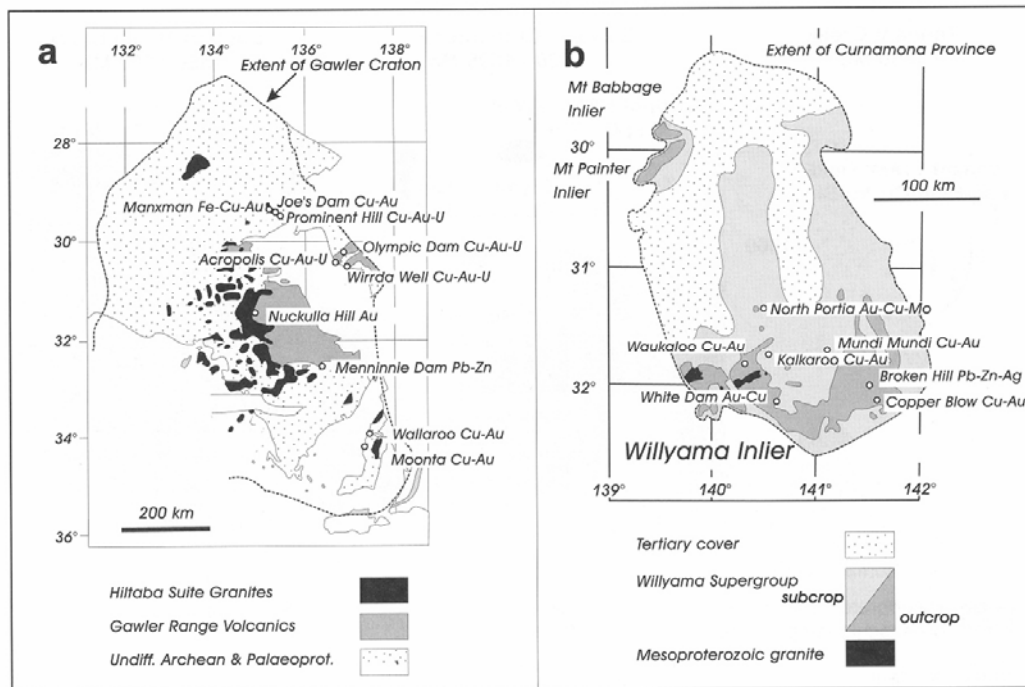
The Gawler and Curnamona Cratons underlie a large part of southern Australia, and are separated by an extensive area of Neoproterozoic supracrustal rocks that were deformed in the early Paleozoic Adelaide Fold Belt (Fig. 1, 5). Large areas of the cratons are concealed by Neoproterozoic and younger cover, but much is thin enough to allow geophysical exploration of the basement. Archean rocks are quite extensively exposed in the west and north of the Gawler Craton, and extend as far east as Broken Hill (Nutman and Ehlers, 1998a). Major banded iron formation containing abundant magnetite occurs locally within the older parts of the Paleoproterozoic sequence (Parker et al., 1993).

Granitoid emplacement in the Gawler and Curnamona Cratons occurred at a number of times (Fig. 2) including at around 1850 Ma (cf. Tennant Creek), 1740 Ma, 1690 Ma, and 1660 Ma (Parker et al., 1993; Daly et al., 1998; Nutman and Ehlers, 1998b). Three compressional deformation events occurred in the southern part of the craton from 1845-1700 Ma in what is referred to as the Kimban Orogeny (e.g. Daly et al., 1998). High grade metamorphism is deduced to have occurred at least locally in association with the 1690 Ma magmatic event in the Curnamona craton (Nutman and Ehlers, 1998b). However, it appears that the Fe-oxide Cu-Au deposits are discreetly associated with the emplacement of the Hiltaba suite granitoids at around 1590 Ma, a timing relationship that is tightly constrained at the Olympic Dam deposit (Johnson and Cross, 1995). Regional metamorphism ranging up to granulite facies occurred in the Curnamona craton at about the same time, during an event that has been termed the Olarian Orogeny.

A major feature that distinguishes the Gawler Craton from the other Australian Fe-oxide Cu-Au districts is the extensive development of a sequence of volcanic rocks (Gawler Range Volcanics) coeval with the granitoids associated with mineralization. These dominate the cratonic associations near Olympic Dam, and host additional Fe-oxide Cu-Au deposits such as Acropolis and Wirrda Well (Blissett et al., 1993; Cross, 1993). This suggests that depth of erosion into the critical rock associations over large areas of the craton is relatively shallow compared to the other districts.

The resource inventory of the Gawler and Curnamona cratons is dominated by the giant deposit at Olympic Dam (Reeve et al., 1990) and the major Prominent Hill deposit, though the region also contains a number of other significant occurrences and is fairly immature from an exploration perspective. A large number of Cu, Au and U-bearing magnetite and/or hematite bodies have been discovered in the basement to the Stuart Shelf south of Olympic Dam, including Acropolis, Wirrda Well, Oak Dam East and Emmie Bluff (Cross, 1993; Gow et al., 1994; Newton, 1996). To the northwest, the Mount Woods Inlier near Coober Pedy contains several very large magnetite bodies with Cu-Au-U-REE mineralization, such as the examples at Manxman A and the Joes Dam area (Newton, 1996; Hampton, 1997). The latter are hosted by granitoids and high grade metamorphic rocks and are characterized by early sodic-calcic alteration assemblages (plagioclase-diopside). The historically significant Cu-Au occurrences of the Moonta-Walleroo area at the northern end of Yorke Peninsula are characterized by various styles of chalcopyrite  $\pm$  magnetite  $\pm$  pyrite  $\pm$  pyrrhotite  $\pm$  hematite mineralization (Conor, 1996; Daly et al., 1998).

Further east, the Curnamona Craton contains many Cu-Au  $\pm$  Mo occurrences that are commonly associated with Fe-oxides (Fig. 5). These deposits are unified by several characteristics, including an abundance of Fe-oxides in some parts of the hydrothermal systems, potassic alteration, syn- to late-tectonic timing (1630-1580Ma?), moderate temperatures of formation ( $\sim$ 300-450°C), and involvement of both hypersaline and carbonic fluids (Williams and Skirrow, 2000).



**Figure 5.** Geology and selected mineral deposits of the Gawler and Curnamona Cratons (from Williams and Pollard, 2001).

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## Olympic Dam Cu-U-Au deposit

The Olympic Dam Cu-U-Au deposit (Fig. 6) is hosted by the Olympic Dam Breccia Complex which occurs in the Roxby Downs Granite in the northeast Gawler Craton, South Australia. The deposit contains an inferred resource of 2000 million tonnes with 1.6% Cu, 0.6 kg/t U<sub>3</sub>O<sub>8</sub>, 0.6 g/t Au and 3.5 g/t Ag (Reeve et al., 1990).

The breccia complex is composed of fractured, brecciated and altered granite, with minor tuffs and sediments. The breccias and sediments display fabrics and textures indicating that they were formed in a high-level, subvolcanic environment characterized by phreatic and phreatomagmatic activity (Reeve et al., 1990, Oreskes and Einaudi, 1990). Mineralization typically occurs as fragments and/or matrix replacement in breccias and as vein infill in granite and breccias. Reeve et al. (1990) divided the breccias into granite-rich and hematite-rich types, with granite-rich breccias being dominated by granite fragments, and hematite-rich breccias being composed of mixed fragments that include granite, hematite, fine-grained mafic and felsic intrusives, sediments, vein quartz, chalcopyrite, fluorite, carbonate and breccia.

The principal minerals in the Olympic Dam deposit include hematite, sericite and chlorite. Magnetite is locally abundant in association with pyrite±chalcopyrite and siderite. Copper sulphides, copper-iron sulphides, uranium minerals, silver and gold are usually intimately associated and mostly occur within hematite-rich breccia bodies and in close association with hematite-rich matrices and hematite alteration in granite-rich breccias (Haynes et al., 1995).

Sulphide mineralization is zoned from chalcocite to bornite to chalcopyrite to pyrite, with chalcopyrite- and pyrite-bearing zones usually developed beneath and fringing the bornite-bearing zones (Roberts and Hudson, 1983, Haynes et al., 1995). Barren hematite breccias and quartz-hematite breccias occur above, or adjacent to chalcocite-bearing zones.

Haynes et al. (1995) recognize three distinctive mineral associations at Olympic Dam: 1) early magnetite±hematite, chlorite, sericite, siderite, pyrite, chalcopyrite and pitchblende, 2) hematite, sericite, chalcopyrite, bornite, pitchblende, barite, fluorite and chlorite, and 3) late, porous, vuggy or massive hematite, quartz, barite and local silicification and quartz veining. There is a gross vertical zonation from association one at deeper levels to association three at shallow levels (Haynes et al., 1995).

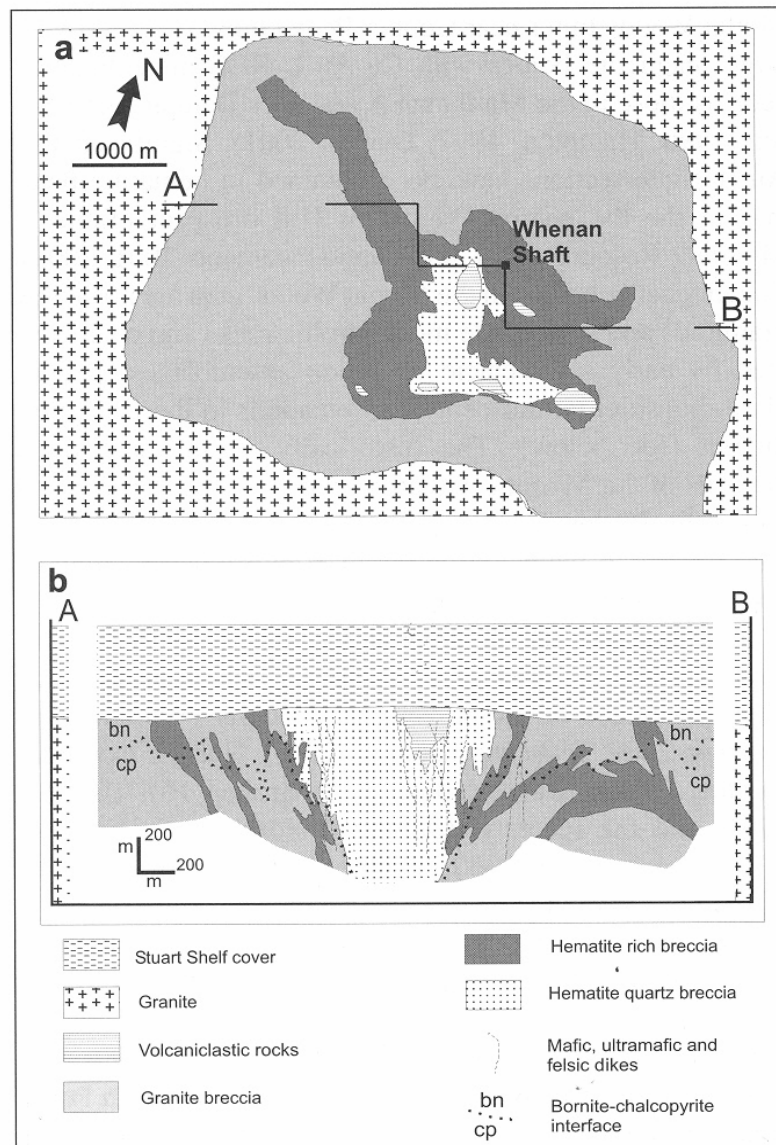
Oreskes and Einaudi (1992) observed that liquid-vapor fluid inclusions associated with hematite-rich breccias contained between 7 and 23wt.% NaCl equivalent and homogenized between 130-280°C, but mostly in the range 170-190°C. Inclusions associated with early magnetite, pyrite and siderite were higher temperature (~400°C). Interpretation of microthermometric data suggests the fluids contain a variety of components including H<sub>2</sub>O, NaCl, KCl, CaCl<sub>2</sub>, FeCl<sub>2</sub>, CaF and CO<sub>2</sub> (Oreskes and Einaudi, 1992).

Fluid inclusion data, together with stable isotope data were interpreted to indicate that two different fluids were involved in formation of the mineralization, a deep-seated, possibly magmatic fluid, and a lower temperature, surface-derived fluid (Oreskes and Einaudi, 1992). The high temperature, magmatic fluid may have evolved via unmixing

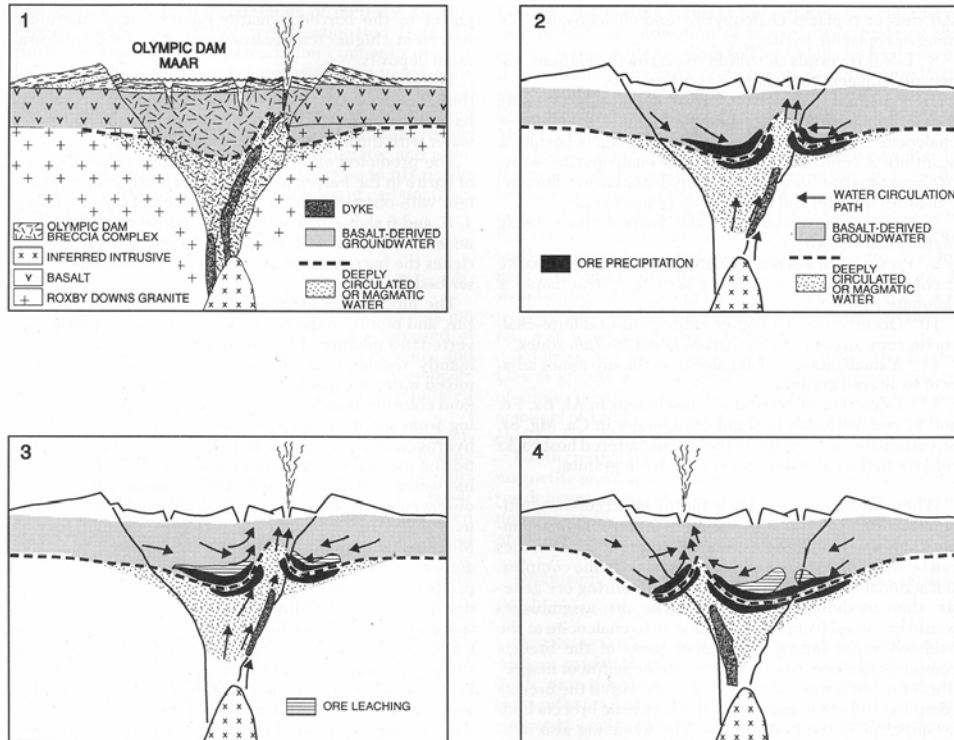


of a CO<sub>2</sub>-rich vapor phase, as both hypersaline and CO<sub>2</sub>-rich fluid inclusions have been identified by Conan-Davies (1987).

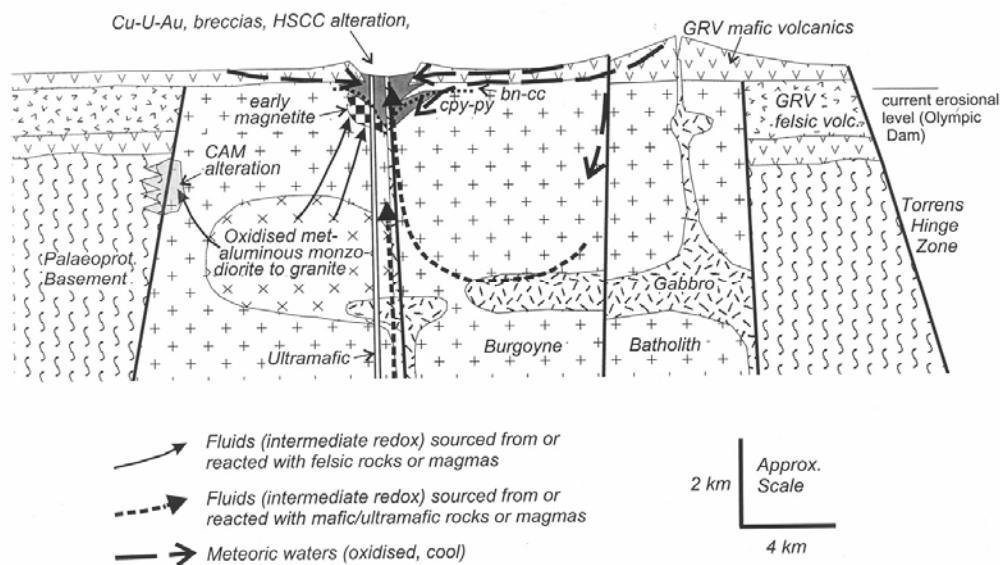
At least two sources of components are suggested by Nd isotope data which indicate that early magnetite-rich assemblages have  $\epsilon\text{Nd}$  similar to the host Roxby Downs Granite, while the hematitic-ores require a contribution from a mantle-derived source rock or magma (Johnson and McCulloch, 1995). Haynes et al. (1995) propose the mixing of the hotter, more reduced possibly magmatic fluid with a lower temperature, more oxidized meteoric fluid may have occurred repetitively in a high level setting (Figs. 7 and 8), with typical mineralizing events evolving from reduced, hotter, more saline conditions to oxidized, cooler, less saline conditions.



**Figure 6.** a. Simplified geologic plan of the Olympic Dam Breccia Complex at about 400m below surface; b. cross section of the Olympic Dam deposit (adapted from Reeve et al., 1990).



**Figure 7.** Model for mineralization via fluid mixing in the Olympic Dam Breccia Complex (from Haynes et al., 1995).

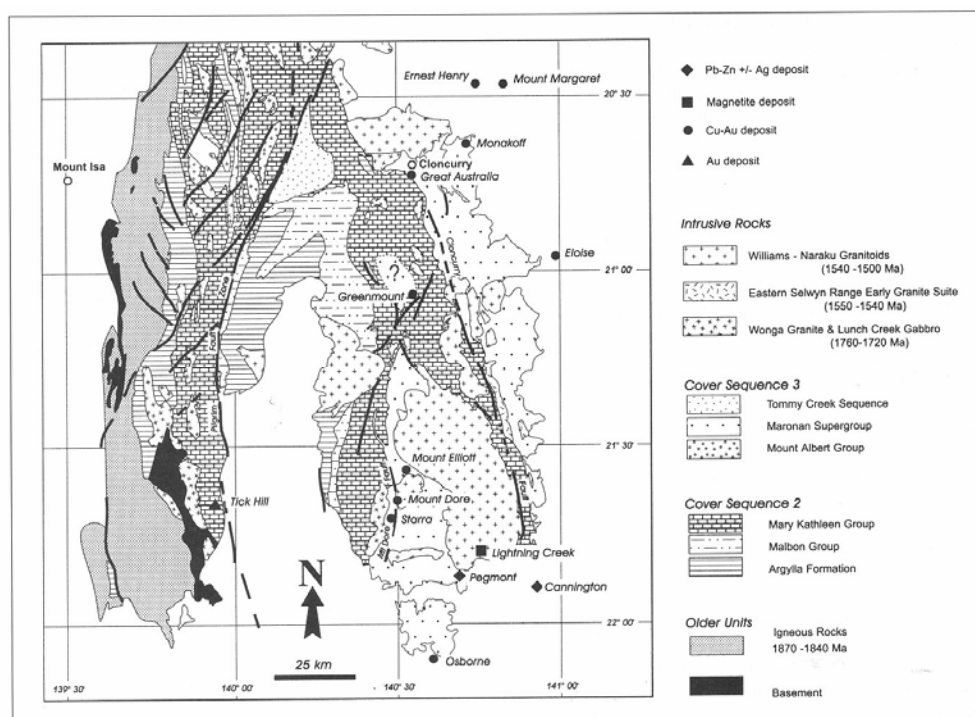


**Figure 8.** Schematic cross section of an Olympic Dam style Cu-U-Au system, showing Cu-U-Au mineralization associated with hematite-sericite-chlorite-carbonate (HSCC) alteration. Deeper level and/or distal calcsilicate-alkali feldspar-magnetite alteration (CAM) and alternative fluid types are also shown (from Skirrow et al., 2002).

## Cloncurry district

The Cloncurry district occupies the eastern part of the Mount Isa Inlier in northwest Queensland (Figs. 9). Young supracrustal rocks near Cloncurry belong to two main sequences which are generally observed in faulted contact with each other. The older sequence deposited between 1780-1720 Ma is a diverse package of metamorphosed basic to felsic volcanic rocks, carbonates, evaporites and siliciclastic sediments. It is dominated by the Mary Kathleen Group in the eastern part of the fold belt. The second late Paleoproterozoic sequence is informally termed the Maronan supergroup (Beardsmore et al., 1988), and consists of metasiliciclastic rocks and subordinate metabasic rocks that appear to have been deposited after 1670-1660 Ma (Page and Sun, 1998). This sequence contains thin, laterally extensive magnetite-quartz “BIF” horizons.

The oldest intrusive rocks in the Cloncurry district have SHRIMP U/Pb ages of approximately 1750 Ma (Pollard and McNaughton, 1997; Page and Sun, 1998). This is similar to the age of intrusive rocks of the Wonga Batholith to the east, and to the age of the Mount Fort Constantine volcanics which are thought to host the Ernest Henry Cu-Au deposit. Adjacent to the Ernest Henry deposit, two diorite intrusions have been dated at approximately 1670 Ma using SHRIMP U/Pb dating of zircon and titanite (Pollard and McNaughton, 1997). These intrusions are equivalent in age to the volcanic components of the Soldiers Cap Group in the Cloncurry district, and the host rocks of the Mount Isa deposit in the Western Succession of the Mt Isa Inlier.

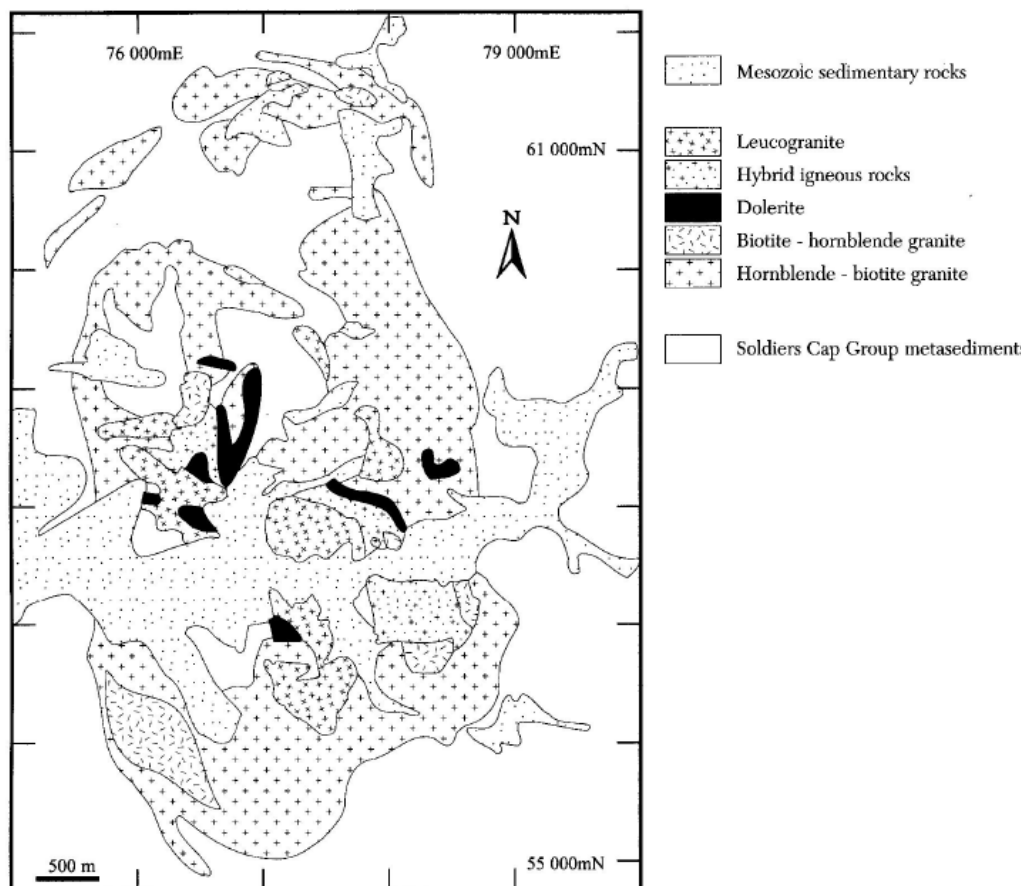


**Figure 9.** Simplified geology of the Cloncurry district (from Williams and Pollard, 2001).

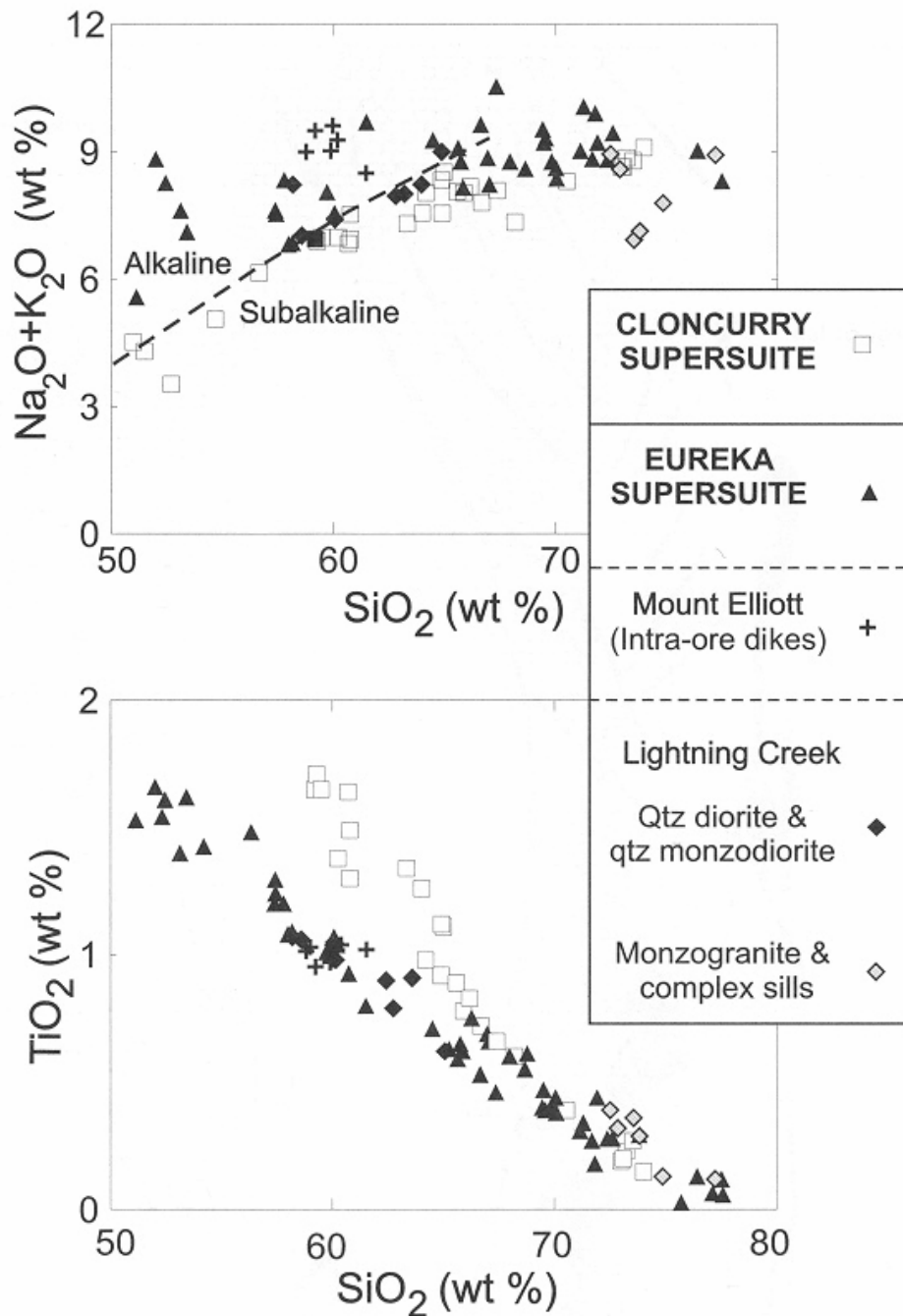
In the southeastern part of the Cloncurry district, close to the Cloncurry fault, a group of intrusions known collectively as the Eastern Selwyn Range early granite suite (Fig. 9) has been dated at approximately 1545 Ma (Page and Sun, 1998). This group of intrusions includes trondjhemitic rocks and was at least partly overprinted by the regional D<sub>2</sub> deformation (see Williams and Blake, 1993).

A major portion of the outcrop in the Cloncurry district is composed of granitoids of the Williams and Naraku Batholiths (Wyborn, 1998; Fig. 9). Most of these rocks were emplaced in the interval 1540-1500 Ma (see below), which coincides with the timing of much of the sodic-calcic alteration, Cu-Au mineralization and D<sub>3</sub> deformation. Detailed mapping (e.g. Fig. 10) and geochemical study of these intrusive rocks has identified two groups of intrusives which are referred to as the Cloncurry Supersuite and Eureka Supersuite (Pollard et al., 1998, Fig. 11).

The Cloncurry Supersuite is composed mainly of granodiorite, monzogranite and syenogranite, and includes parts of the Saxby Granite, Dipvale Granodiorite and Malakoff Granite. The granites are subalkaline, high-K, metaluminous rocks and range from 59-74wt% SiO<sub>2</sub> (Fig. 11). Increasing SiO<sub>2</sub> is accompanied by increasing K<sub>2</sub>O, Rb, and Th, and decreasing MgO, Fe<sub>2</sub>O<sub>3</sub>, FeO, TiO<sub>2</sub>, CaO, P<sub>2</sub>O<sub>5</sub>, Sr, Eu, Zr, Cu, Zn, Ni, V and Sc. This is consistent with fractionation of observed phases including plagioclase, amphibole, biotite, magnetite, titanite and apatite.



**Figure 11.** Simplified geological map of the Wiley igneous complex showing the distribution of intrusive rock types (from Pollard et al., 1998).



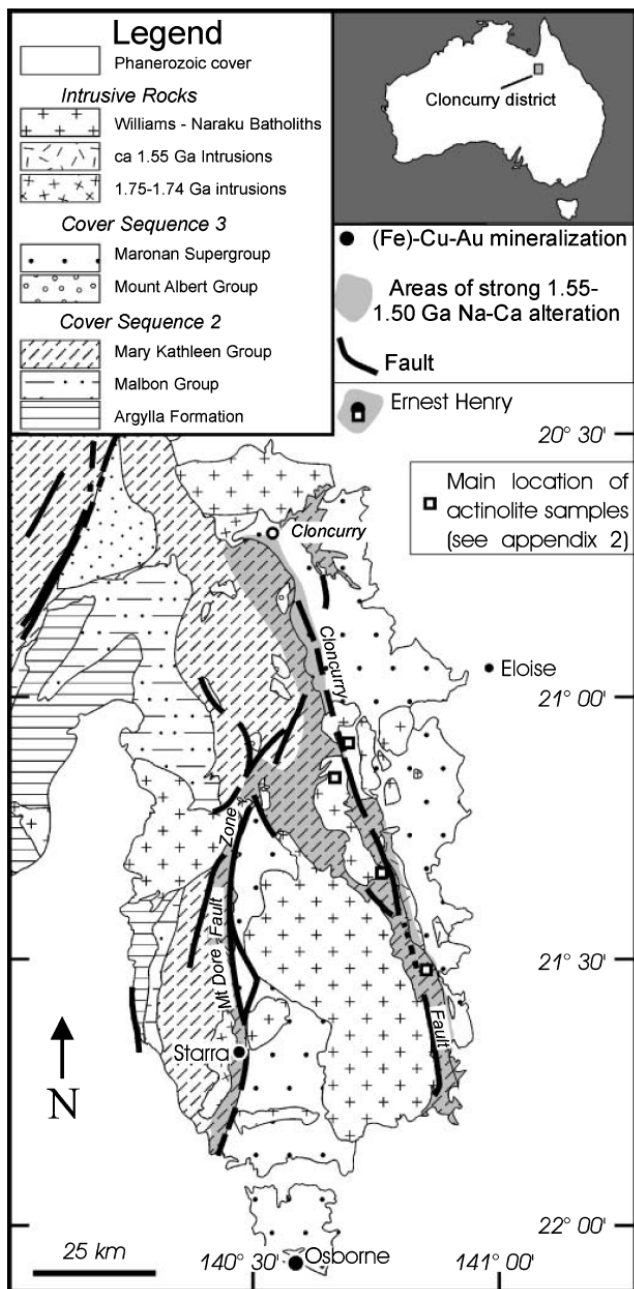
**Figure 12.** Total alkalis- $\text{SiO}_2$  and  $\text{TiO}_2$ - $\text{SiO}_2$  variation diagrams for post-1540 ma intrusive rocks in the Cloncurry district (data from Pollard et al. (1998), Perring et al., (2000) and Wang and Williams (2001)).

The Eureka Supersuite ranges from diorite to syenogranite in composition. More mafic varieties contain pyroxene and hornblende, while more felsic varieties contain biotite as the only ferromagnesian silicate. These rocks are represented in the Mount Angelay igneous complex, Squirrel Hills Granite and Wimberu Granite. Eureka Supersuite intrusive rocks range from 51 to 76 wt%  $\text{SiO}_2$  (Fig. 11). The granites are alkaline and  $\text{K}_2\text{O}$ -rich, leading to their classification as a shoshonite series. Rare-earth element

patterns show enrichment in LREE and negative Eu anomalies that become more pronounced in the more evolved samples. With increasing SiO<sub>2</sub> content, TiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, Fe<sub>2</sub>O<sub>3</sub>(T), MnO, MgO, CaO, P<sub>2</sub>O<sub>5</sub>, Sr, Eu, Cu, Zn, Ni, V and Sc decrease, indicating fractionation of observed minerals including apatite, magnetite, pyroxene, hornblende, plagioclase and biotite. Compared with many other shoshonitic rocks, high-K Eureka Supersuite samples are strongly depleted in Sr and to a lesser extent Eu, and enriched in Y and HREE. This suggests that garnet was not a stable phase in the source region, that the source contained plagioclase as a stable phase, and/or the granites evolved by extensive plagioclase fractionation. The Eureka Supersuite includes monzogranite sills at the Lightning Creek prospect that are associated with the generation of Cu-rich magmatic fluids (see below), and with intra-ore dykes at the Mt Elliott Cu-Au deposit (Fig. 11).

Sodic-calcic alteration occurs over the whole outcrop extent of the Cloncurry district but varies greatly in intensity and only selected areas have been mapped in detail (Oliver, 1995; De Jong and Williams, 1995; De Jong, 1995; Mark, 1998). A major focus of sodic-calcic alteration is in Mary Kathleen Group metasediments (Corella, Staveley and Doherty Formations) which consist of a carbonate-evaporite-clastic sequence containing abundant evidence for development of scapolite during metamorphism. Sodic-calcic alteration also occurs in the vicinity of major fault systems such as the Cloncurry fault, and is a common alteration assemblage within granites of the Williams and Naraku Batholiths (Fig. 12).

Multiple generations of sodic-calcic alteration have been recognized in several areas (Williams and Blake, 1993; Rubenach and Barker, 1998; Mark, 1998), and timing evidence suggests that sodic-calcic alteration extended from syn-D<sub>2</sub> (~1540 Ma?) to ca.1500 Ma, and was episodic in nature. The alteration occurred over the same time period as intrusion of Williams and Naraku Batholith granites, and during the period when Cu-Au mineralization formed. Sodic-calcic alteration is present as a pre-ore alteration stage in all Cu-Au deposits in the region. Some Cu-Au deposits have early pre-ore alteration dominated by albite, with late pre-ore and/or ore-stage assemblages dominated by sodic-calcic minerals, e.g. Mount Elliott (pyroxene, scapolite, actinolite, magnetite, titanite; Wang and Williams, 1997), Brumby (pyroxene, apatite, titanite, magnetite; Foster, 1995), Osborne (albite-actinolite-magnetite; Adshead, 1995), and Lightning Creek (pyroxene, albite, magnetite, quartz, titanite; Perring et al., 2000). This suggests a close relationship between sodic-calcic alteration and Cu-Au mineralization.



**Figure 10.** Major IOCG deposits and distribution of strong sodic-calcic alteration in the Cloncurry district (from Mark et al., 2004).

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## Ernest Henry

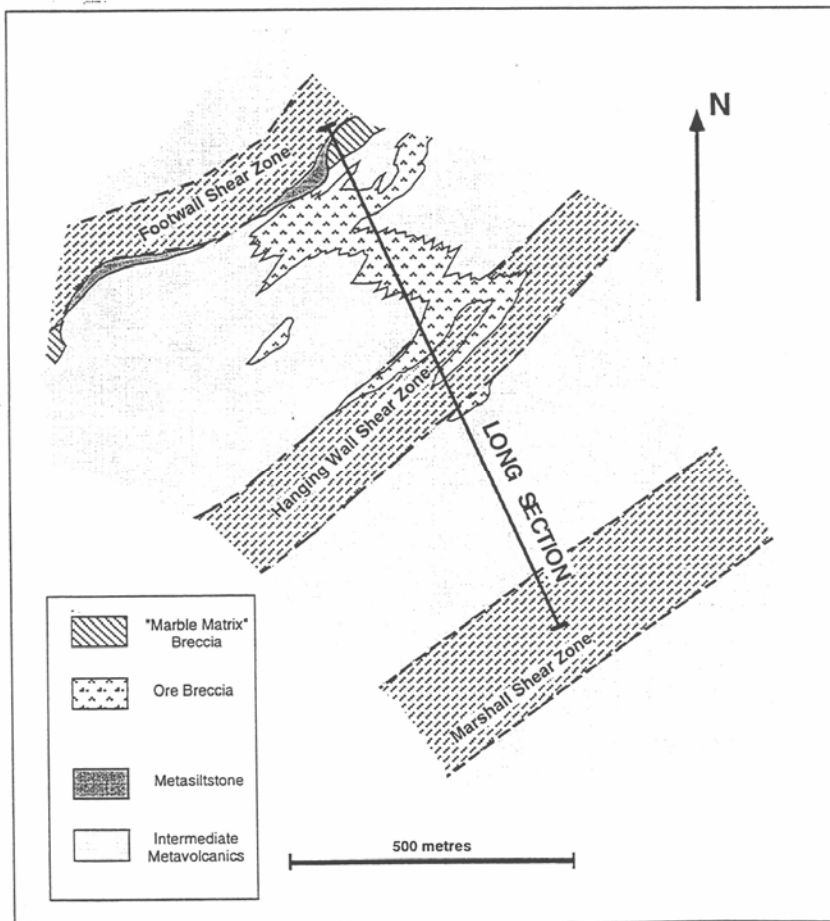
The Ernest Henry mine (Figs. 13 and 14); approximately 40 km NE of Cloncurry is the major Cu-Au deposit in the district. It is a blind deposit, situated some 12 km from the nearest Proterozoic basement outcrop and was discovered by Western Mining Corporation in 1991 by drilling coincident magnetic and electrical features beneath loosely consolidated cover that underlies an extensive black soil plain (Webb and Rowston, 1995). There is a leached cap overlying a zone containing native copper which proved to be the essential source of the TEM anomaly that led to the discovery of the deposit (Webb and Rowston, 1995). The identified resource is 166Mt @ 1.1% Cu and 0.54g/t Au, though the deposit is open at depth and displays no reduction in thickness in the portion that has been explored (Ryan, 1998). Given that some of the deposit was lost by erosion prior to deposition of the Mesozoic cover, this means that the total size may have been substantially larger than the delineated resource.

The dominant structural and geophysical grain at Ernest Henry trends NE and the main fabrics exhibit moderate dip to the SE (Webb and Rowston, 1995; Twyerould, 1997). This is an anomalous structural setting in a regional context. The host rocks to the ore are part of a sequence of strongly altered basic to felsic metavolcanic rocks which may correlate with outcrops further west dated at about 1745 Ma (Page and Sun, 1998). Drilling and geophysical data have delineated extensive dioritic intrusions that are as close as about 500m south of the deposit (Twyerould, 1997; Ryan, 1998). SHRIMP dating suggests these were emplaced at around 1660 Ma (Pollard and McNaughton, 1997).

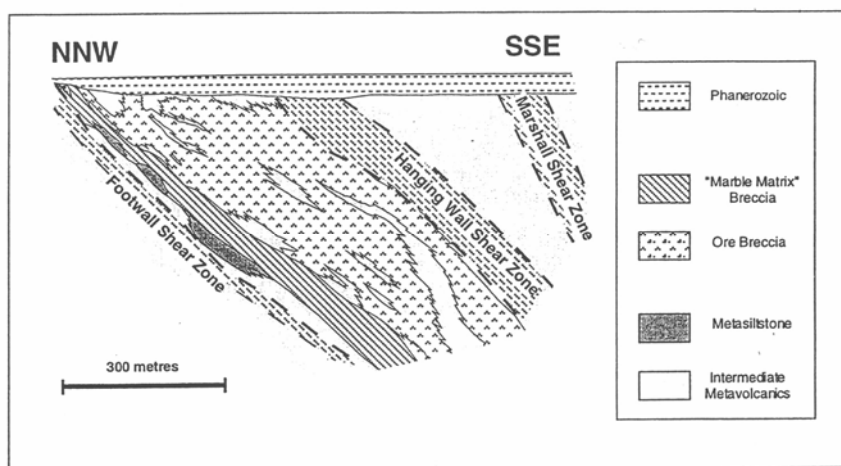
The original character of the metavolcanic rocks near the deposit is obscured by widespread, fine-grained magnetite-biotite alteration, though rocks with porphyritic, amygdaloidal and flow banded textures can be recognized locally. This fine-grained magnetite alteration predates a coarser type of magnetite that occurs with sulphides in the ore (Ryan, 1998). The ore zone is a large breccia body that has been interpreted to be localized between two shear zones characterized by strong fabrics and intense biotite-magnetite-(garnet) alteration (Twyerould, 1997).

The ore breccias contain matrix infill and alteration mineralization with clasts of pervasively K-feldspathized metavolcanic rocks (Twyerould, 1997; Mark and Crookes, 1999). The breccias display a range of textures from incipient cracking, grading to matrix-dominated breccias characterized by the highest ore grades. Some of the breccias have stretched and flattened clasts implying they developed in a brittle-ductile deformation regime. The mineralized matrix in the primary zone consists of chalcopyrite, pyrite and magnetite which are associated with calcite, quartz, biotite and chlorite, along with minor amounts of epidote, allanite, tourmaline, fluorite, REE fluorcarbonates, monazite, barite, apatite, scheelite, arsenopyrite, cobaltite, galena, molybdenite, uraninite and brannerite (Fairall, 1994; Twyerould, 1997; Ryan, 1998). The average magnetite content of the ore is 20-25% (Ryan, 1998). Ar-Ar dating of biotites from Ernest Henry suggests the ore was emplaced at about 1505 Ma (Twyerould, 1997).





**Figure 13.** Plan geology of the Ernest Henry Cu-Au deposit at 60 metres below surface (after Craske, 1995).



**Figure 14.** Long section of the Ernest Henry Cu-Au deposit interpreted from drill hole data (adapted from Craske, 1995).

## Monakoff

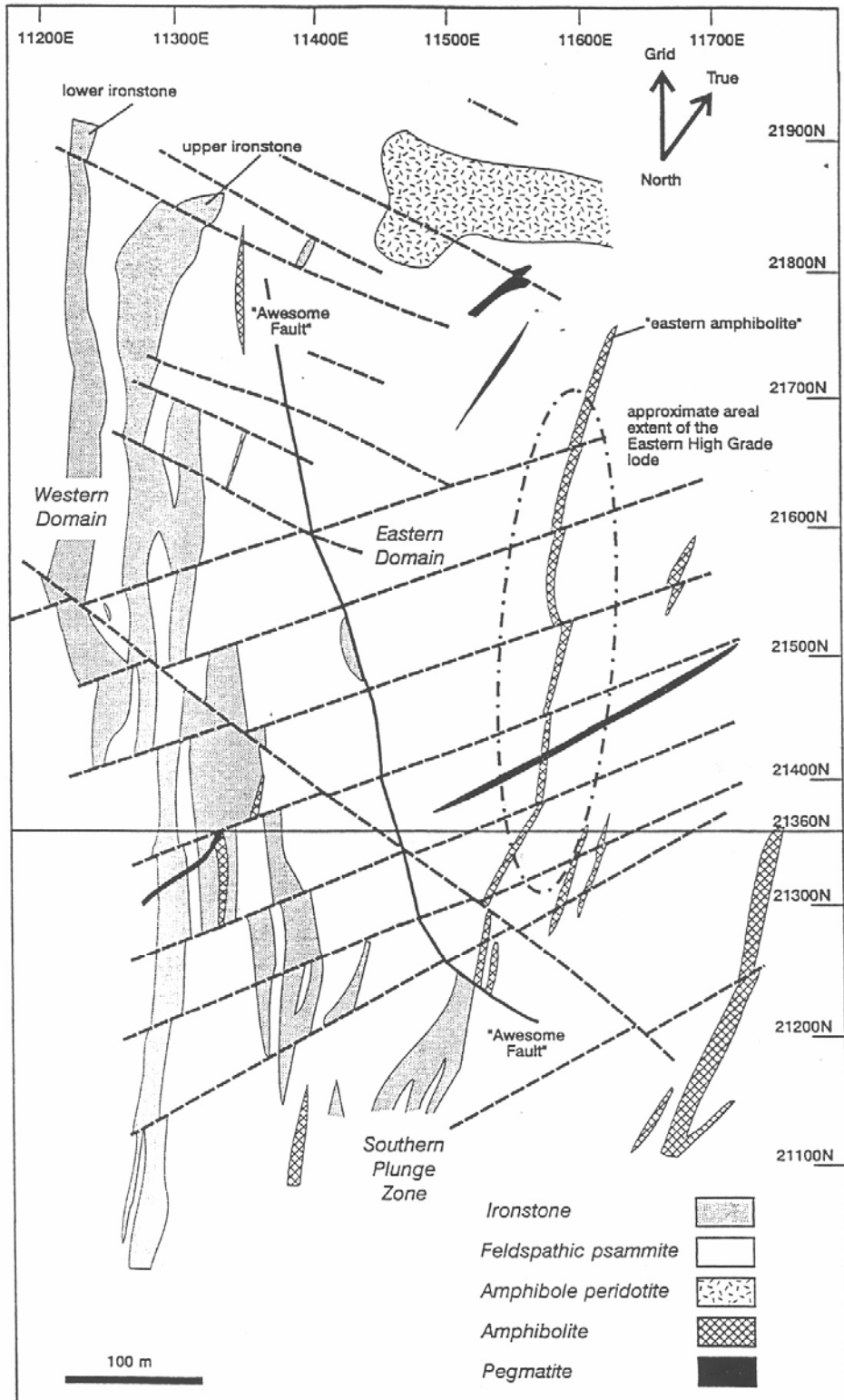
Monakoff is the nearest significant occurrence to Ernest Henry in the exposed Proterozoic basement, and like its much larger neighbor, has a complex Cu-Au-(F-Co-As-Ba-REE-U) geochemical association (Davidson, et al., 2002). Monakoff is hosted by metasediments (quartz-muscovite-plagioclase schists) and consists of foliation-parallel massive siderite-fluorite-barite-magnetite-quartz-sulphide lenses that formed by replacement of rocks with intensified fabrics in a ductile shear zone (Davidson et al., 2002; Laing, 1998). The deposit is localized at a stratigraphic position close to a regionally extensive magnetite iron formation unit that displays a more complex deformation history than the ore (Ryburn et al., 1988; Davidson et al., 2002). Ar-Ar dating of secondary biotite from the deposit gave an age of  $1508 \pm 10$  Ma, similar to those of alteration biotites from Ernest Henry (Pollard and Perkins, 1997; Twyerould, 1997).

## Osborne

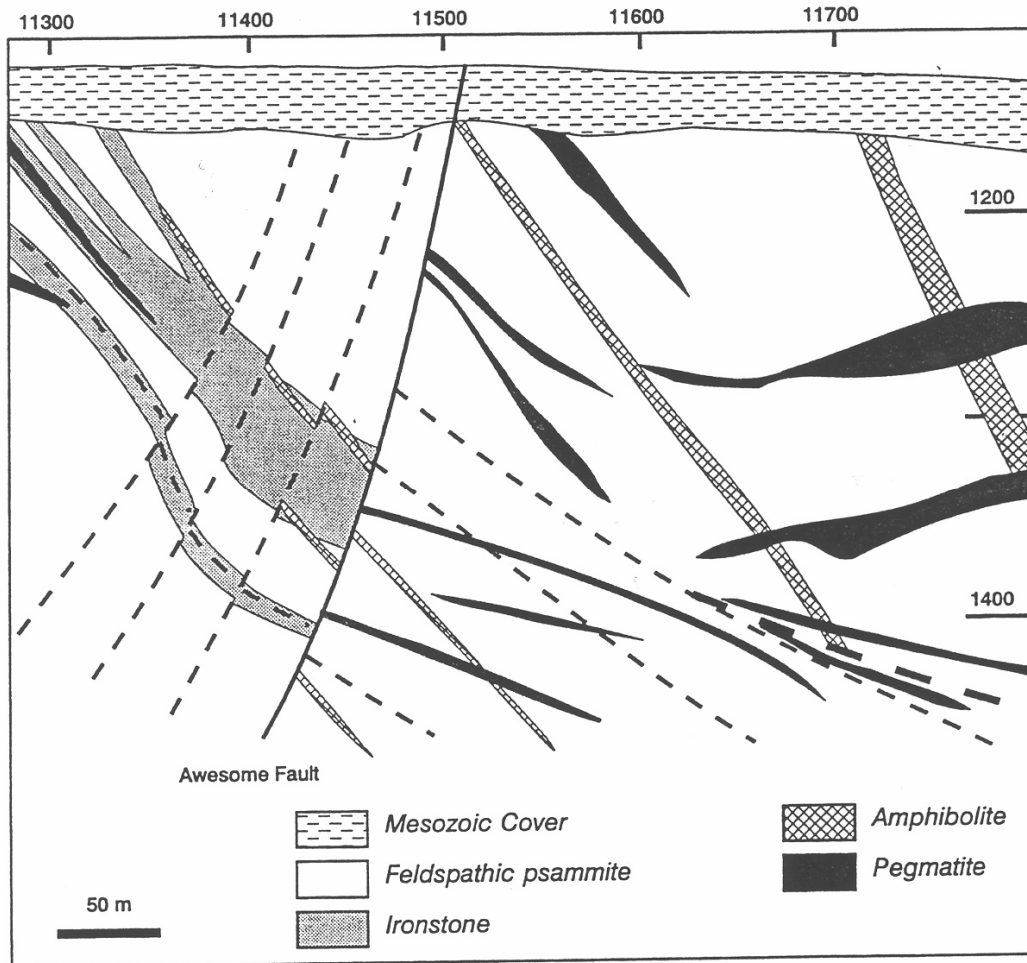
The Osborne Cu-Au deposit (Figs. 15-17) is situated 155 km south of Cloncurry and is a blind deposit concealed by 30-40 metres of Mesozoic cover. It was discovered by drill testing a regionally significant aeromagnetic anomaly first identified in 1974 (Anderson and Logan, 1992).

Osborne is hosted within upper amphibolite facies metamorphic rocks and pegmatites. The metamorphic rocks are dominated by Na-rich feldspathic psammities, with lesser amphibolites, and a discrete body of metaperidotite some 30 metres thick (Adshead, 1995). Ironstones, predominantly composed of magnetite and quartz with minor apatite are an important lithology. These commonly have a distinct compositional layering that is far less well-differentiated than that of typical BIF. These are significant as they localize much of the mineralization in the deposit, though this is typically at subeconomic grades.

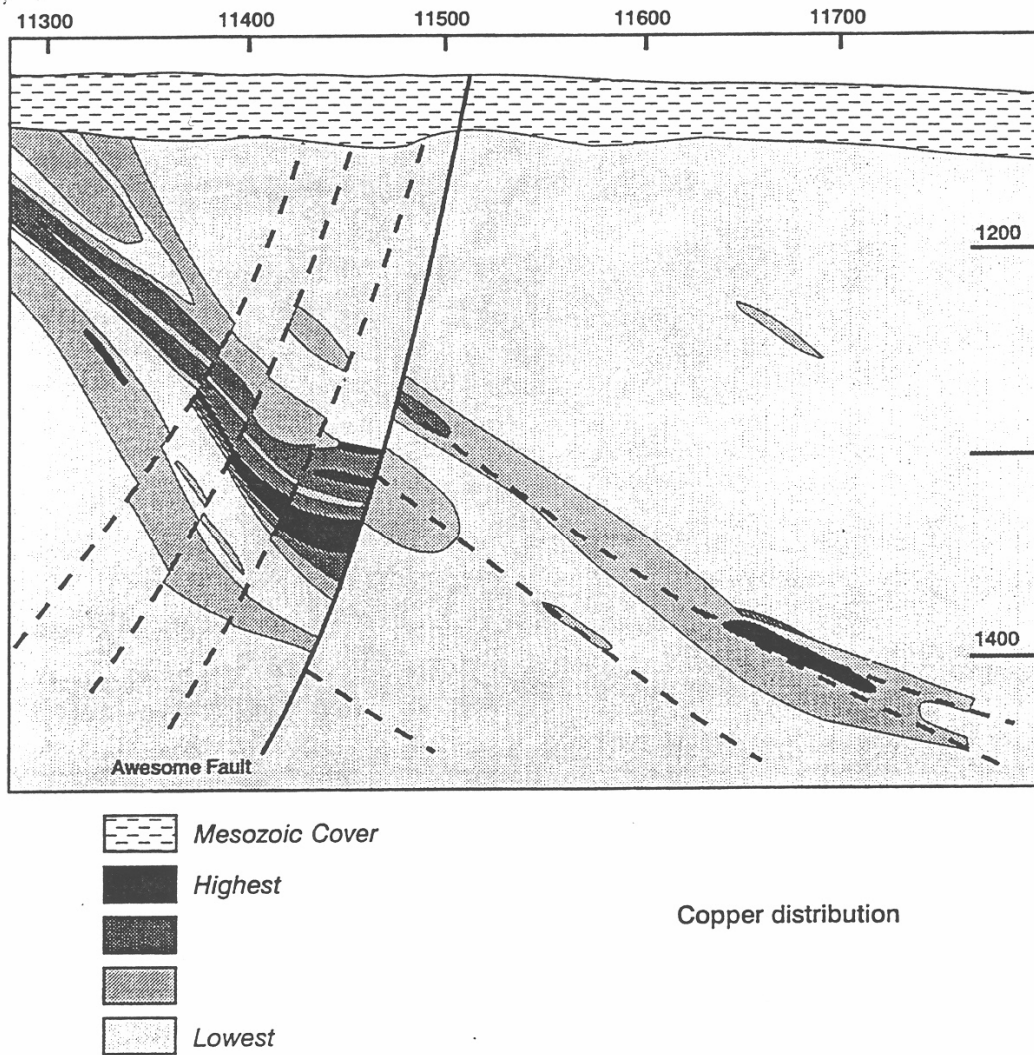
The deposit area is divided into two discrete geological domains. The western domain contains two substantial ironstone units that individually exceed 50m in thickness. These strike ESE, dip steeply to the north and exhibit complex branching (folded?) geometries. The eastern domain lacks ironstones, but has a comparatively large proportion of pegmatite intrusions. Both domains contain ore zones associated with intense postmetamorphic quartz veining selectively developed within, and close to the ironstones in the west, and interpreted to be due to dilation during faulting subparallel to major contacts (Adshead, 1995). The western domain contains pyrite-chalcopyrite with a second generation of magnetite and locally developed hematite. Eastern domain ore is concentrated in a discrete body referred to as the Eastern High Grade Lode, which consists of pyrrhotite-chalcopyrite-magnetite-pyrite-quartz. The ores are geochemically complex, being enriched in Co-Mo-Ag-Sn-W-Bi-Hg-Se-Te reflected in the presence of a variety of accessory ore minerals including cobaltian forms of pyrite and pyrrhotite, bravoite, molybdenite, wolframite, argentian native gold and a variety of bismuth sulphides, selenides and tellurides (Adshead et al., 1998). Ar-Ar dating of pre- to synmineralization hornblende and biotite gave ages close to 1540 Ma, while Re-Os dating of molybdenite indicates a mineralization age of 1595Ma (Perkins and Wyborn, 1998; Gauthier et al., 2001).



**Figure 15.** Geological plan of the Osborne Cu-Au deposit approximately 70 metres below surface (from Adshead, 1995).



**Figure 16.** Interpreted geological cross section of the Osborne Cu-Au deposit at 21360N (from Adshead, 1995).



**Figure 17.** Copper distribution in the Osborne Cu-Au deposit at 21360N (from Adshead, 1995).

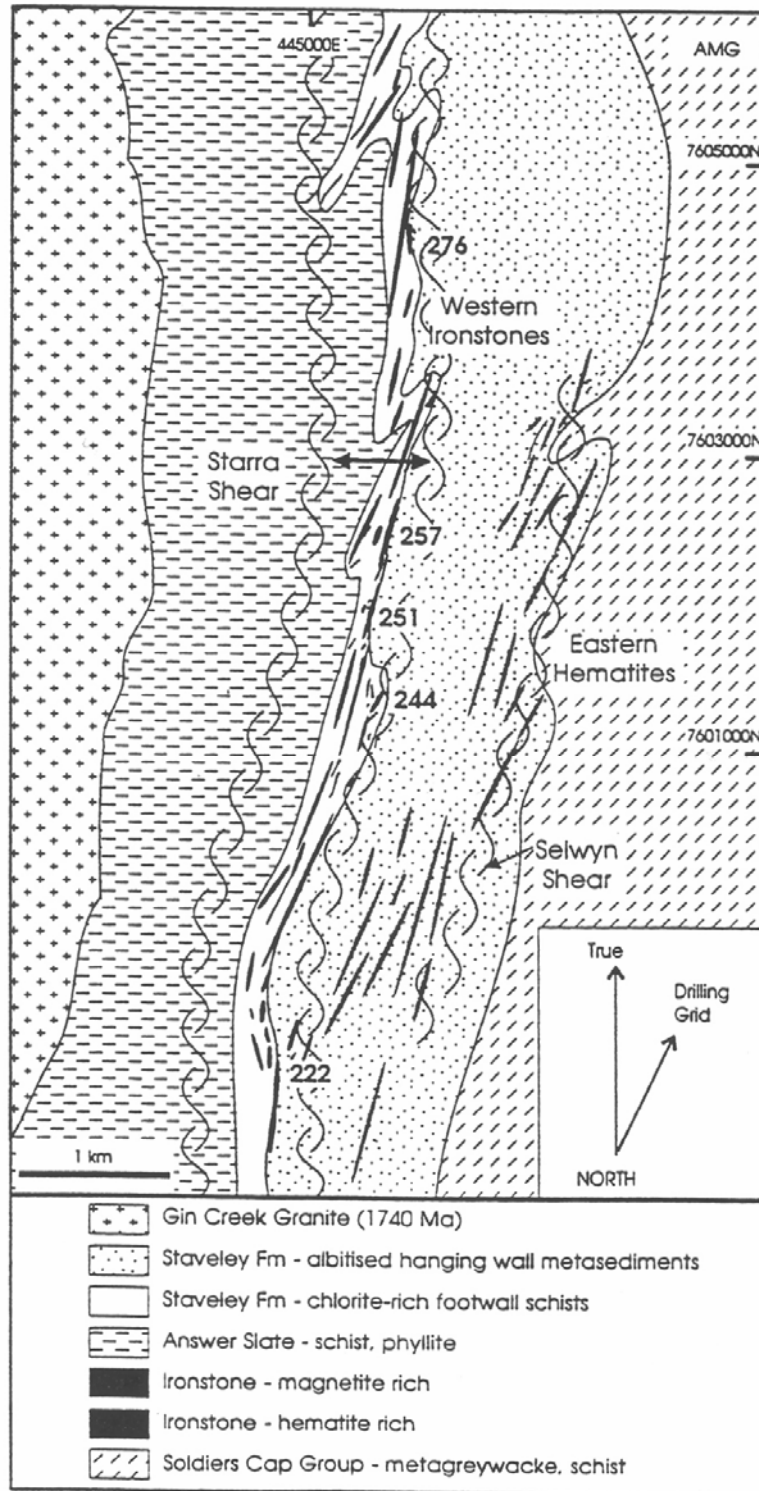
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## **Starra**

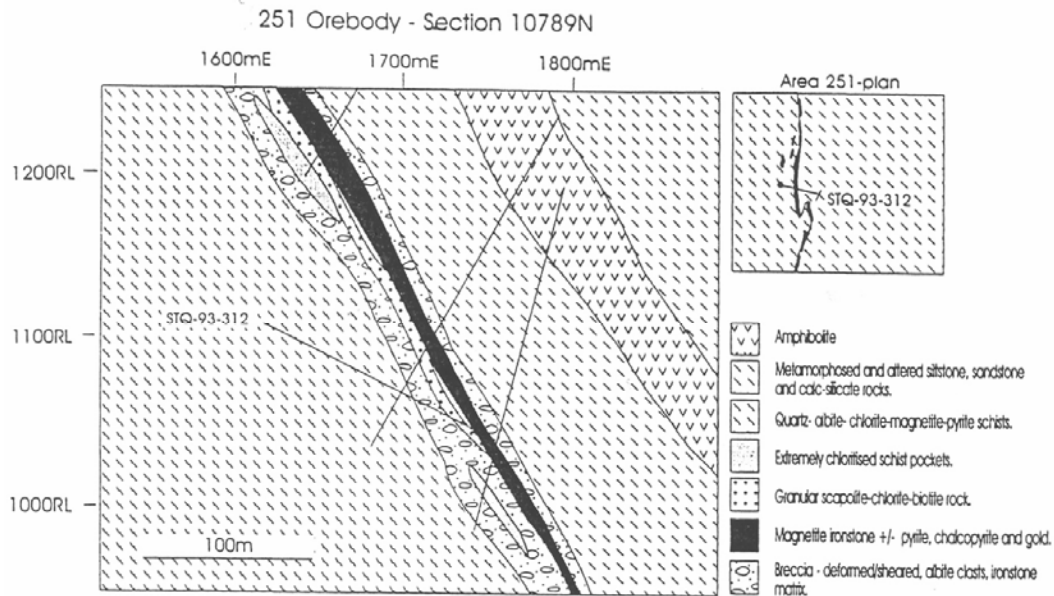
The Starra Au-Cu orebodies (Figs. 18 and 19) occur in magnetite-hematite ironstones and display some parallels with occurrences at Tennant Creek, although structural styles suggests development in a deeper-seated environment. The deposit was initially interpreted as a deformed cupriferous sedimentary iron formation (Davidson et al., 1989). However, subsequent microstructural and paragenetic studies suggested that albitization, magnetite lens formation and magnetite-biotite alteration occurred by replacement of schist and calc-silicate host rocks that occurred late in the deformation history of the region during fabric intensification in a steeply dipping shear zone (Rotherham, 1997; Adshead-Bell, 1998). This is supported by a recent Ar-Ar date of ~1503 Ma for alteration biotite (Perkins and Wyborn, 1998). Au-Cu ore is dominantly confined to ironstones and occurs in seven discrete zones over 5.5 km of strike. Pyrite, chalcopyrite, anhydrite, barite, calcite and gold, locally with hypogene bornite and chalcocite were precipitated from late oxidizing fluids in association with brecciation and hematization of the magnetite host (Rotherham, 1997).

## **Eloise**

The Eloise deposit is also localized by a steep shear zone and distinguished by a pronounced mineral zoning over a distance of some 2 km that sees the economic lode separated from the main concentrations of magnetite by several hundred metres (Brescianini et al., 1992; Baker, 1998; Baker and Laing, 1998). Lodes are hosted by amphibolite facies meta-arkoses and characterized by early albitization overprinted by hornblende-biotite-quartz alteration and infill which was selectively replaced by pyrrhotite and chalcopyrite during mineralization (Baker, 1998). Mineralization was accompanied by the development of a lower temperature alteration paragenesis containing actinolite, chlorite, muscovite and carbonate. The ore is characterized by high levels of Ni and Co plus anomalous zinc (Baker, 1998).



**Figure 18.** Geological setting of the Starra ironstones and orebodies (after Rotherham, 1997).



**Figure 19.** Cross section through the 251 orebody at Starra (from Rotherham 1997).

### Carbonaceous metasediment-hosted deposits

Greenmount, Mount Dore and Mount Elliott are some of the more significant of many loosely “stratabound” deposits in the district hosted by carbonaceous pelitic schists. These three are all associated with the regionally extensive Mount Dore Fault Zone (Laing, 1998). Greenmount occurs at a faulted contact with an oxidized calc-silicate sequence which has magnetite alteration but is unmineralized (Krcmarov and Stewart, 1998). The Cu-Au-Co resource occurs in altered and brecciated pelites with extensive K-feldspar alteration. Mount Dore is hosted by a fault-bounded slice of carbonaceous mica schists sandwiched between calcareous and calc-silicate rocks of the Staveley Formation (Mary Kathleen Group) to the west, and the Mount Dore Granite to the east (Beardsmore, 1992). Mineralization and associated alteration are localized by tabular dilational breccia bodies that were formed during post-metamorphic, pre-granite emplacement, movement on the fault system and associated hydrothermal activity (Beardsmore, 1992).

Mount Elliott is a diopside ( $\pm$ minor andraditic garnet) skarn deposit containing both magnetite-pyrite and pyrrhotite-rich chalcopyrite ore types (Wang and Williams, 1997; Fortowski and McCracken, 1998). Unusually, the Ca-Mg-Fe bearing skarn alteration overprints the non-calcareous pelites. The ore has high levels of Co and Ni. Mount Elliott is the only significant deposit in the region where intra-ore intrusions have been identified. These are trachyandesite dykes which cut the skarn alteration, but were affected by the sulphide mineralization (Wang and Williams, 1997; Drabsch, 1998).



## Lightning Creek Cu-Au prospect

The Lightning Creek Cu-Au prospect, which is located in the southern part of the Squirrel Hills Granite (Figs. 9 and 10), preserves a remarkable transition from magmatic to hydrothermal conditions, with melt crystallization accompanied by the evolution of magmatic-hydrothermal fluids strongly enriched in copper (Perring et al., 2000). The fluid inclusion and stable isotope characteristics of the magmatic-hydrothermal fluids at Lightning Creek are very similar to those from the Cu-Au deposits in the region and provide compelling evidence of a genetic link between Eureka Supersuite magmas and Cu-Au mineralization in the Cloncurry district.

The Lightning Creek prospect forms part of an approximately 20 square kilometre, 9000 nT dipolar magnetic anomaly and consists of a series of flat-lying granitic sills which are interpreted to be late-stage differentiates of Eureka Supersuite magmas (Fig. 12). The sills are associated with extensive development of magnetite-rich veins containing minor amounts of pyrite and chalcopyrite.

The texturally-complex sills (Fig. 20) occur as a series of subhorizontal intrusions which generally range from a few millimetres to a few metres in thickness, and rarely reach thicknesses of up to 12 metres. The sills display a range of mineral compositions and textures, and varying combinations of these may be present in any one sill. The majority of sills are quartzofeldspathic and aplitic, with miarolitic patches. These typically consist of albite (<75%), microcline (<30%) and quartz (25-30%) with accessory apatite, calcite, chlorite, magnetite, pyrite, titanite and zircon. Fe-rich assemblages are present as discrete intrusions or zones within predominantly quartzofeldspathic intrusions and are dominated by spherulitic textures: they are marginally younger than the aplites in which they occur. The spherulitic assemblages are commonly dominated by albite with lesser magnetite (5-20%) and quartz (10-15%) and traces of clinopyroxene and pyrite.

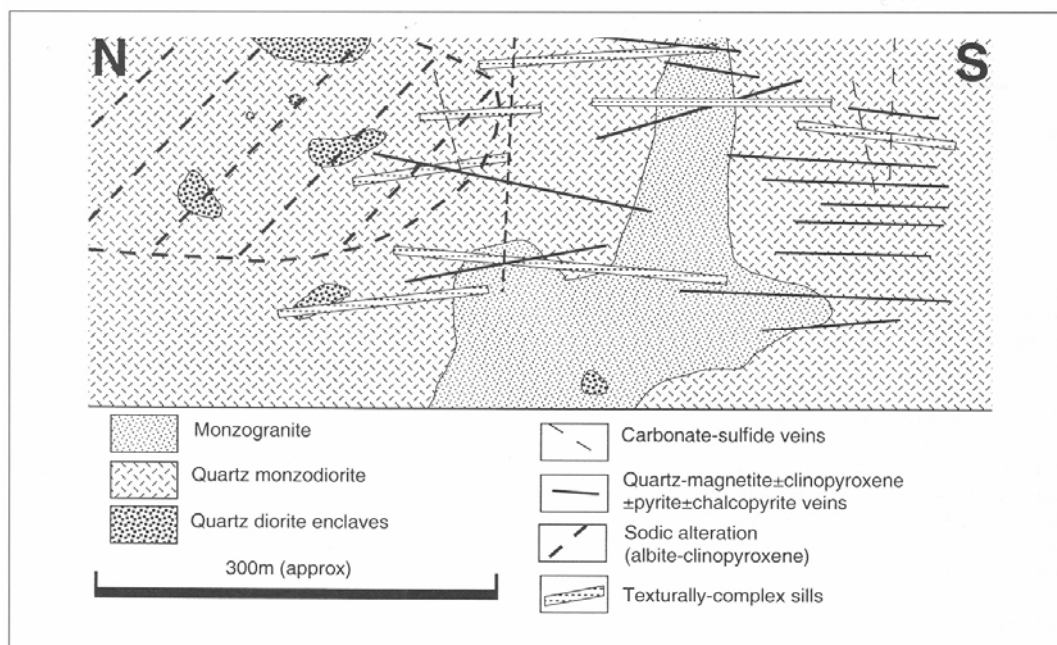
Patches of coarse-grained quartz–magnetite ± clinopyroxene ± titanite ± pyrite ± albite appear to be an integral part of some sills and are referred to as “endogenous” veins. These occur either within the body of the sill or on the margins. Zones of Fe-rich, spherulitic to balloon-textured material are commonly cored by endogenous veins. Endogenous veins which occur within equigranular sills have narrow comb-textured, coliform or spherulitic selvages. Other endogenous veins comprise crystals of coarse-grained clinopyroxene–magnetite±quartz which occur on the very margins of sills, present faceted terminations towards the interior of the sill, and are overgrown by graphic or spherulitic quartzofeldspathic material.

Four generations of brittle veins are recognized. The first two are broadly coeval with the sills and have a similar mineralogy to the endogenous veins. A minor phase of narrow magnetite veins with halos of magnetite ± clinopyroxene alteration appear marginally to predate the dominant set of flat-lying quartz–magnetite ± clinopyroxene ± pyrite ± albite ± titanite ± apatite veins. The latter do not have alteration halos. These magnetite-rich vein sets are sequentially overprinted by: (i) narrow, steeply dipping calcite ± chlorite ± pyrite ± chalcopyrite veins, and (ii) late epidote veins which are associated with intense epidote alteration of the host rocks.

The textures displayed by the Fe-rich parts of the sills (complex spherulites with multiple growth rings) suggest episodic release of a fluid phase during solidification. Much of this fluid presumably escaped along zones which subsequently crystallized as endogenous veins (in the sill from which the fluid originated) or brittle veins (in other intrusives). Some of the fluid apparently was not channeled away, either crystallizing as quartz-rich patches in the balloon-textured zones or causing calcic  $\pm$  ferruginous  $\pm$  sodic alteration in the recrystallized zones. While this alteration is also characterized by the growth of clinopyroxene and albite, it differs from the pervasive Na–Ca alteration in that quartz is destroyed and magnetite is commonly a dominant component of the alteration assemblage.

Both sills and veins contain the same high-temperature, hypersaline (Type 1) and CO<sub>2</sub>-rich (Type 3) fluid inclusions as have been found in the Cu–Au deposits in the Cloncurry district. The common co-existence of these two inclusion types and their respective liquid- and vapor-rich compositions suggest that phase separation occurred within an original H<sub>2</sub>O–CO<sub>2</sub>–salts fluid generated during crystallization of the sills.

Minor Cu–Au mineralization is associated with the late calcite  $\pm$  chlorite  $\pm$  pyrite  $\pm$  chalcopryrite veins at Lightning Creek. PIXE analysis of hypersaline fluid inclusions from one of the sills indicates copper contents of >1 wt.% in the fluid. However, the deposition of significant quantities of Cu from the fluid would have required an adequate supply of reduced S to form sulphide, and this may have been lacking at Lightning Creek. Although sulphur-isotopic data are compatible with a magmatic source (-3.2 to -5.0‰), no sulphur species were detected in any of the Lightning Creek fluids and the high Ba contents of the hypersaline fluids indicate that the abundance of oxidized S species was low (barium sulphate is extremely insoluble) at this stage.



**Figure 20.** Schematic cross section of the Lightning Creek prospect (see Williams and Pollard, 2001).

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Iron-oxide-Cu-Au-(U-REE) deposits – A practical guide  
and update of recent developments

**Part 3. Geological aspects of selected IOCG deposits in  
South America**

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## Coastal Batholith of Chile

### Regional geology

A belt of Cu and Fe mineralization typically less than 30 km wide extends for approximately 1200 km from La Serena to Tocopilla in northern Chile (Figs. 1-3). Mineralization also extends into southern and central Peru where deposits such as Raul, Condestable and Marcona (Vidal et al., 1990; Espinoza, 2002; Hawkes et al., 2002) may belong to the Fe-oxide Cu-Au grouping.

Mineralization in the Coastal batholith of Chile is closely associated with the Atacama Fault Zone and occurs within the remnants of a late Jurassic - early Cretaceous magmatic arc and associated early Cretaceous back arc sedimentary sequences (Fig. 3b). These rocks formed at a continental margin composed of older Paleozoic sediments and Permian to Triassic intrusions belonging to an older magmatic arc.

Late Jurassic to early Cretaceous basaltic andesite to andesite flows and volcanic breccias with intercalations of felsic tuffs, sandstone conglomerate and limestone characterize the eastern parts of the belt, with predominantly shallow marine carbonate sequences becoming more abundant higher in the stratigraphic sequence and to the east. The western part of the belt is dominated by the Coastal Batholith composed of Jurassic to early Cretaceous calc-alkaline plutons predominantly of intermediate composition, but ranging from gabbro to granite.

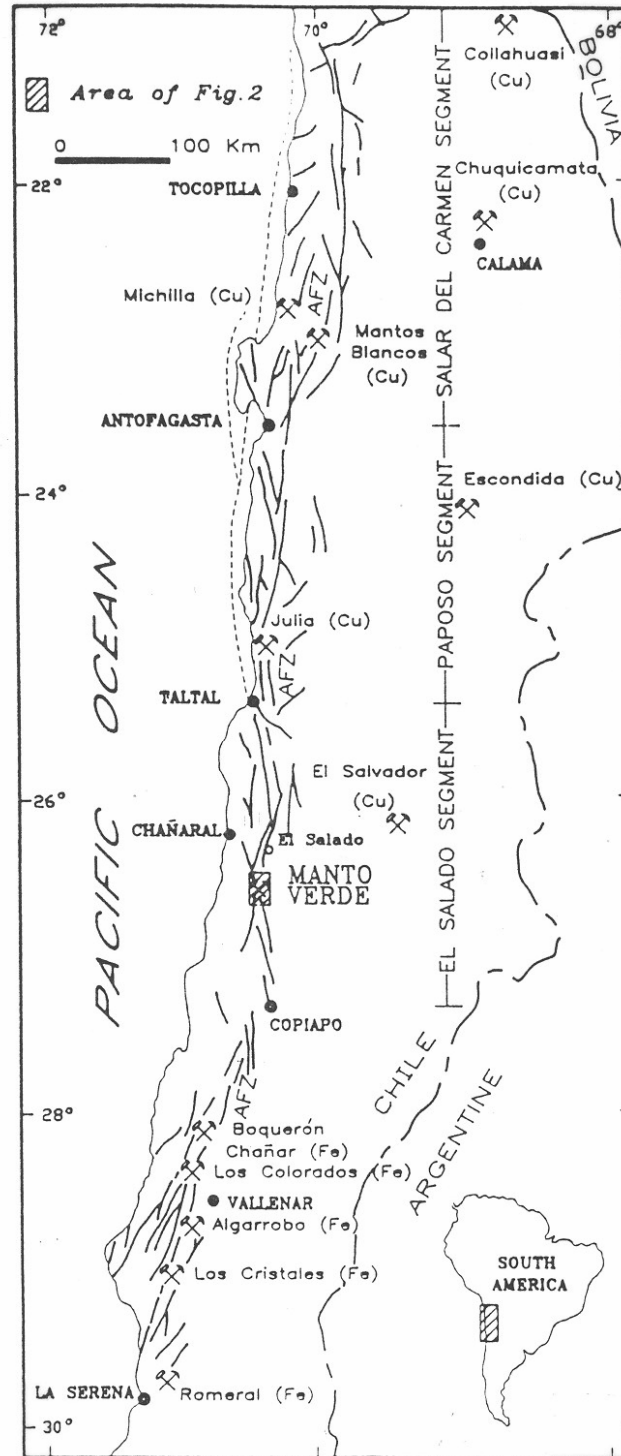
The Atacama Fault Zone is developed parallel to the strike of the Mesozoic magmatic arc and was active during arc formation (Brown et al., 1993). The fault system was initiated under mid-crustal conditions and brittle faults overprint ductile shear zones marked by mylonitic schists.

### Magnetite-apatite deposits

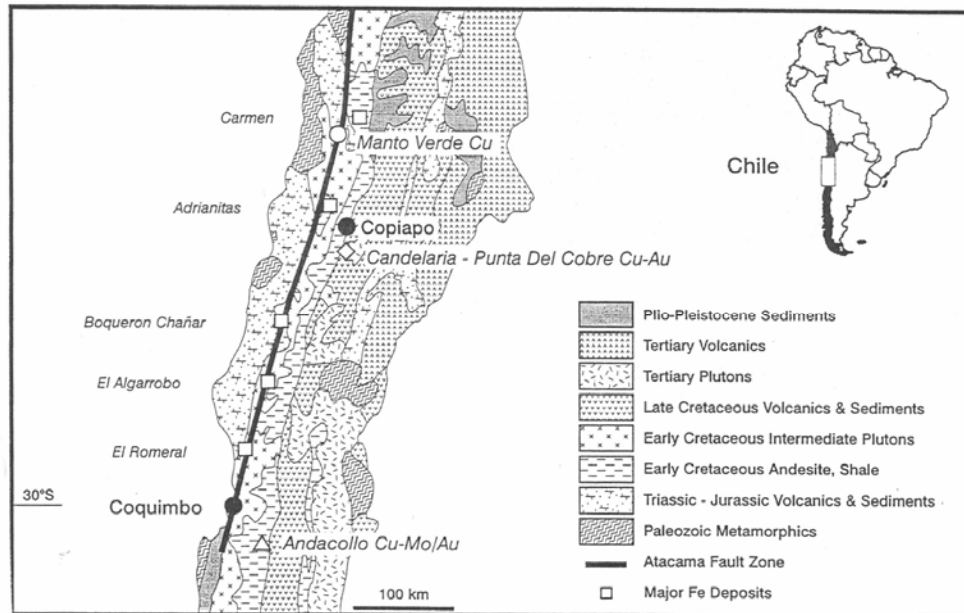
Approximately 40 iron deposits have been exploited in the Coastal Batholith, with deposits ranging up to approximately 150 Mt of high grade material (Espinoza, 1990). Most of the deposits are hosted within andesitic volcanic rocks in mylonitic domains along the Atacama Fault Zone (e.g. Fig. 1 and 2).

The iron deposits consist principally of magnetite with varying amounts of apatite and actinolite. The deposits are commonly located on major fault structures, and the ores commonly have vein and breccia textures. Alteration assemblages associated with the magnetite-apatite deposits include combinations of albite-oligoclase, scapolite, actinolite-hornblende, magnetite, biotite, K-feldspar, titanite, quartz and calcite (Brookstrom, 1977; Espinoza, 1990; Fig. 4). Pyrite and chalcopyrite are common constituents of hypogene ores.

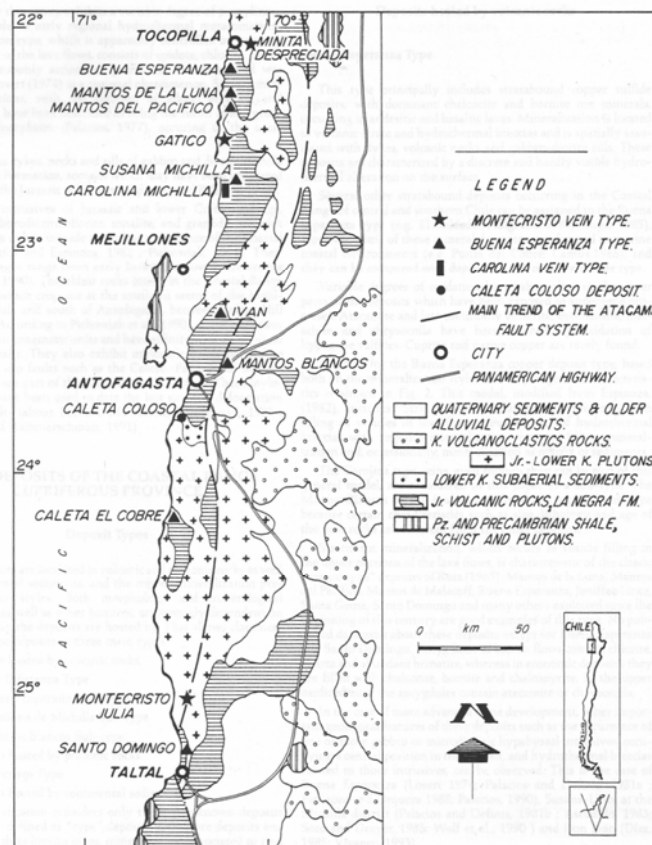
The magnetite-apatite deposits are sometimes closely associated with intrusive activity (e.g. Brookstrom, 1977; Ménard, 1995) and are believed to have formed at depths of approximately 4 km.



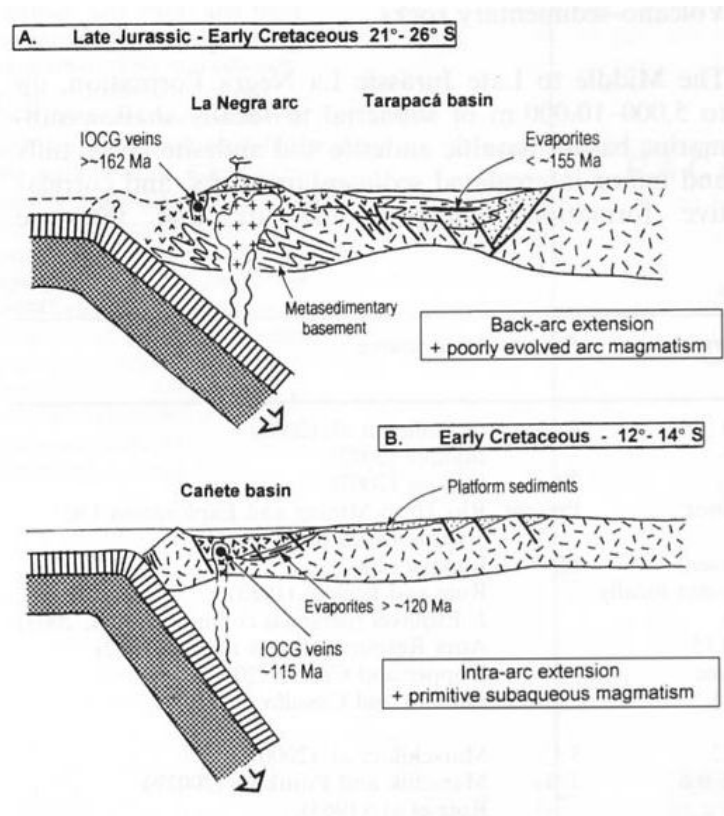
**Figure 1.** Major segments of the Atacama Fault Zone and major iron deposits of the Chile Coastal Batholith (from Vila et al., 1996).



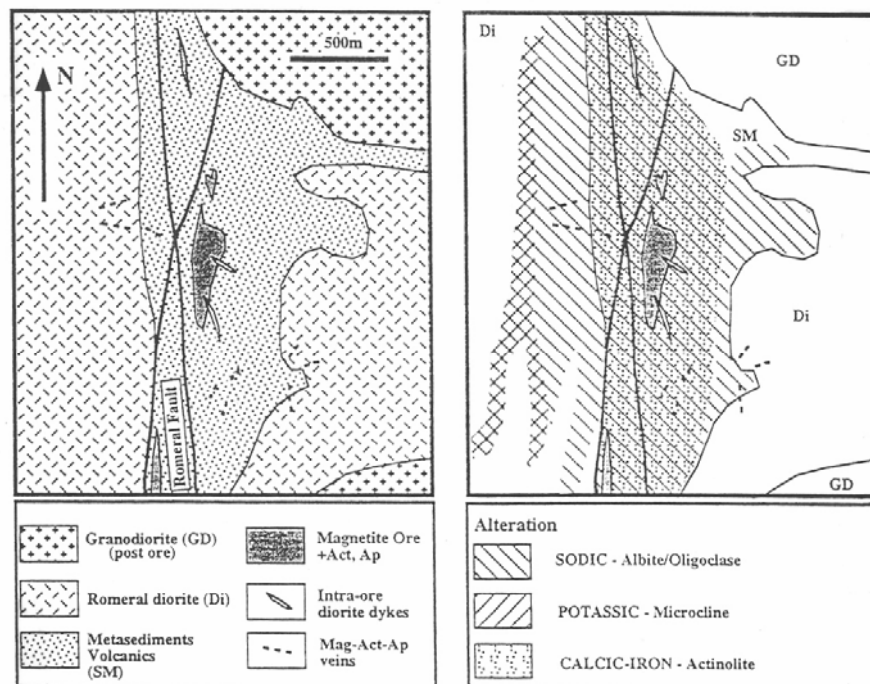
**Figure 2.** Geology and major ore deposits of the southern portion of the Chile Coastal Batholith (adapted from Espinoza, 1990).



**Figure 3a.** Simplified geology and distribution of copper deposits in the northern part of the Chile Coastal Batholith (from Espinoza et al., 1996).



**Figure 3b.** Tectonic setting of IOCG formation in the late Jurassic and early Cretaceous of northern Chile and Peru (from Sillitoe, 2003)



**Figure 4.** Fault-related alteration and magnetite deposits of the El Romeral Fe deposit, northern Chile (modified from Brookstrom, 1977).

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### **Hematite-sericite/K-feldspar Cu(-Ag-Au) deposits**

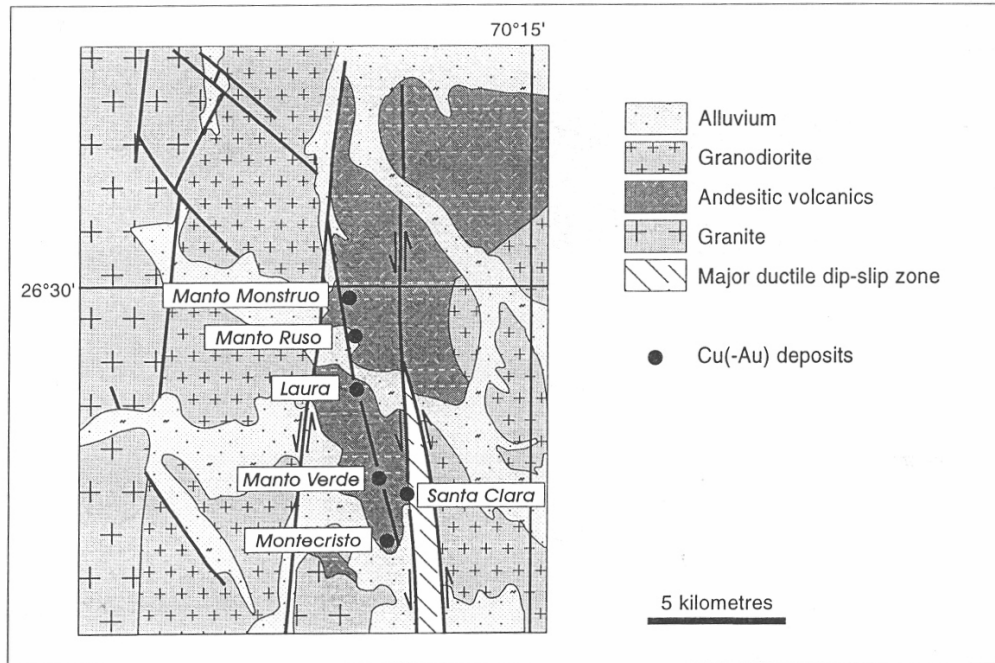
The coastal belt of northern Chile also contains numerous examples of specularite-rich Cu(-Ag-Au) deposits hosted by volcanic and volcano-sedimentary sequences. These deposits occur along the Atacama Fault Zone and include deposits such as Mantos Blancos, Buena Esperanza, El Salado, Teresa de Colmo (Barreal Seco) and Manto Verde. Mineralization is hosted by steep structures and occurs as veins, stockworks and hydrothermal breccia systems. An important component of mineralization in some deposits occurs in permeable zones in the host sequence, typically in vesicular flow tops, volcanoclastic breccias and sandy horizons. Two main types of deposits can be recognized, one with chalcopyrite and pyrite as the principal hypogene sulphides (e.g. Manto Verde, Teresa de Colmo), and one in which hypogene bornite, chalcocite and covellite are important components (e.g. El Soldado, Mantos Blancos, Michilla). The chalcopyrite-rich deposits are generally accepted as belonging to the Fe-oxide Cu-Au class of deposits, while the inclusion of the bornite-chalcocite deposits is more problematical (see below).

#### ***Chalcopyrite type: Manto Verde (see Vila et al., 1996)***

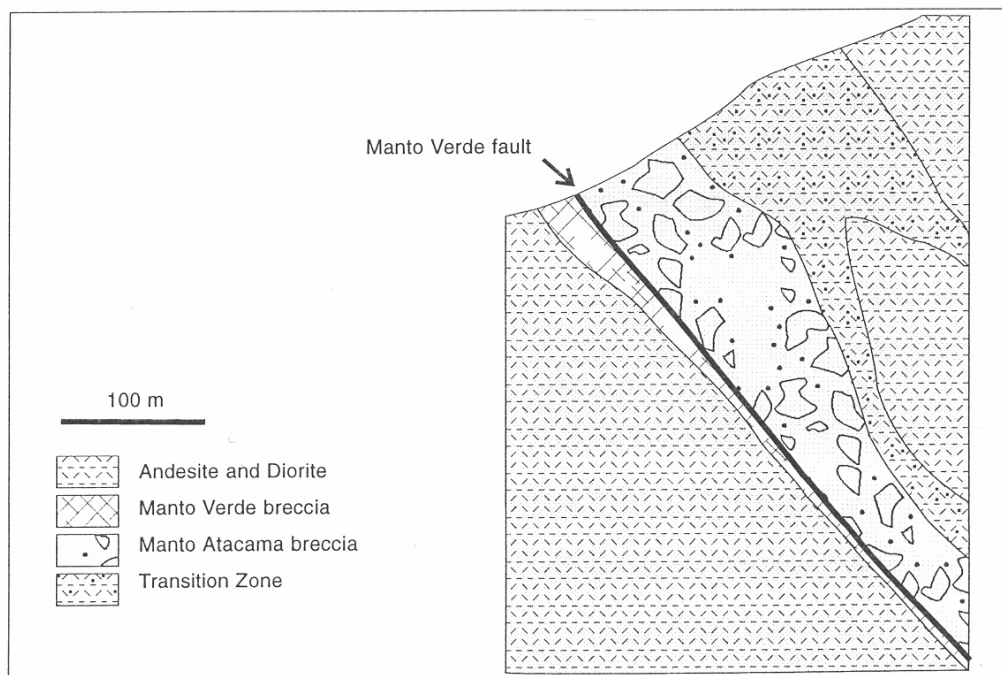
The Manto Verde deposit has a copper oxide resource on the order of 120 million tonnes at 0.72% Cu (Vila et al., 1996; Figs. 5 and 6). Mineralization is hosted by andesitic volcanics and diorite porphyry stocks of Early Cretaceous age which have been subjected to district scale K-metasomatism characterized by microcline and chlorite. Mineralization occurs within three breccia units at least 1500 metres long which parallel the Manto Verde fault, a 12 km long east dipping structure that joins two major branches of the Atacama Fault Zone (e.g. Fig. 5). Approaching the Manto Verde fault, K-feldspar alteration is overprinted by sericitic alteration. The most extensive breccia is the Manto Atacama breccia, a 10 to 100 metre wide, matrix-supported and poorly sorted hydrothermal intrusive breccia which grades outwards into a transition zone of stockwork fractures (Vila et al., 1996; Fig. 6). The breccia contains andesite and/or diorite clasts (1-30 cm) in a matrix of deformed specularite and rock flour, with patches of calcite. Hypogene sulphides which accompany specularite consist of chalcopyrite and pyrite, and gold grades of 0.3 to 0.5 g/t occur in the breccias (Vila et al., 1996).

Fluid inclusions from pre-ore quartz and ore stage calcite generally homogenize in the range 180-250°C, with salinity decreasing from 30-40wt.% NaCl equivalent in early quartz to 14-21wt.% NaCl equivalent in ore-stage calcite (see Vila et al., 1996). The presence of coexisting vapor-rich inclusions in early quartz is consistent with boiling, and the subsequent decrease in salinity may be due to influx of a low salinity fluid.

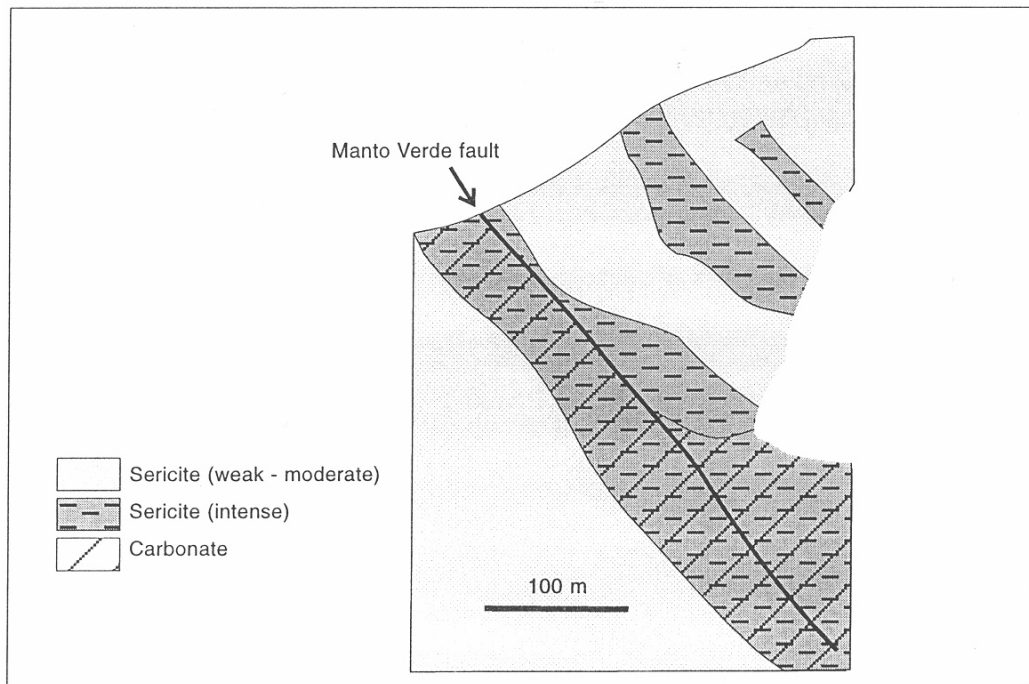
Vila et al. (1996) suggest that Manto Verde formed from magmatic-hydrothermal fluids released from an underlying granitic pluton which is expressed as a dyke system at surface which parallels the Manto Verde fault.



**Figure 5.** Geological setting of the Manto Verde Cu deposit, northern Chile. Modified from Vila et al. (1996).



**Figure 6a.** Generalized cross-section showing the distribution of geological units in the Manto Verde Cu deposit (from Vila et al., 1996).



**Figure 6b.** Generalized cross-section showing the distribution of hydrothermal alteration in the Manto Verde Cu deposit (from Vila et al., 1996).



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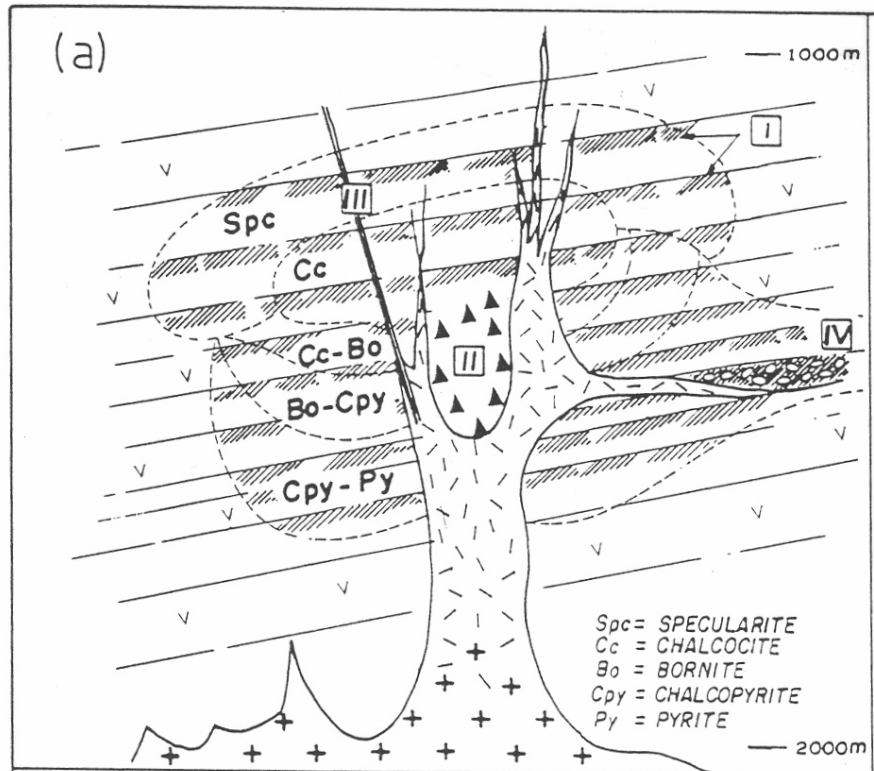
***Bornite-chalcocite type: e.g. Buena Esperanza and Mantos Blancos***

The bornite-chalcocite deposits are hosted within volcanic and volcano-sedimentary rocks, with mineralization occurring in vesicular flow tops, volcanoclastic breccias and sandy horizons, as well as in hydrothermal breccias and veins (Chavez, 1985; Espinoza et al., 1996, Figs. 7-9). Gabbro and/or microdiorite intrusions occupy central positions in some deposits and are related to the hydrothermal breccias (Espinoza et al., 1996, Fig. 7). At the Susana deposit (Fig. 8), mantos ranging from 1 to 25 metres thick extend approximately 100 metres outwards from a central breccia body, and are developed preferentially in the vesicular tops of lava flows (Espinoza et al., 1996). At El Soldado, the orebodies occur principally within an albitized basalt-rhyodacite sequence and show strong control by subvertical faults.

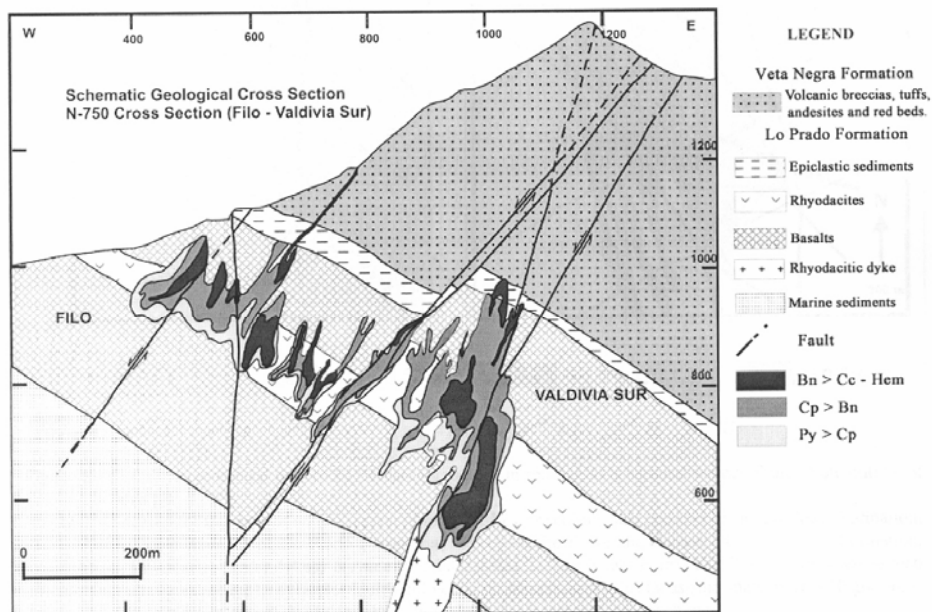
Deposits are typically found in relatively reduced horizons juxtaposed with red, commonly hematized units (Sillitoe, 1992). Primary ore zoning from top to bottom and from the centre outwards includes: specularite-calcite, chalcocite-digenite-covellite, bornite-digenite-covellite-chalcocite-galena, bornite-chalcopyrite and chalcopyrite-pyrite (Espinoza et al., 1996). This zoning is considered to result from an early stage consisting of hematite, pyrite and chalcopyrite, and a later stage chalcopyrite, bornite and covellite (Espinoza et al., 1996). The silver content of the copper sulphides increases progressively through the hypogene paragenesis (Chavez, 1985).

Hydrothermal alteration is sometimes difficult to distinguish from the low grade (prehnite-pumpellyite to greenschist) metamorphic assemblages of the host rocks and consists mainly of chlorite, hematite, sericite, calcite and titanite, with albitization in mineralized zones (Sillitoe, 1992; Espinoza et al., 1996). The general paragenetic sequence at Mantos Blancos consists of early specularite, followed by quartz, albite, chlorite, sulphides and sericite (Chavez, 1985, Fig. 10).

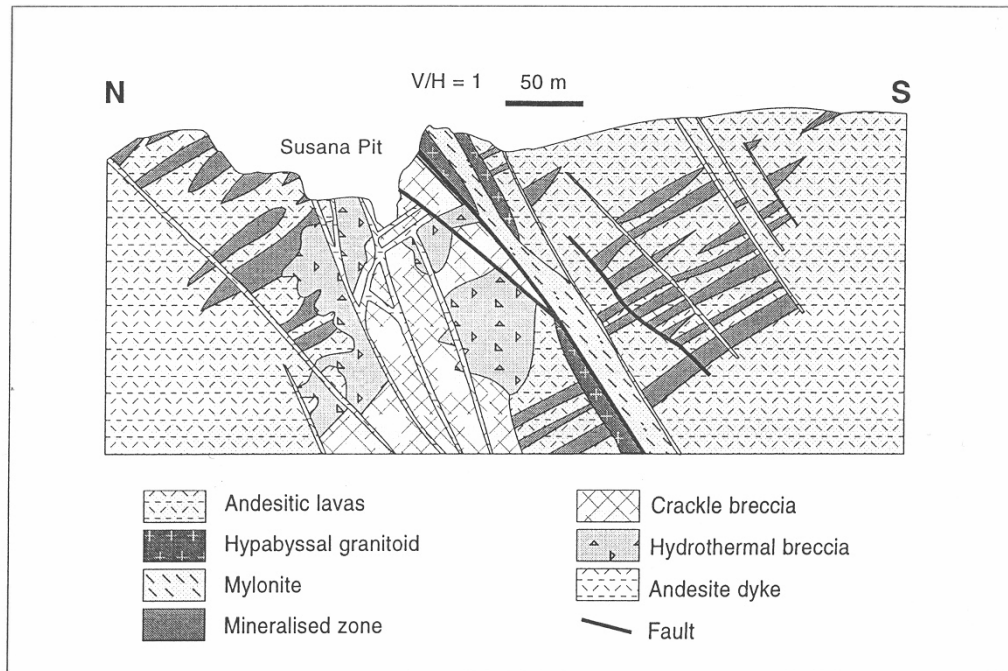
The relationship between the bornite-chalcocite deposits and the Fe-oxide Cu-Au deposits in northern Chile remains a matter of some debate (Maksaev and Zentilli, 2002; Sillitoe, 2003). The deposits occur distal to contemporaneous Upper Jurassic or Lower Cretaceous granodiorite batholiths (e.g. Maksaev and Zentilli, 2002; Boric et al., 2002), and stable and radiogenic isotope evidence has been used to suggest that the fluids responsible for mineralization are mixtures of basin-derived and meteoric fluids (Maksaev and Zentilli, 2002; Kojima et al., 2003; Wilson et al., 2003). Fluid inclusion data indicate that the fluids were predominantly H<sub>2</sub>O-NaCl-CaCl<sub>2</sub> fluids of variable salinity (3-34 wt.% equivalent NaCl) with ore formation occurring at temperatures ranging from approximately 200 to >300°C (Boric et al., 2002; Kojima et al., 2003). The zoning pattern of oxide and sulphide minerals in the bornite-chalcocite deposits is the same as that observed in the Olympic Dam deposit in Australia.



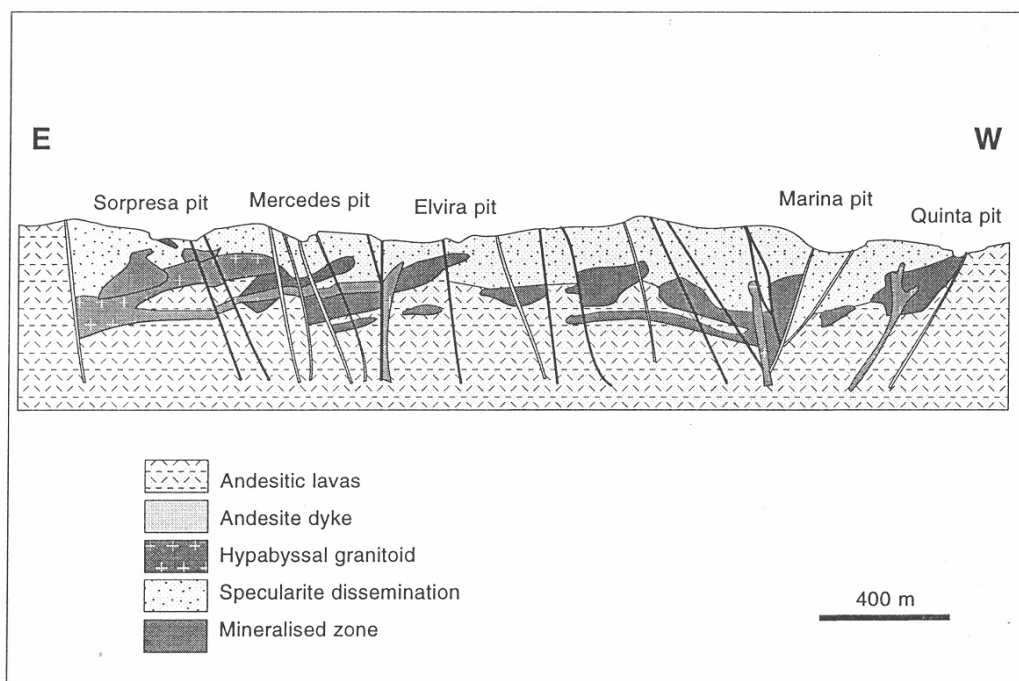
**Figure 7a.** Schematic cross-section of the Buena Esperanza deposit, northern Chile. I - mineralization in the vesicular portion of lava flows, II - mineralization in hydrothermal breccias, III - mineralization in fractures and faults, and IV - mineralization in porous sedimentary rocks. From Espinoza et al. (1996).



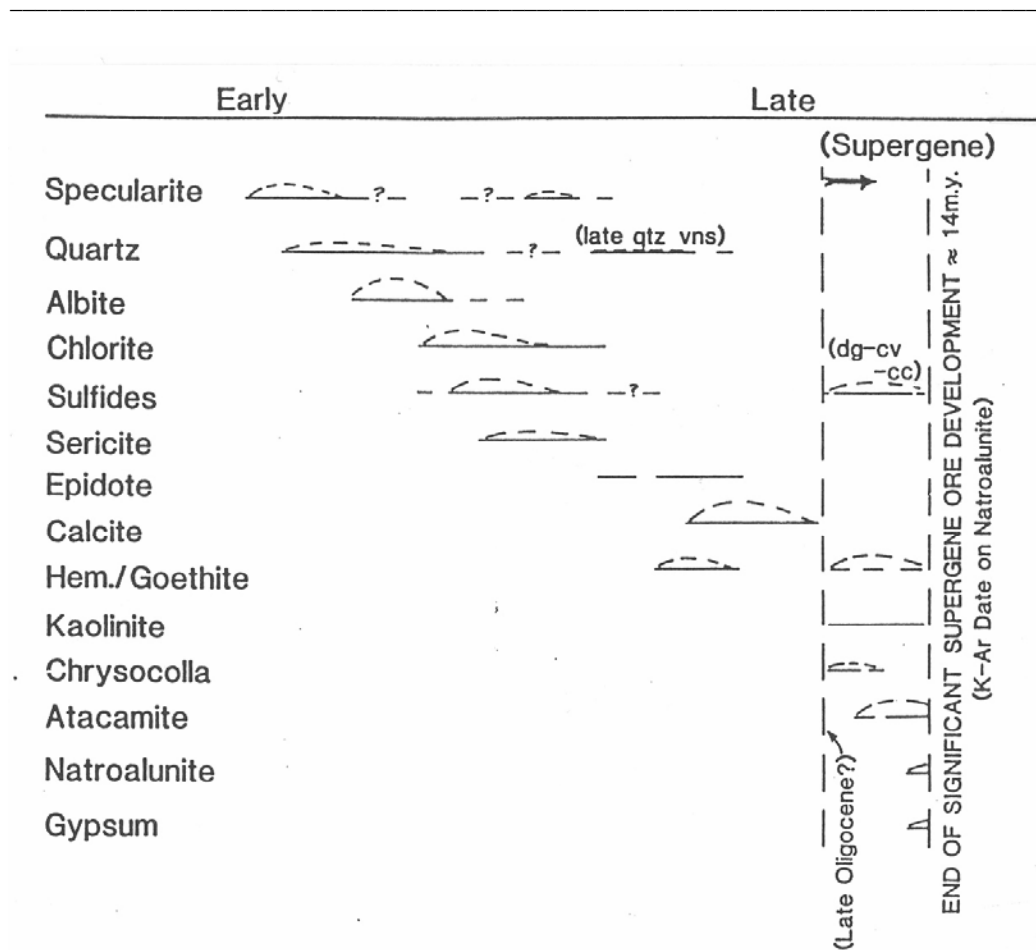
**Figure 7b.** Cross section at N-750 in the El Soldado Cu deposit, northern Chile (from Boric et al., 2002).



**Figure 8.** Schematic cross section of the Susana Cu deposit, northern Chile. Modified from Espinoza et al. (1996).



**Figure 9.** Schematic cross section of the Mantos Blancos Cu deposit, northern Chile. Modified from Espinoza et al. (1996).



**Figure 10.** Mineral paragenesis of the Mantos Blancos copper deposit (from Chavez, 1985).

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## **The Candelaria copper-gold deposit, Chile**

The Candelaria copper-gold deposit (Fig. 11) is located in the Copiapo district near the town of Tierra Amarilla in the Atacama Desert region of Chile. The deposit was discovered by Phelps Dodge in 1987 when conducting an IP survey over an area of known mineralization. Mineable reserves are estimated at 366 million tonnes averaging 1.08% Cu and 0.26 g/t Au (Ryan et al., 1995).

The oldest rocks in the area of the Candelaria deposit are metavolcanics of the Punta del Cobre Formation of Early Cretaceous age. Approximately 1 kilometre west of the deposit is a mid-Cretaceous batholith composed of pyroxene diorite and albite granite.

### *Local Geology*

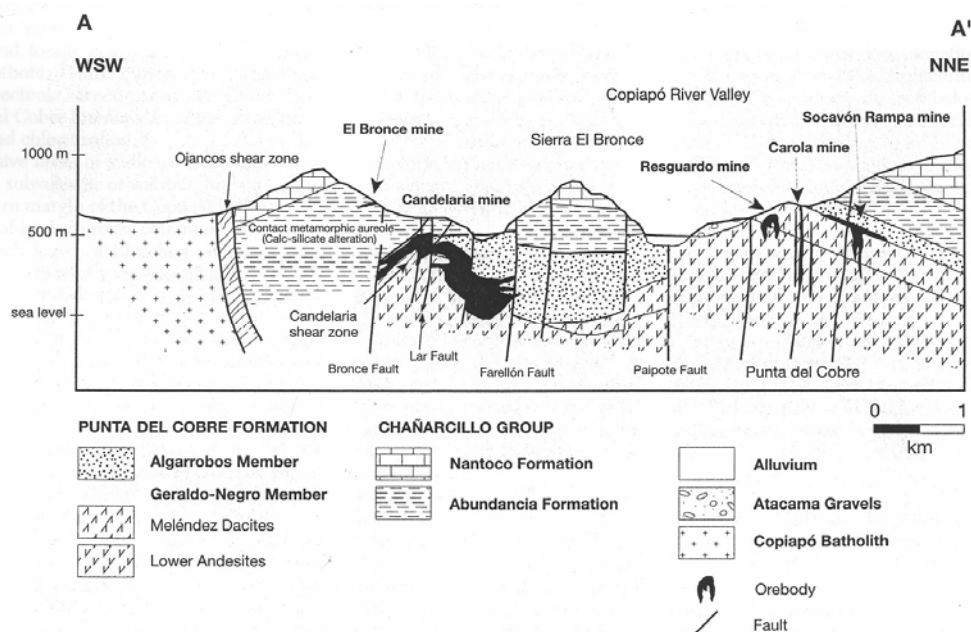
The host rocks at Candelaria are divided into four major groups: lower andesite, tuff, upper andesite and metasediments (Fig. 12). Most of the mineralization is hosted within the lower andesites which are in excess of 600 metres thick, and consist dominantly of andesite flows, crystal tuffs and volcanic breccias. The lower andesites are commonly intensely altered, with potassic alteration dominated by biotite with actinolite, apatite and titanite being prevalent in the mine area.

Tuffs or volcanoclastic sediments which overly the lower andesites range up to 200 metres thick. The original character of the rocks is difficult to determine due to alteration. Mineralogically, they consist largely of biotite, andraditic garnet, actinolite, hornblende, pyroxene, quartz, orthoclase, plagioclase, scapolite, titanite and apatite. There appears to be a crude zoning from hornblende-rich rocks lower in the unit, grading upwards into pyroxene-rich rocks and garnet-scapolite-diopside rocks at the top (Ryan et al., 1995).

The upper andesites which overly the tuff unit are composed dominantly of andesite flows with interbedded breccia and tuff. Near the contact with the overlying sediments, the upper andesites are commonly converted to skarns composed mainly of andraditic garnet, actinolite, hornblende, pyroxene, sodic plagioclase and scapolite (Ryan et al., 1995).

The upper andesites are unconformably overlain by skarn-altered and hornfelsed metasediments. The skarns are composed predominantly of andraditic garnet, pyroxene, actinolite and scapolite. Thin, interbedded garnetites are composed of greater than 90% andraditic garnet which is commonly zoned to more aluminous rims. Hornfelses are composed of plagioclase, biotite, pyroxene, actinolite and scapolite. The garnet-pyroxene-scapolite skarns host small lenses of magnetite, pyrite and chalcopyrite which were mined in the Lar and Bronce mines.

The host rocks to Candelaria have been cut by dykes, sills and possibly small stocks of dacite, tonalite and granodiorite that have been partially altered to actinolite, diopside, albite and scapolite. In places they are cut by garnet veins or partially altered to scapolite skarn.



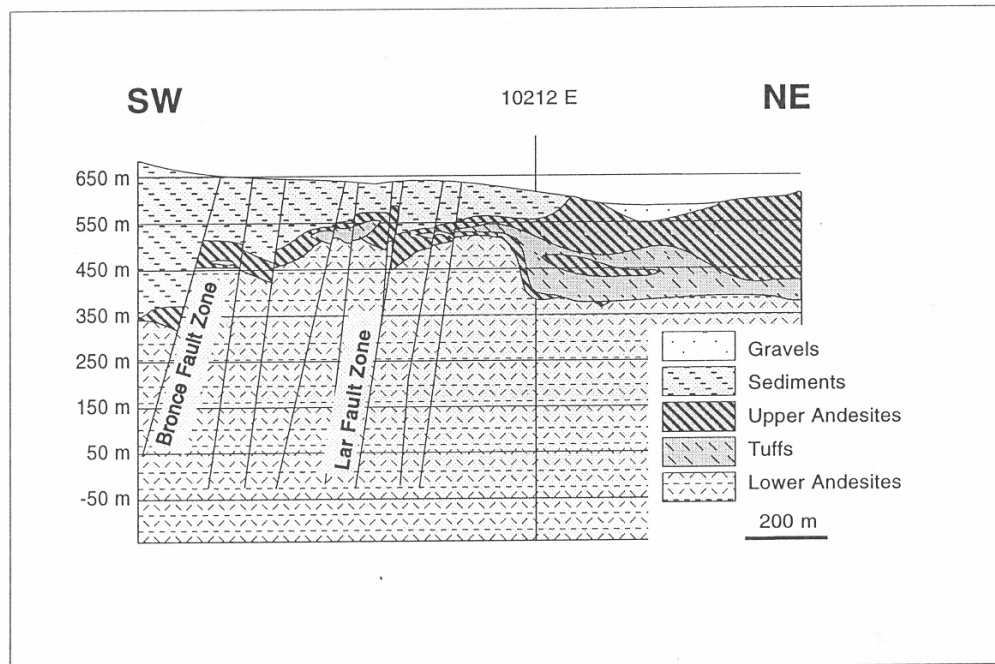
**Figure 11.** Schematic cross section through the Candelaria-Punta del Cobre area (from Marschik and Fontboté, 2001).

### *Mineralization*

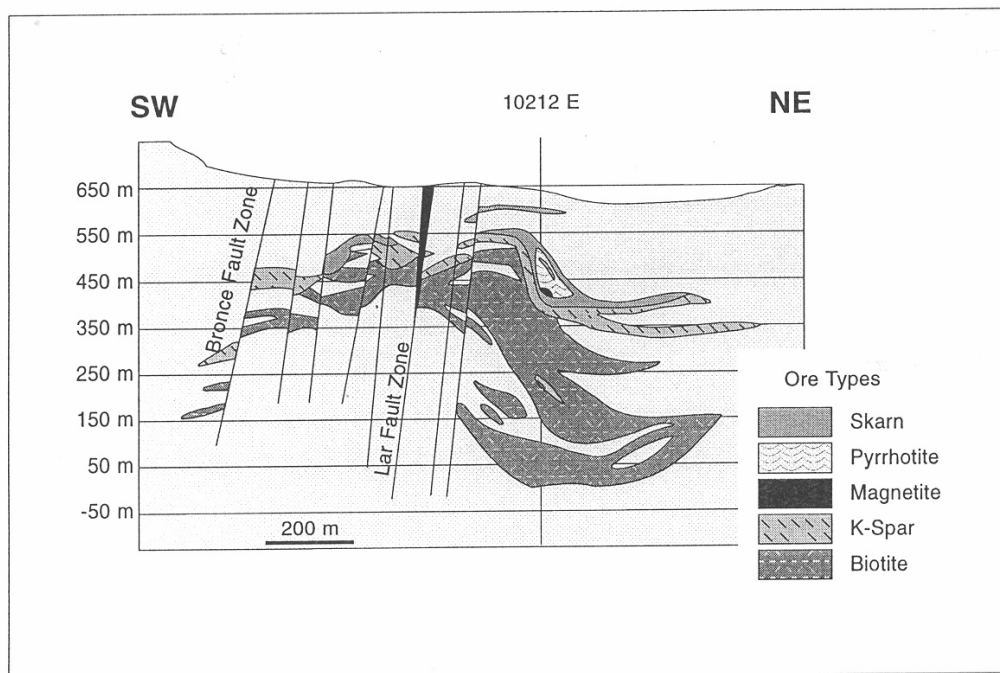
The main part of the Candelaria deposit is approximately 2000 metres long by 600 metres wide, and elongated north-south. The most prominent structural feature is the north-south trending Tierra Amarilla Anticlinorium, which is cut by a series of high angle, north-south trending faults, many of which offset ore zones (Fig. 13). In the tuff unit, mineralization has a vertical extent of at least 350 metres, and the eastern flank of the fold structure hosts the highest grade mineralization. Five main ore types are recognized based on their metallurgical characteristics: skarn ore (10%), pyrrhotite ore (5%), potassium feldspar ore (10%), magnetite ore (3%) and biotite ore (72%).

Skarn ore consists of pyroxene, andraditic garnet, calcite, quartz, actinolite, hornblende and scapolite, with sulphides in fractures and disseminated in the wall rocks. Skarn ore occurs predominantly near the top of the orebody, within the metasediments and tuff unit.

Pyrrhotite ore is similar to skarn ore, but contains visible pyrrhotite. This ore type occurs predominantly in the metasediments and tuff unit, but is also present in the lower andesites. A common assemblage in pyrrhotite ore is pyrrhotite-chalcopyrite-pyrite-magnetite-hedenbergite, indicating lower oxygen fugacity compared to typical skarn ore.



**Figure 12.** Geologic cross-section through the Candelaria deposit. Note the steep-sided asymmetrical fold located near the centre of the section (from Ryan et al. 1995).



**Figure 13.** Cross-section showing the distribution of ore types in the Candelaria deposit. From Ryan et al. (1995).

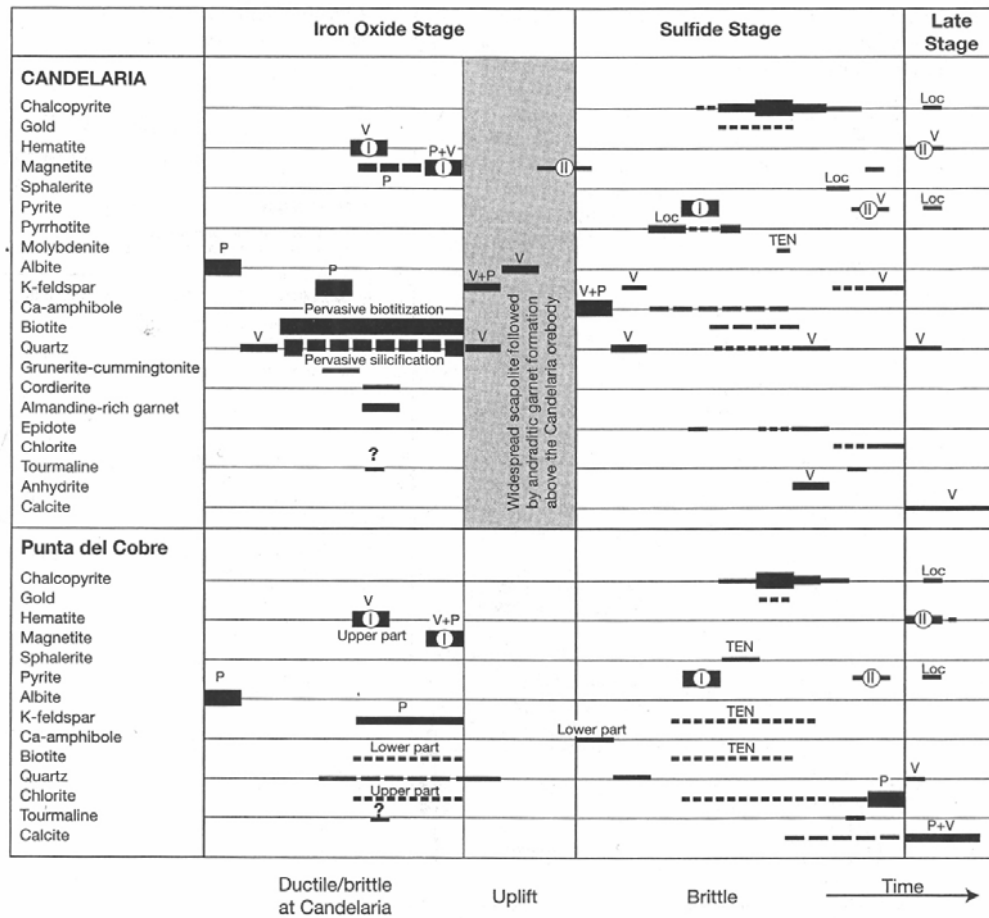
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Potassium feldspar ore contains K-feldspar flooding as well as veins and vein selvages, and occurs mainly along the contact between the tuff unit and the lower andesites. Magnetite ore is present as small lenses and breccia filling throughout the deposit, with significant amounts occurring in areas below current economic cut-off grade. Biotite ore is characterized by biotite alteration and occurs almost exclusively in the lower andesites.

The paragenesis of alteration and mineralization has been presented by Ullrich and Clark (1999) and Marschik and Fontboté (2001; Fig 14). Ullrich and Clark (1999) recognize an early magnetite-quartz-biotite stage, followed by a main ore stage that includes chalcopyrite and Ca-Fe amphibole, and several later stages that include carbonate, epidote, hematite and sulphides. These occur mainly in overprinting shear-, crackle- and stockwork-style fracture systems localized within the core of the anticlinal structure.

Sulphur isotope data indicate the presence of early fluids with near magmatic sulphur compositions (Marschik et al., 2001), and later, more oxidized fluids that were probably evaporite-sourced brines (Ulrich and Clark, 1999). Main-stage mineralization may have been dominated by mixing of S-bearing fluids from magmatic and evaporitic sources (Ulrich and Clark, 1999). Pb isotope data are consistent with the derivation of ore lead directly from a magmatic fluid exsolved during crystallization of the adjacent batholith and/or from hydrothermal leaching of the volcanic rocks originally having similar isotopic compositions to the batholith (Marschik et al., 2003). Re-Os data also suggest that the granitoid batholith was the source of metals for the deposit (Mathur et al., 2002).





**Figure 14.** Paragenetic sequence of the Candelaria and Punta del Cobre Cu-Au deposits (from Marschik and Fontboté, 2001). Abbreviations: v – vein, p – pervasive, TEN – tentatively.

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## Carajás district, Brazil

The Carajás mineral province consists of Archean rocks of the Xingu granite-gneiss complex and a younger volcano-plutonic and sedimentary sequence (Fig. 15). These rocks are intruded by Archean and middle Proterozoic granitoids, the latter of which appear to be associated with Cu-Au mineralization (Dardenne and Schobbenhaus, 2000). The district hosts a number of Fe-oxide Cu-Au deposits, including Salobo, Igarapé Bahia/ Alemão, Sossego and Cristalino (Huhn et al., 1999; Huhn et al., 2000; Villas and Santos, 2001; Fig. 15) which are Archean in age (e.g. Marschik et al., 2001; Requia et al., 2003). Granite-hosted and granite-related Cu-Au-(Sn-W) deposits in the Carajás district, including Breves and Gameleira, are linked to Proterozoic granites intruded at approximately 1.8 Ga (e.g. Pimentel et al., 2003; Tallarico et al., 2004).

The Carajás district also hosts the Serra Pelada Au-PGE deposit which has been proposed as part of the spectrum of Fe-oxide Cu-Au deposits (Grainger et al., 2002). Given the enrichments in Bi and W in addition to Cu, Au, Co, LREE and U, and the inferred association with alkaline magmatism (Grainger et al., 2002), a link to the ca.1.8 Ga granite-related Cu-Au deposits in the Carajás district appears equally likely. Until the age of mineralization at Serra Pelada is adequately constrained, it cannot be confidently linked to the Fe-oxide Cu-Au deposits in the region.

### Salobo

The Salobo deposit (Fig. 16) occurs within a shear zone and crops out for 3.5 km along a WNW-trending ridge situated 30 km north of Serra dos Carajás. Reserves are estimated at 789 Mt @ 0.96% Cu and 0.52 g/t Au (Souza and Vieira, 2000). Mineralization is hosted within metagreywacke, amphibolite, and rocks described by Lindenmayer and Teixeira (1999) as silicate iron formation and aluminous iron formation. Two types of granites are present: 1) the deformed Older Salobo granite (2573±2 Ma) and 2) a quartz-syenite dyke called the Young Salobo granite (1880±80 Ma) that contains hydrothermal alteration including fluorite, tourmaline, uraninite, allanite, chlorite, epidote, calcite and chalcopyrite (Lindenmayer and Teixeira, 1999).

The iron formations at Salobo are composed largely of deformed biotite, amphibole and garnet that may represent hydrothermal alteration (Fig. 17). This is overprinted by magnetite veins and breccias associated with extensive magnetite alteration. Later sulphide mineralization consists mainly of chalcopyrite, bornite, chalcocite and gold, while molybdenite, uraninite, Co-pentlandite and ilmenite are also present (Lindenmayer and Teixeira (1999).

Re-Os and Pb-Pb dating indicates that ore formation at Salobo was contemporaneous with the Old Salobo granite at approximately 2.57Ga (Requia et al., 2003).

### Igarapé Bahia/ Alemão

The Igarapé Bahia Cu-Au deposit includes several gold deposits (Fig. 18) that occur in an oxidized zone developed above, and along strike of the Alemão orebody (Tazava et al., 2000). Alemão was discovered in 1996 by drilling an aeromagnetic anomaly, and reserves are estimated at 170 Mt at 1.5% Cu and 0.82 g/t Au (Barreira et

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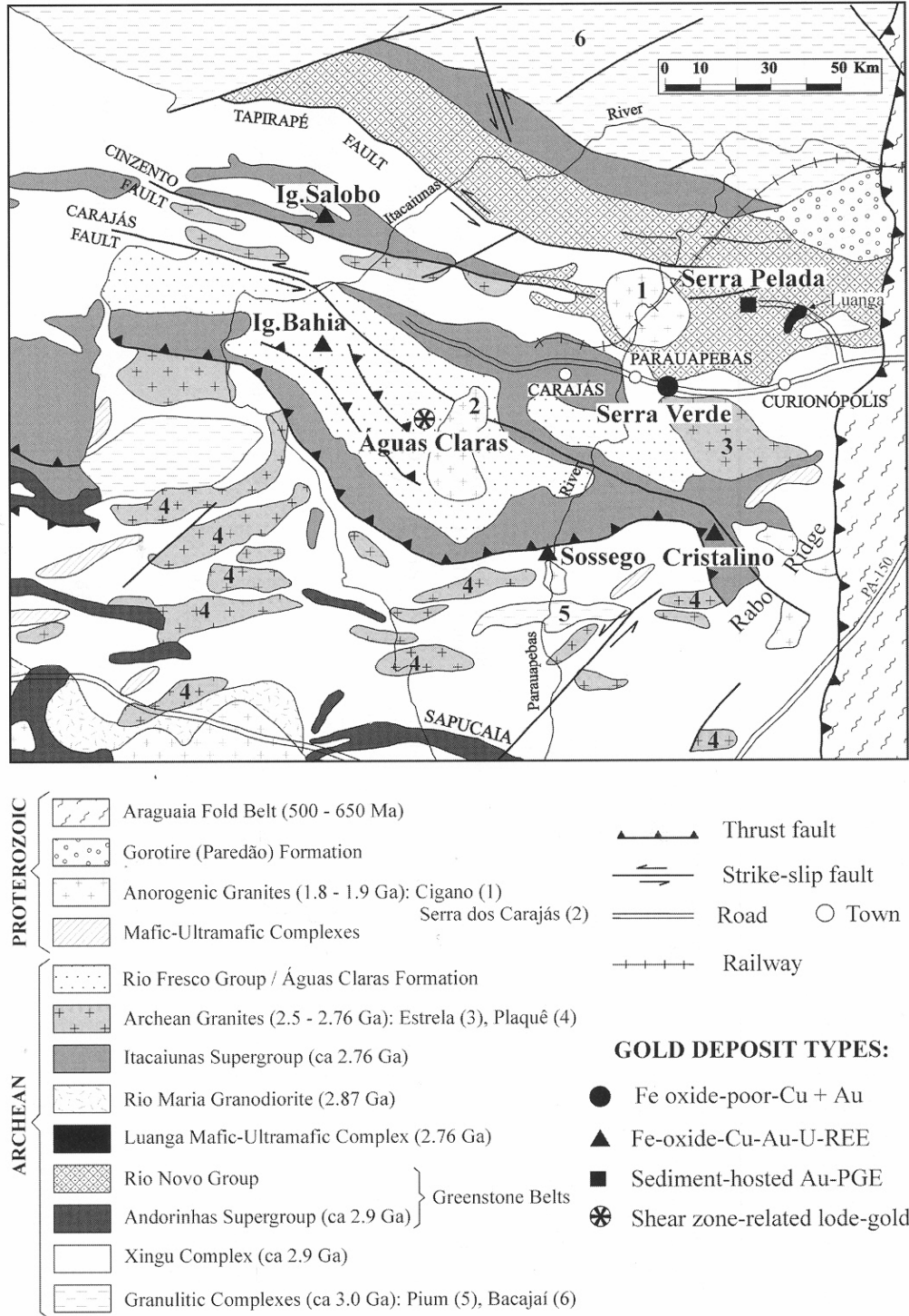
al., 1999). The deposit is hosted within metavolcanic and metasedimentary rocks that include metabasalt and andesite, arenite, siltstone, greywacke, chert and iron formation (Soares et al., 1999; Ronze et al., 2000). Mineralization occurs within breccias in strongly chloritized rocks and includes magnetite, chalcopyrite, bornite, siderite, chlorite, tourmaline, apatite, molybdenite, parisite, allanite and uraninite (Soares et al., 1999).

### **Sossego**

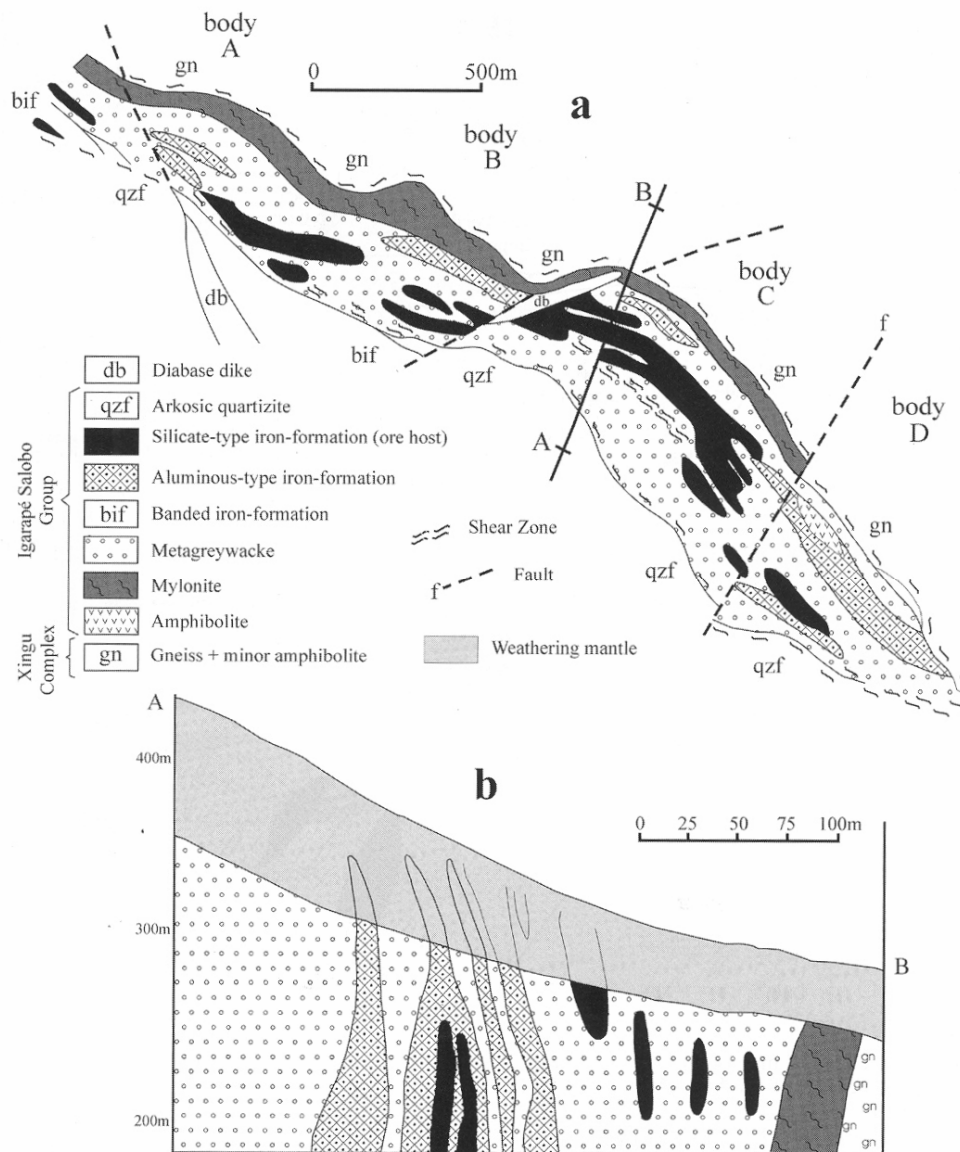
The Sossego Cu-Au deposit (Lancaster et al., 2000; Marschik et al., 2003) consists of two orebodies, Sossego Hill and Sequirinho. Mineralization is hosted within mafic and felsic volcanics and diorite which have been overprinted by albitization and some biotite-scapolite alteration. Sequirinho contains milled breccias with magnetite-amphibole infill and alteration, carbonate apatite veins and a sulphide stage dominated by chalcopyrite. Sossego also contains milled breccia, but in this case there is no fine matrix and the fragments are set in abundant hydrothermal infill material that includes carbonate, magnetite, amphibole, apatite and chalcopyrite.

### **Cristalino**

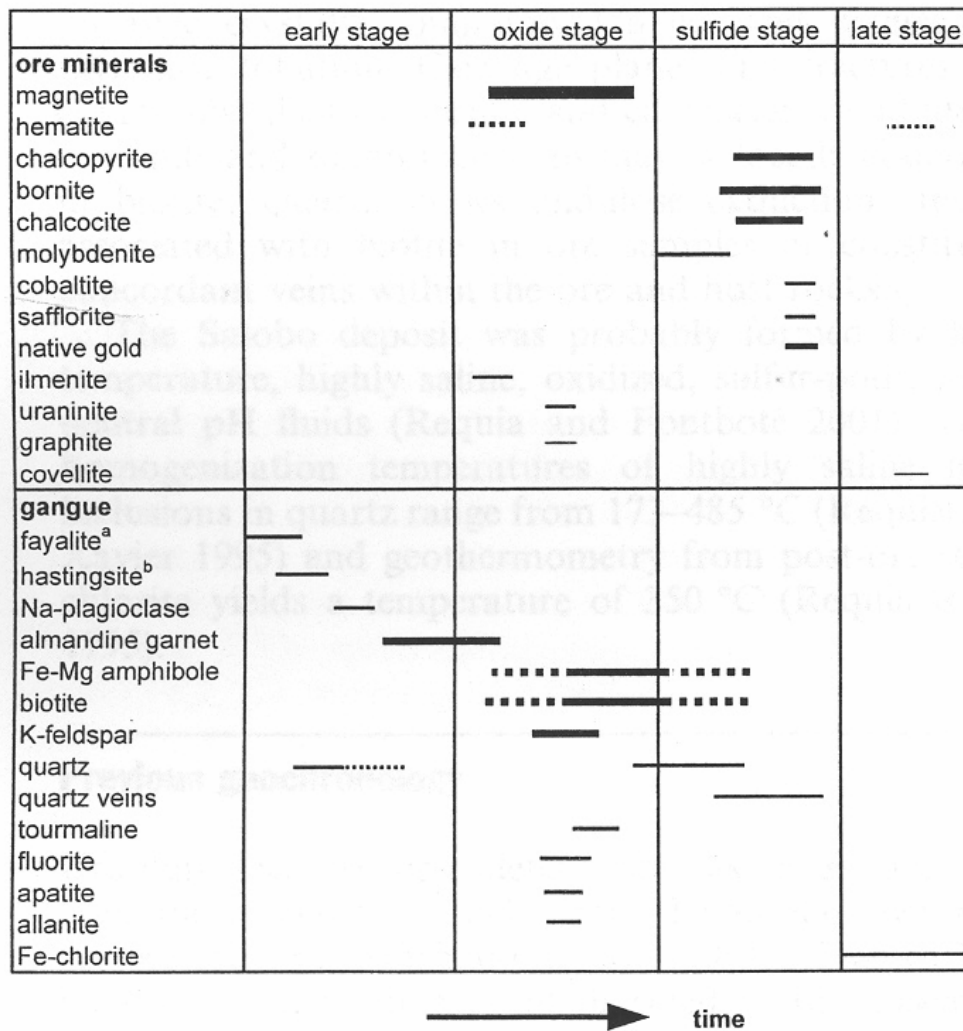
The Cristalino deposit was discovered in 1998, and is hosted within a volcanosedimentary sequence containing banded iron formation that is intruded by diorites and overprinted by hydrothermal alteration including biotite, microcline, albite, amphibole, scapolite and apatite (Huhn et al., 1999, 2000b). Mineralization is hosted within breccia zones and consists of mainly of chalcopyrite, pyrite, magnetite, bravoite, cobaltite, millerite, vesite and gold (Huhn et al., 1999). The resource is estimated at more than 500 Mt with grades of 1.0% Cu and 0.3 g/t Au (Huhn et al., 2000b).



**Figure 15.** Geological map and major gold deposits in the Carajás district (from NentunoVillas and Santo, 2001).



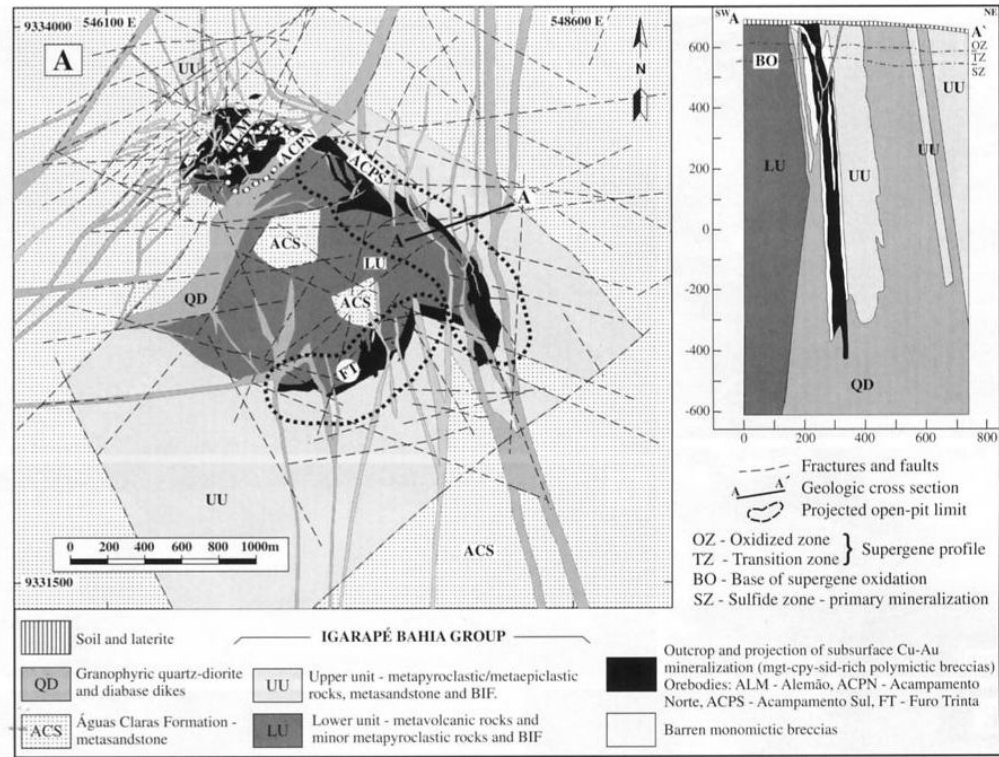
**Figure 16.** (A) Geologic map and (B) cross-section of the Salobo Cu-Au deposit (from NentunoVillas and Santo, 2001).



<sup>a</sup> pseudomorphically altered to magnetite and grunerite

<sup>b</sup> mostly replaced by grunerite or cummingtonite and magnetite

**Figure 17.** Paragenetic sequence of the Salobo Cu-Au deposit (from Requia and Fontboté, 2003).



**Figure 18** a. Geological map and b. cross section of the Igarapé Bahia deposit (from Tallarico et al., 2005).

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Iron-oxide-Cu-Au-(U-REE) deposits – A practical guide and  
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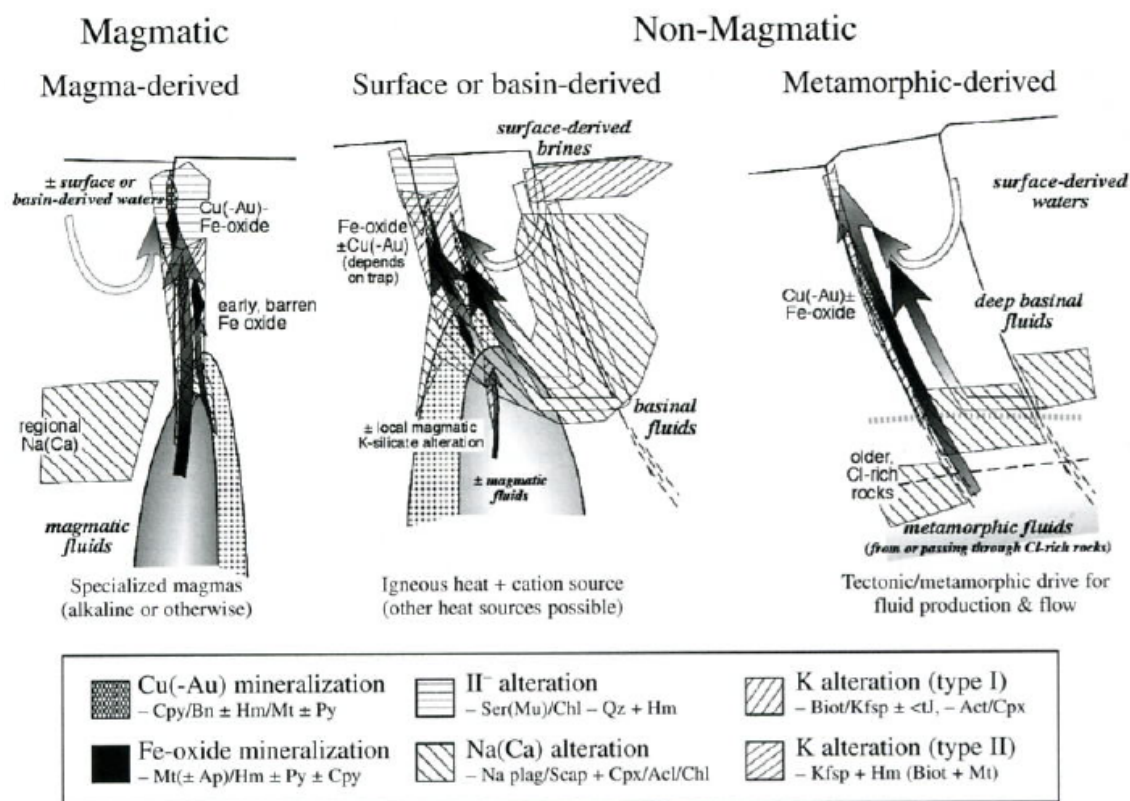
**Part 4. Metallogenic and exploration issues**

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## Metallogenic issues

Models for IOCG systems (Fig. 1) are comparatively new and exploration and research are still at an early stage in most districts. As yet there is no consensus view on the genesis of the deposits, although the growing body of knowledge does enable some comment to be made concerning their metallogenesis.

At present, the most contentious issue concerns the source of the hydrothermal fluids and the various components transported by the fluids. The main types of models that have been proposed that could account for alteration and mineralization in these systems include: (1) an evaporitic source model (e.g. Barton and Johnson, 1996), (2) a fluid heating model (e.g. Dilles et al., 1995), and (3) a magmatic-hydrothermal model (e.g. Pollard, 2000, 2001, 2006) and a metamorphic model (Fig. 1).



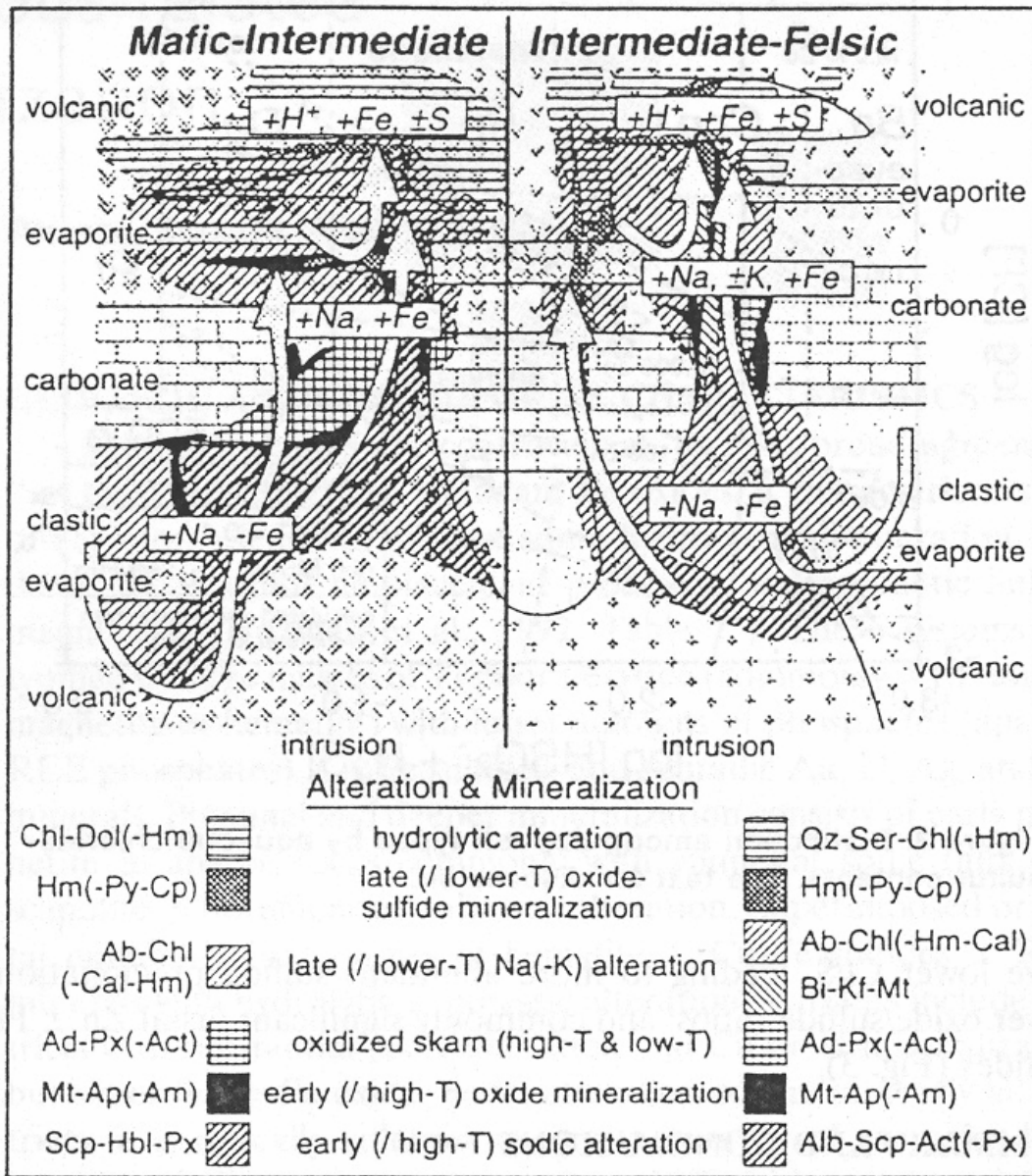
**Figure 1.** Schematic illustration of alternative models for the formation of IOCG deposits (from Williams et al., 2005).

Fluid source	Magmatic	Nonmagmatic	
		Basin/surface	Metamorphic
Fundamental processes	Release of S <sup>2-</sup> -poor metal-bearing brine from magma; rise by buoyancy Cooling, wall-rock reaction ± fluid mixing provide trap	Thermal convection of non-magmatic brines; wall-rock reaction provides metals Cooling, wall-rock reaction or fluid mixing provide trap; second fluid may provide metals	Metamorphic release of brine components by devolatilization or reaction with other aqueous fluids; rise by buoyancy Cooling, wall-rock reaction ± fluid mixing provide trap
Igneous associations	High K, oxidized suites ranging in composition from diorites to granites Carbonatite and strongly alkaline connections proposed by some	Igneous rocks diverse (gabbro to granite); nonmagmatic examples known Key heat source in most Material source, diversity reflected in geochemistry	No necessary connection, though commonly present Could be heat source in some settings Can be material source
Hydrothermal alteration in feldspathic hosts	Na(Ca) another types (K, H <sup>+</sup> ) link to magmas Regional Na(Ca) coincident but not directly related to Cu(-Au)	K (type I), H <sup>+</sup> ± Na(Ca) in upwelling zones Na(Ca) ± K (type II) in recharge zones	Primarily K and H <sup>+</sup> alteration associated with deposits Regional Na(Ca) association reflects sources
Relationship of Fe oxides to Cu(-Au)	Some Fe oxides with Cu(-Au), may be deeper or higher T equivalents Barren Fe oxides may form from distinct fluids and commonly in older hydrothermal systems in same area	Mt-rich are deeper, earlier, higher-T parts of ore-forming; Mt or Hm also typical with Cu Barren Fe oxides represent lack of S trap for Cu or lack of second Cu-bearing fluid	Fe oxides present but relatively minor (Bi or Chl common); Fe oxides commonly generated by breakdown of mafic minerals rather than Fe introduction
Local setting: depth/structure	Shallow to midcrustal levels; commonly along regional structures but near causative intrusions	In (mainly) brittle upper crust; plumbing provided by regional or volcano-tectonic structures	Mid- to shallow crustal levels near or on major structures; surface fluids require shallow levels
Global setting	Ares or extensional environments that produce characteristic magmas (oxidized high K or alkaline)	Regions with appropriate brine sources (arid settings or older Cl-rich materials), plumbing systems, and thermal drives	Regions with Cl-rich low- to intermediate-grade source rocks; compressional setting (e.g., basin-collapse) or prograde metamorphism
Key references	Hanek (1989), Pollard (2000), Groves and Vielreicher (2001), Sillitoe (2003)	Barton and Johnson (1996, 2000), Haynes et al. (1995), Haynes (2000)	Williams (1994), De Jong et al. (1998), Hitzman (2000)

**Table 1.** Summary of alternative genetic models for IOCG systems (from Williams et al., 2005).

### Evaporitic source

Barton and Johnson (1996; Fig. 2a) considered that Na-rich alteration and the distinctive element enrichment in Fe-Cu-Au(-REE-U) deposits reflect a process involving generation of hot S-poor brines by interaction of fluids with evaporite minerals to produce Na-rich alteration on the downwelling part and metal precipitation in the cooler, upwelling parts of large thermal convection systems driven by granitoid emplacement. Within this model, mineralization is a direct consequence of the presence of evaporitic rocks, even though fluids and metals may be derived from contemporaneous granites and metamorphic and/or meteoric fluid systems.



**Figure 2a.** Schematic model of igneous-driven circulation of evaporitic fluids showing alteration zoning in mafic and felsic systems (from Barton and Johnson, 1996).



## Heating external fluids

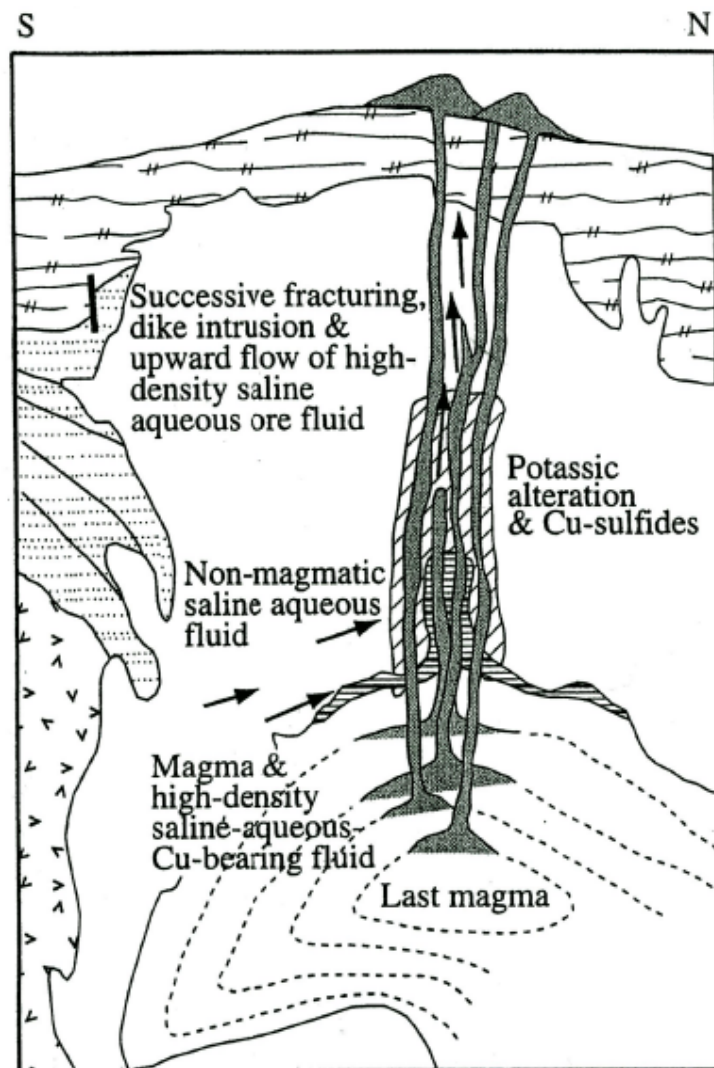
Sodic-calcic alteration is increasingly recognized as a component of porphyry-related Cu-Au deposits, e.g. El Salvador (Gustafson and Quiroga, 1995), deposits in the Yerington Batholith, Nevada (Dilles et al., 1995), and at Island Copper (Arancibia and Clark, 1996). In the Yerington Batholith, Dilles and Einaudi (1992) and Dilles et al. (1995) suggest that sodic-calcic alteration was formed from connate fluids which became heated as they were drawn toward the intrusive centres, with potassic alteration formed in the central part of the system as the fluids flowed upwards/outwards and cooled (Fig. 2b).

The Dilles et al. (1995) model relies on feldspar-fluid exchange equilibria where the Na/(Na+K) ratio of a fluid in equilibrium with alkali feldspars has an inverse relationship with temperature (Orville, 1963), i.e. in hydrothermal systems in feldspathic rocks, increasing temperature would be accompanied by albitization and decreasing temperature would be accompanied by K-feldspar alteration. The sodic-calcic alteration observed in the Yerington Batholith has removed components such as K, Fe and Cu from the altered rocks and this may account for approximately 15% of the copper in the porphyry copper systems (Dilles et al., 1995).

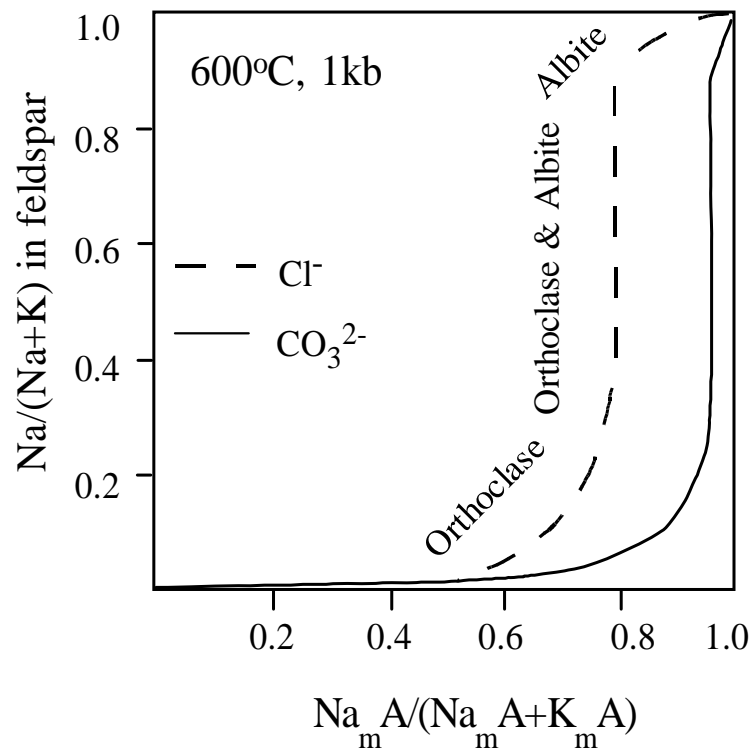
## Magmatic-hydrothermal model

Pollard (2000, 2006) proposed that Fe-oxide Cu-Au deposits are primarily products of magmatic-hydrothermal systems linked to oxidized mafic to felsic I-type magmas. Fluid inclusion and stable isotope data for sodic-calcic alteration and Fe-oxide Cu-Au deposits commonly point to a predominantly magmatic fluid source (Pollard, 2000; Perring et al., 2000; Mark et al., 2004). Hypersaline aqueous fluids and CO<sub>2</sub>-rich fluids which appear to be products of unmixing of H<sub>2</sub>O-CO<sub>2</sub>-salt fluids are commonly observed in the alteration systems (Pollard, 2000; 2001). Experimental evidence (Iiyama 1965) indicates that the Na/(Na+K) ratio of fluids in equilibrium with two alkali feldspars in CO<sub>3</sub><sup>2-</sup>-bearing fluids (0.96 at 600°C; 1 kilobar) is significantly higher than in chloride-bearing fluids (0.79)(Fig. 3). Unmixing of H<sub>2</sub>O-CO<sub>2</sub>-salt fluids caused by decreases in temperature and/or pressure will therefore result in albitization of the wall rocks, and may be succeeded at lower temperature by K-feldspathization due to the temperature dependence of Na/(Na+K) in chloride-bearing fluids buffered by alkali feldspars (Pollard, 2001). The low solubility of CO<sub>2</sub> in granitic magmas (Holloway 1976) suggests that pressure decrease during upward intrusion of the magmas may cause CO<sub>2</sub> saturation, although the addition of CO<sub>2</sub> from associated basic magmas and/or devolatilization of carbonate-bearing host rocks could also play a role. Saturation of the magma in CO<sub>2</sub> would result in partitioning of water, chlorine and metals to the fluid phase.

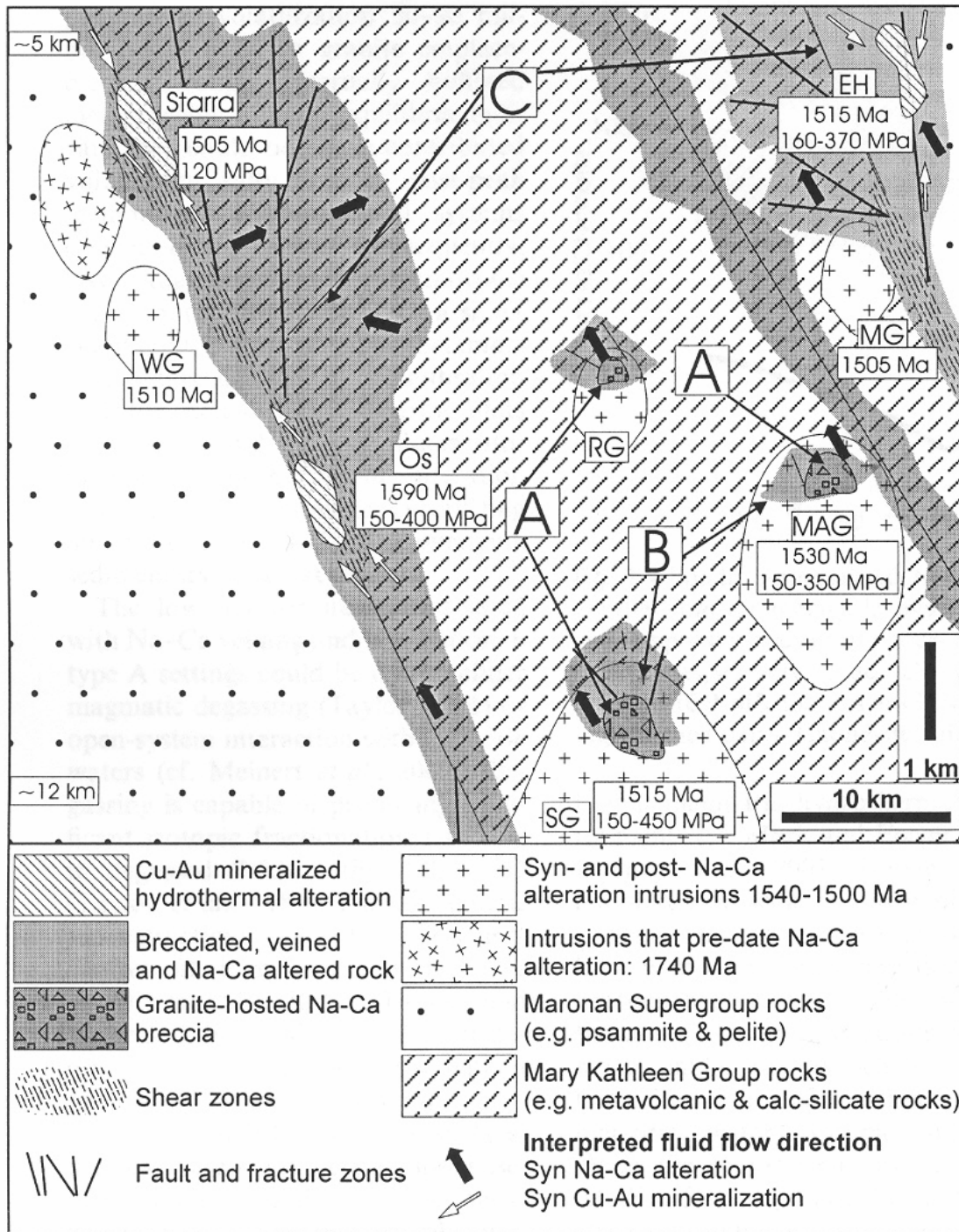
The extensive hydrothermal breccias associated with sodic(-calcic) alteration in the Cloncurry district (Fig. 4) may have formed by volume expansion during fluid unmixing where brecciation was primarily caused by escape of the CO<sub>2</sub>-rich vapour phase, and albitization was primarily due to later infiltration of the salt-rich liquid phase (Pollard, 2000a). Stable isotope study of the sodic-calcic altered rocks indicates a dominantly magmatic composition for the hydrothermal fluids (Mark et al., 2004; Fig. 5a). With decreasing temperature, the equilibrium Na/(Na+K) ratio of the fluid increases and the fluids would therefore cause potassic (or sericitic) alteration of the wall-rocks, consistent with observations of paragenetically later and/or shallower K-feldspar and/or sericitic alteration in Fe-oxide Cu-Au systems (see Hitzman et al., 1992).



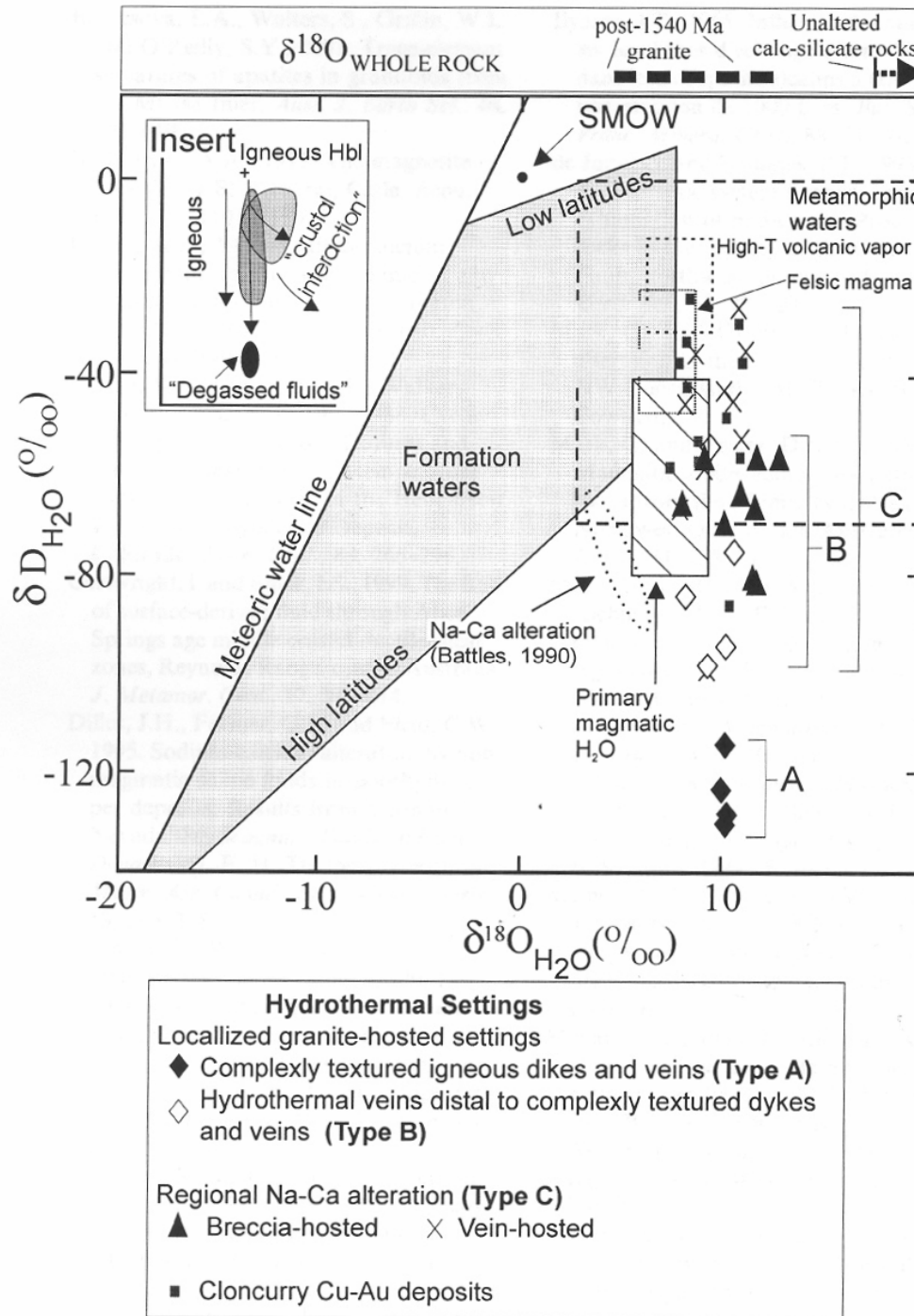
**Figure 2b.** Fluid flow paths and alteration styles in the Yerington Batholith (from Dilles, 1987).



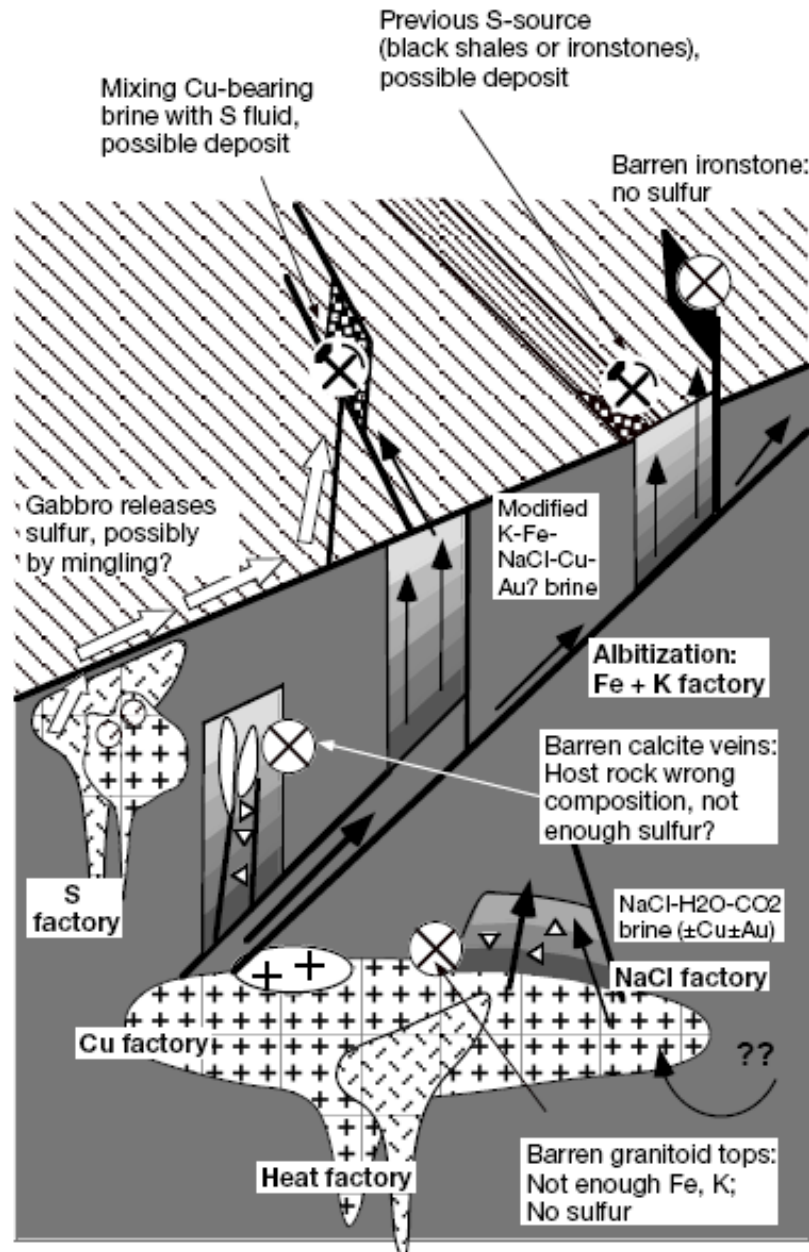
**Figure 3.** Exchange equilibria between alkali feldspars and K-Na solutions with Cl<sup>-</sup> and CO<sub>3</sub><sup>2-</sup> anions at 600°C and 1 kilobar. The Na<sub>m</sub>A/(Na<sub>m</sub>A+K<sub>m</sub>A) values are concentrations in solution, where A is the anionic species and m is the number of cations. Modified from Iiyama (1965).



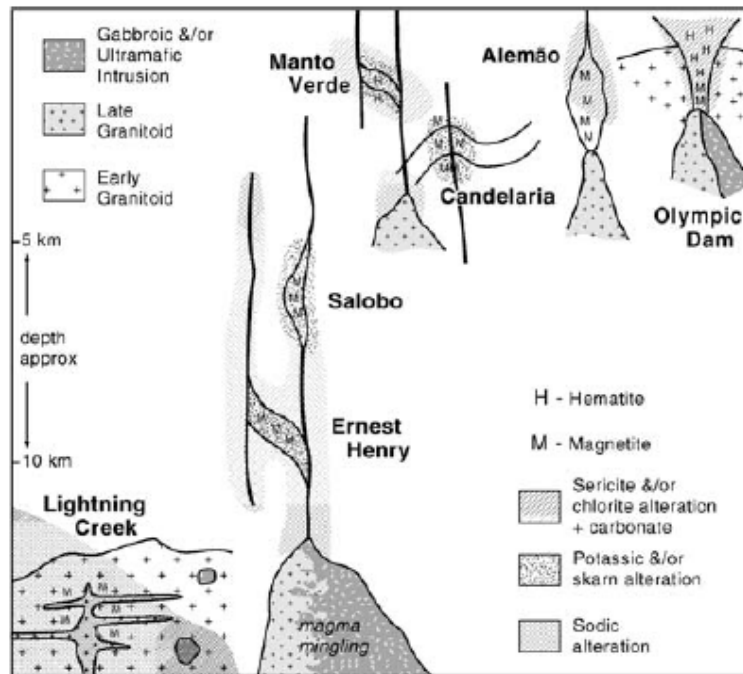
**Figure 4.** Schematic cross section showing spatial relationships between the main types of Na-Ca alteration in the Cloncurry district (from Mark et al., 2004).



**Figure 5a.** Calculated isotopic composition of fluids associated with sodic-calcic alteration (from Mark et al., 2004).



**Figure 5b.** Model for IOCG mineralization in the Cloncurry district (Oliver et al., 2004).



**Figure 5c.** Schematic model to show the relationships between different deposits and the major alteration styles in IOCG systems (from Pollard, 1996).

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## **Ingredients in exploration models**

### **Fertile intrusive rocks**

Depending on the preferred genetic model, the role of intrusive rocks in these systems is quite variable. However, at this stage it appears that mildly alkaline to subalkaline, magnetite-series mafic to intermediate magmas that have not undergone significant biotite fractionation are most closely linked to Fe-oxide Cu-Au deposits. Although low-K rocks have been implicated in some districts (e.g. Vidal et al., 1990), evidence from porphyry Cu-Au systems suggests that high-K suites are likely to be more fertile (Sillitoe, 1997). Pollard (2006) has argued that similar suites of intrusive rocks are linked to porphyry and Fe-oxide Cu-Au deposits, even though trace element differences are evident due to different tectonic settings and magma sources. For example, U and REE enrichment is a feature of intrusive rocks in the Proterozoic back arc environments, and these elements are commonly enriched in Fe-oxide Cu-Au deposits in these environments (e.g. Olympic Dam; see Pollard, 2006).

### **Structural controls**

At the broad scale, major, commonly transcurrent fault systems appear to be the major controls on the localization of intrusive rocks and mineralization. At the local scale, a variety of brittle and brittle-ductile structural situations have the potential to host mineralization, and the most common of these include combinations of fracturing, fracture intersections and fracture-lithology intersections which have resulted in the development of brecciation and/or extensional sites for fluid flow and precipitation. Commonly it appears to be relatively massive, competent lithologies such as ironstone, amphibolite, volcanics, granitoids and albitized rocks that have been brecciated while zones of shearing strain have commonly anastomosed around the rigid bodies. For example, Ernest Henry is located in volcanic rocks that comprise perhaps the most extensive group of fine-grained relatively isotropic rocks in the Cloncurry district. These types of rocks offered an opportunity for large-scale brecciation because they contained minimal anisotropies (bedding, cleavage) along which the strain could be partitioned. Ore zones at the Osborne deposit are selectively developed adjacent to ironstones as a result of dilation during faulting subparallel to contacts between ironstone and feldspathic psammite (Adshead, 1995).

Several of the major Fe-oxide Cu-Au deposits are hosted within breccias that contain textures indicative of fluidization. These breccias are composed of rounded to angular fragments of different (mixed) rock types within variable proportions of rock flour matrix. Mineralization may occur as fragments in the breccias and/or accompanying overprinting alteration. Examples include Olympic Dam, Australia (Reeve et al., 1990), Manto Verde, Chile (Vila et al., 1996), Susana, Chile (Espinoza et al., 1996), Mount Dore, Australia (Beardmore, 1992) and Kuridala, Australia (Laing, 1998). These hydrothermal intrusive breccia/diatreme systems may reflect escape of the vapour phase from an underlying intrusive and/or local boiling within fault/fracture systems

### **Physicochemical controls**

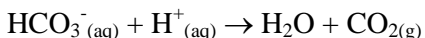
A very wide range of physicochemical controls have been proposed to account for mineralization in Fe-oxide Cu-Au systems. The physicochemical characteristics of hydrothermal fluids responsible for mineralization commonly favour transport of Cu and Au as



chloride complexes, and the high temperature, high salinity character of the hydrothermal fluids suggests that the most likely controls on Cu and Au deposition relate to processes that contribute to decreasing temperature and pressure, changes in oxidation state and/or changes in pH (e.g. Davidson and Large, 1994). Among the wide variety of factors that could cause changes in these parameters, the most important appear to be cooling, fluid unmixing, wall rock interaction and fluid mixing. The key underlying factors are the requirements for a supply of Cu and Au in a physicochemical regime where their solubilities decrease in tandem with the availability of reduced sulphur for sulphide precipitation.

#### Fluid unmixing

Unmixing of highly saline H<sub>2</sub>O-CO<sub>2</sub> fluids has been proposed by Ettner et al. (1993) as an important mechanism for metal precipitation in shear zone-hosted Cu-Au deposits in Norway. Fluid unmixing resulted in formation of a CO<sub>2</sub>-rich fluid and a highly saline H<sub>2</sub>O fluid, and may involve an increase in pH of the aqueous fluid due to the reaction:

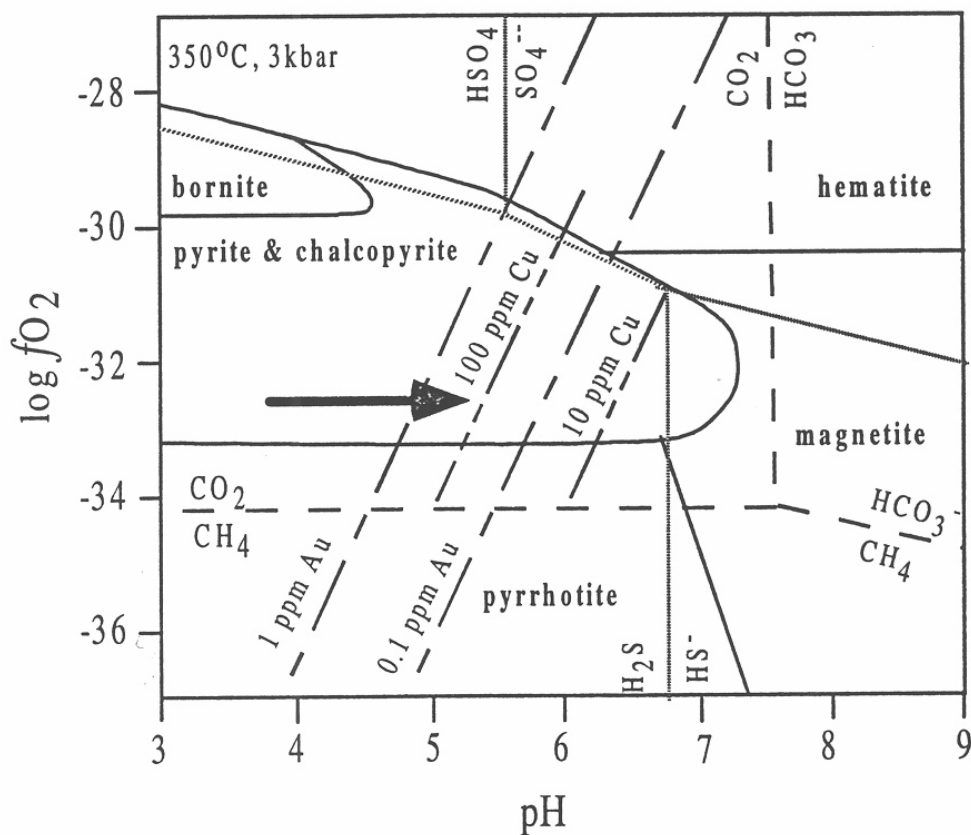


An increase in pH contributes to precipitation of metals transported as chloride complexes due to decreased solubility. Over a wide range of oxygen fugacity (pyrite-stable, magnetite-stable and pyrrhotite stable) the magnitude of the decrease in the solubility of Cu and Au chloride complexes due to increasing pH is very similar (e.g. Fig. 6). The presence of hypersaline and CO<sub>2</sub>-rich fluids in Fe-oxide Cu-Au deposits appears to be ubiquitous (Pollard, 2000 and 2001), and fluid unmixing is potentially a significant mechanism of metal precipitation.

#### Fluid mixing

Fluid mixing can be an effective metal precipitation mechanism through potentially affecting pH, oxygen fugacity and temperature. Deposits such as Olympic Dam (Haynes et al., 1995) and the hematite-sericite systems of northern Chile may have formed by mixing of magmatic and meteoric fluids at moderate to low temperature in shallow level environments. This has the potential to create zoned sulphide assemblages, to fractionate Cu and Au, and to create extensive barren hematite-rich zones.

Of perhaps greater significance is the possibility that the original magmatic fluids carried copper but lacked sulphur (e.g. Perring et al., 2000). The presence of significant barium in inclusion fluids (Perring et al., 2000; Williams et al., 2001) under oxidized conditions could be taken to imply that the fluid contained negligible sulphur and that the formation of a Cu-Au deposit requires mixing with a sulphur-bearing fluid or interaction with a sulphur-bearing rock (e.g. Fig. 5b)



**Figure 6.** Log  $fO_2$  versus pH diagram for the Fe-S-O system for 350°C and 3 Kbar for 25 wt.% NaCl solution with  $\Sigma S = 0.01m$  and showing Cu- and Au-chloride solubilities (modified from Ettner et al., 1993). Arrow shows possible fluid path caused by unmixing of  $CO_2$ -rich vapour phase from an  $H_2O$ - $CO_2$ -salt fluid.

Many Fe-oxide Cu-Au deposits have fluid inclusion evidence to suggest that Na-Ca-rich fluids were present around the time of metal precipitation (e.g. Pollard, 2000 and 2001). The role of these fluids in mineralization is uncertain, although they may reflect dilution of the hypersaline fluids by low salinity fluids and/or introduction of a saline fluid from the wall rocks. It is also possible that such fluids may have introduced sulphur for sulphide precipitation, given the sulphur isotope evidence for derivation of sulphur from the host sequences in some cases (Pollard et al., 1997; Ullrich and Clark, 1999)

#### Fluid-wall rock interaction

Interaction between hydrothermal fluids and wall rocks can be a major factor in ore deposition since it may result in changes in pH and/or oxygen fugacity that decrease the solubility of ore components in the fluid. Interaction of hot, relatively oxidized fluids with more reduced host rocks, including graphitic rocks, amphibolites and magnetite ironstones (or magnetite-rich alteration), can be an effective mechanism for Cu and Au deposition. Oxidation of magnetite ironstones by magmatic-hydrothermal fluids in equilibrium with hematite has been proposed by Huston et al. (1993) as a mechanism for metal precipitation in Tennant Creek Cu-Au-Bi deposits. A decrease in  $fO_2$  of the fluid during alteration caused reduction of  $SO_4^{2-}$  to  $H_2S$ , and

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was accompanied by increasing pH which caused Cu and Au precipitation (Huston et al., 1993).

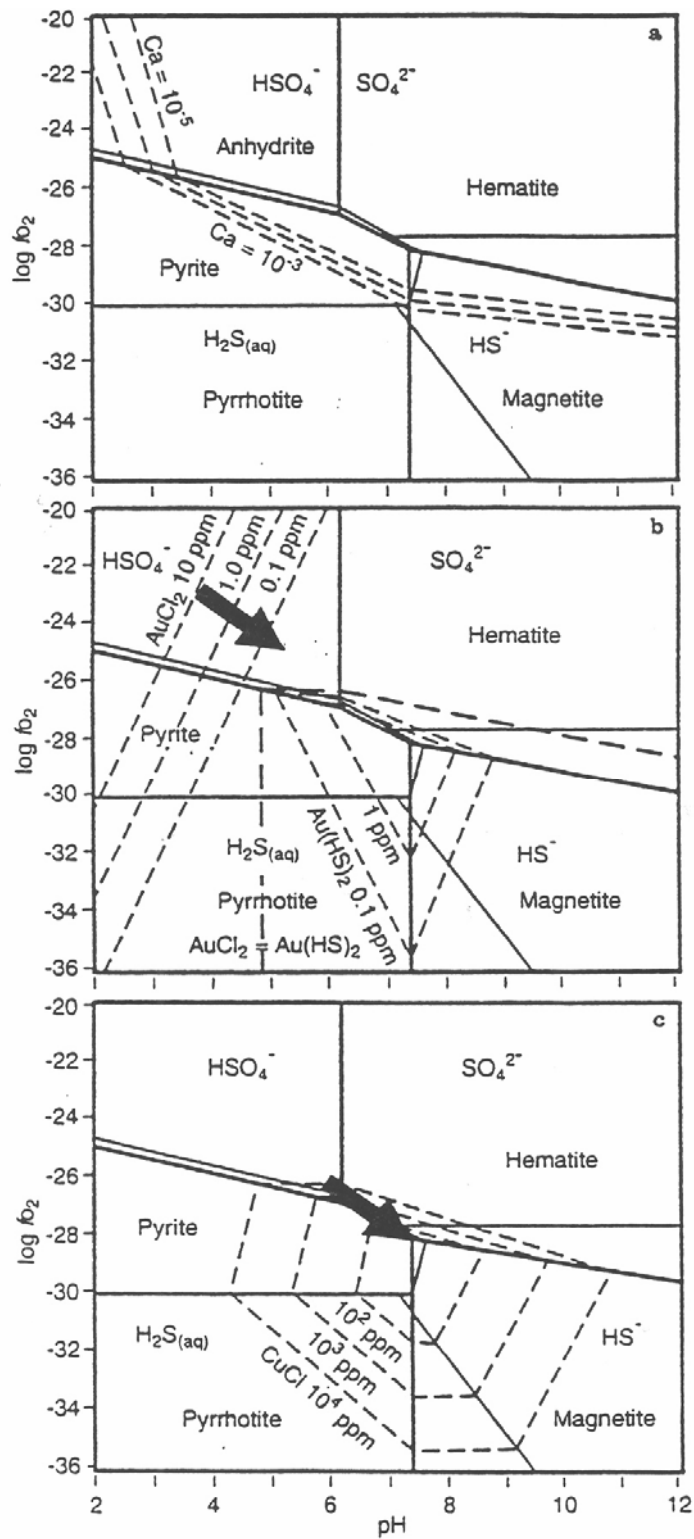
The gold-rich Starra orebodies in the Cloncurry district occur dominantly within and adjacent to magnetite ironstones where they have been intersected by late brittle and brittle-ductile structures. The ore stage paragenesis at Starra is characterized by oxidized assemblages including anhydrite-calcite-hematite-gold-pyrite-chalcopyrite-magnetite and localized barite, bornite and chalcocite. Copper and gold mineralization is spatially associated with magnetite, and commonly occurs where magnetite is replaced by hematite (Rotherham, 1997). The high Au:Cu ratio has been attributed to preferential Au deposition by reaction of the oxidized ore fluid with pre-existing magnetite ironstones, followed by precipitation of pyrite and chalcopyrite as the fluid became progressively more reduced (Rotherham, 1997; Fig. 7).

A role for carbonaceous material in the precipitation of mineralization has also proposed for Fe-oxide Cu-Au deposits, including the Greenmount and Mount Dore deposits in the Cloncurry district (Krcmarov, 1995; Beardsmore, 1992). The Greenmount deposit is localized near a major fault that separates carbonaceous metapelites of the Marimo Shale from scapolite marbles (meta-evaporites) and metasiltstones of the Staveley Formation (Stewart, 1994; Krcmarov, 1995). The stratigraphic contact is the locus of mineralogically complex alteration systems that host Cu-Au mineralization in quartz-K feldspar bearing veins, with pyrite and chalcopyrite (+Au and arsenopyrite) being concentrated in the altered carbonaceous rocks.

The Mount Dore Cu deposit is hosted by a fault-bounded slice of carbonaceous mica schists sandwiched between calcareous and calc-silicate rocks of the Staveley Formation (Mary Kathleen Group) to the west, and the Mount Dore Granite to the east (Beardsmore, 1992). Mineralization and associated alteration are localized by tabular dilational breccia bodies, with sulphide precipitation related to sulphate reduction by the carbonaceous host (Beardsmore, 1992) in conjunction with other processes.

Davidson (1998) proposed that amphibolites may affect a chemical control on ore deposition by increasing fluid pH during alteration of amphibole to biotite, as is observed in Mount Freda style deposits in the Cloncurry district.

Sulphide alteration of earlier minerals has the potential to create higher grades than if the sulphides occur only as infill. Sulphidation appears to be common where Fe-rich lithologies and/or Fe-rich alteration minerals are present. For example, in the Eloise deposit (Baker and Laing, 1998) early hornblende-biotite-rich alteration and infill appears to have been comparatively susceptible to sulphidation. Similarly, at the Alemão deposit, chalcopyrite alteration of pre-existing chlorite and siderite alteration appears to have been a significant factor in creating some of the high grade zones.



**Figure 7.** Log oxygen fugacity-pH diagrams calculated for 350°C and 1000 bars pressure to illustrate the likely fluid pathway for Starra ore stage fluids. From Rotherham, 1997.

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## Comments on geophysical prospecting

A common misconception concerning Fe-oxide Cu-Au deposits is that mineralization is coincident with magnetite. Aeromagnetic data are commonly used as a prime exploration technique with the result that large numbers of magnetite bodies (and basic intrusives) are found. However, magnetics is not always effective in finding ore because the Cu-Au mineralization may have a different timing to magnetite, and/or the system may be zoned (Williams and Pollard, 2001; Smith, 2002; Carter et al., 2003). The sorts of relationships between magnetite, sulphides and Cu-Au mineralization in Fe-oxide Cu-Au systems can be illustrated by deposits in the Cloncurry district (Fig. 8), where the nature of the principal iron-bearing minerals is variable. Similarly, the geochemistry of these deposits is also highly variable (Figs. 9 and 10).

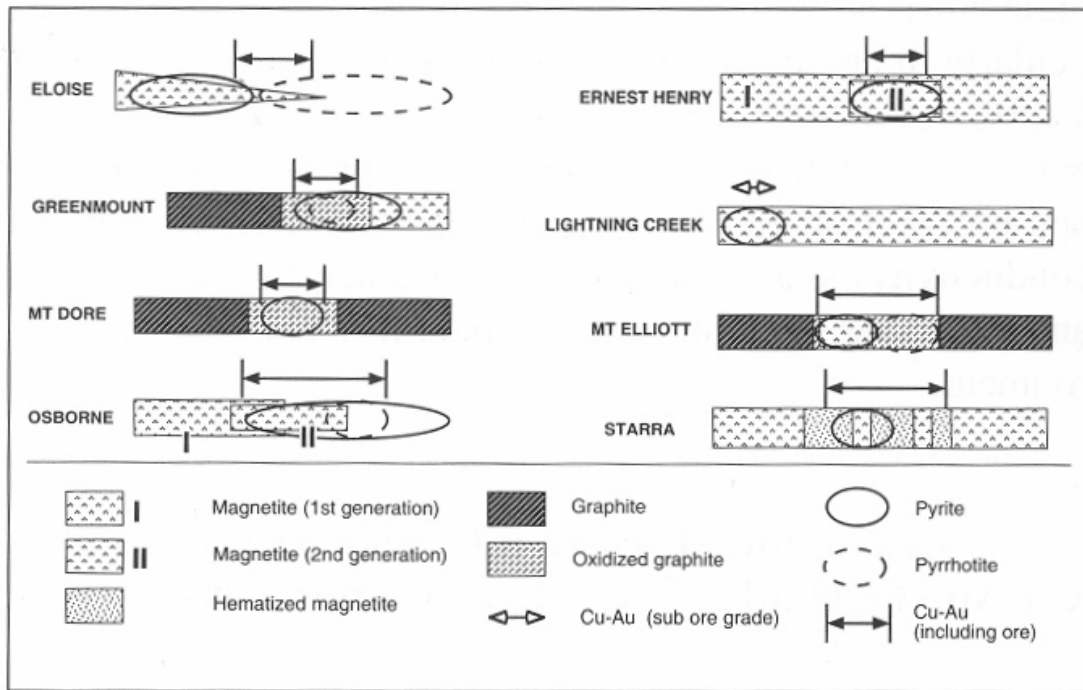
At Osborne, the western ironstones were probably formed well before metamorphism and deformation and are unrelated genetically to the Cu-Au mineralization. The spatial association between some mineralization at Osborne and the magnetite ironstones may be due to the competency contrast between ironstone and psammites which led to development of zones of brecciation along the contact (e.g. Adshead, 1995). Magnetic data may be effective in finding the ironstone, but could not locate the Eastern High Grade zone which contains only minor magnetite (see Anderson and Logan, 1992).

In zoned systems such as Eloise, drilling based on magnetic data may not be effective in finding the Cu- and Au-rich parts of the system (Figs. 11-13). Similarly, electrical methods may be effective in finding the sulphide-rich part of the system, but not the Cu-Au-rich part.

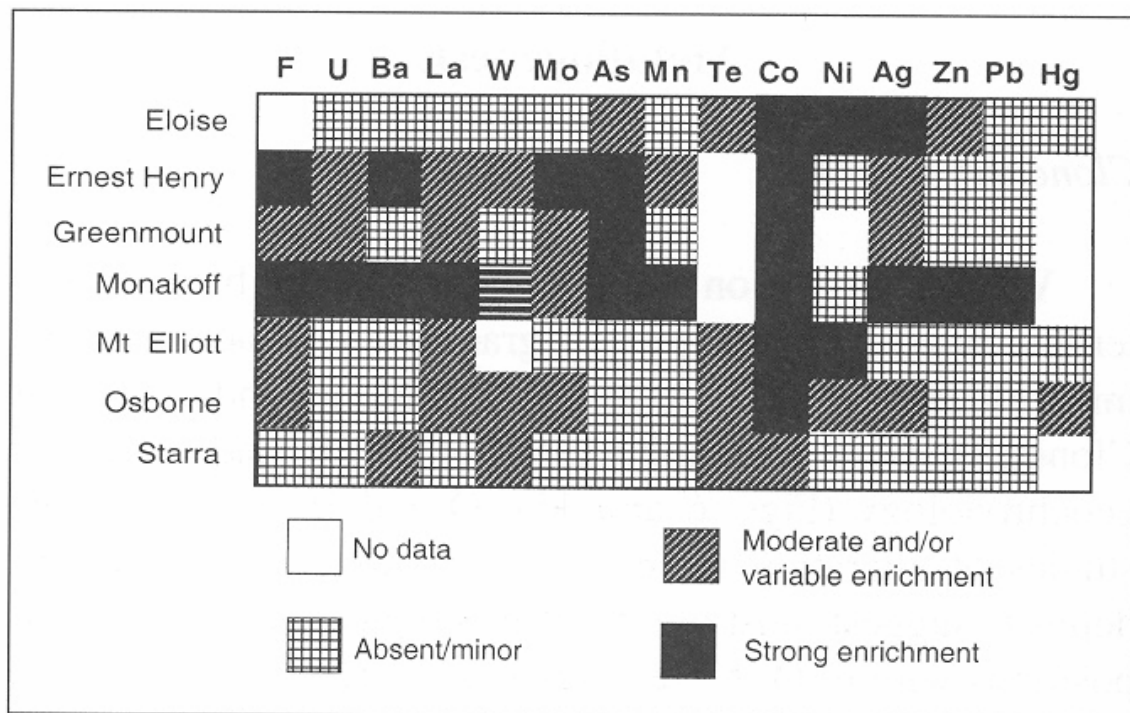
At Ernest Henry, there is a strong correlation between sulphide, Cu, Au and magnetite. However, the deposit is associated with development of other, essentially barren magnetite alteration in the surrounding area which has magnetic response similar in magnitude to Ernest Henry (Fairall, 1994; Fig 14). It is possible that the deposit might not have been found were it not for the supergene native copper which was responsible for the discrete TEM anomaly that was initially drilled (Webb and Rowston, 1995). Olympic Dam and Candelaria may be more conspicuous on magnetic images (Figs. 15 and 16), but the cause of the magnetic anomaly at Olympic Dam is uncertain.

Deposits such as Mt Elliott, which have more oxidized and more reduced portions (Wang and Williams, 2001), may have Cu-Au mineralization associated with magnetite-rich zones and/or with pyrrhotite-rich zones. Other deposits, such as Starra, have high-grade gold mineralization associated with zones containing little sulphide and abundant hematite relative to magnetite (Rotherham, 1997).

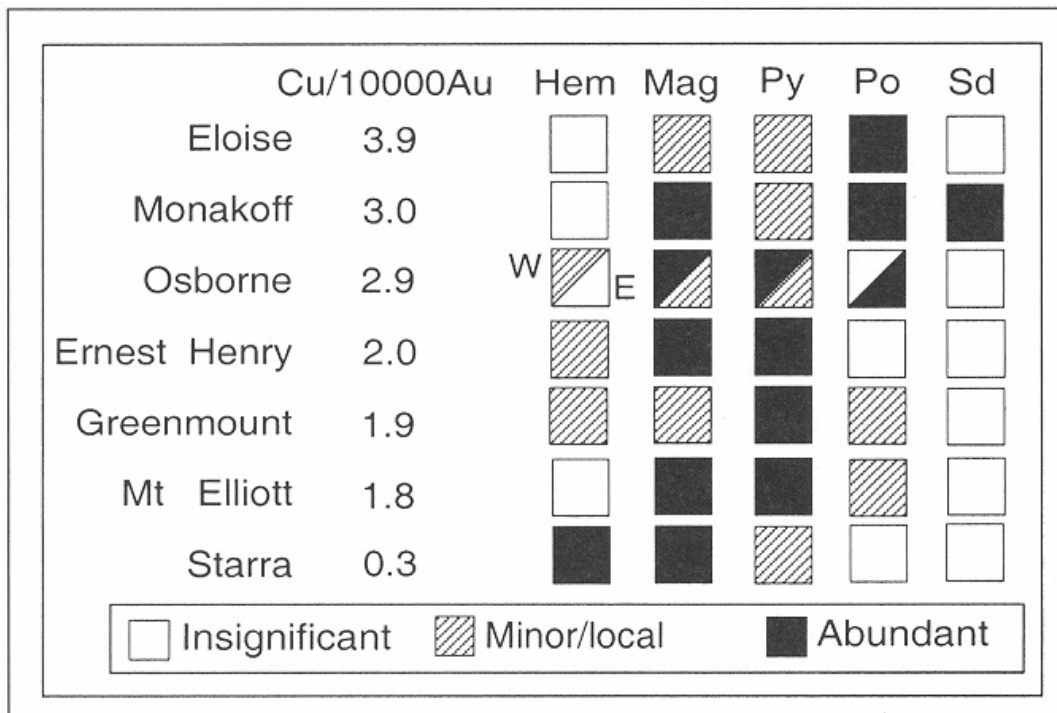




**Figure 8.** Schematic representation of the spatial relationships of Cu-Au mineralization to magnetite, sulphide and graphite in the Cloncurry district (from Williams and Pollard, 2001).



**Figure 9.** Empirical representation of element enrichments in selected Cloncurry district Cu-Au deposits (from Williams and Pollard, 2001).



**Figure 10.** Correspondence of Cu:Au ratios of Cloncurry district resources with variations in the assemblages of Fe phases that coexist with chalcopyrite in the ore (from Williams and Pollard, 2001).



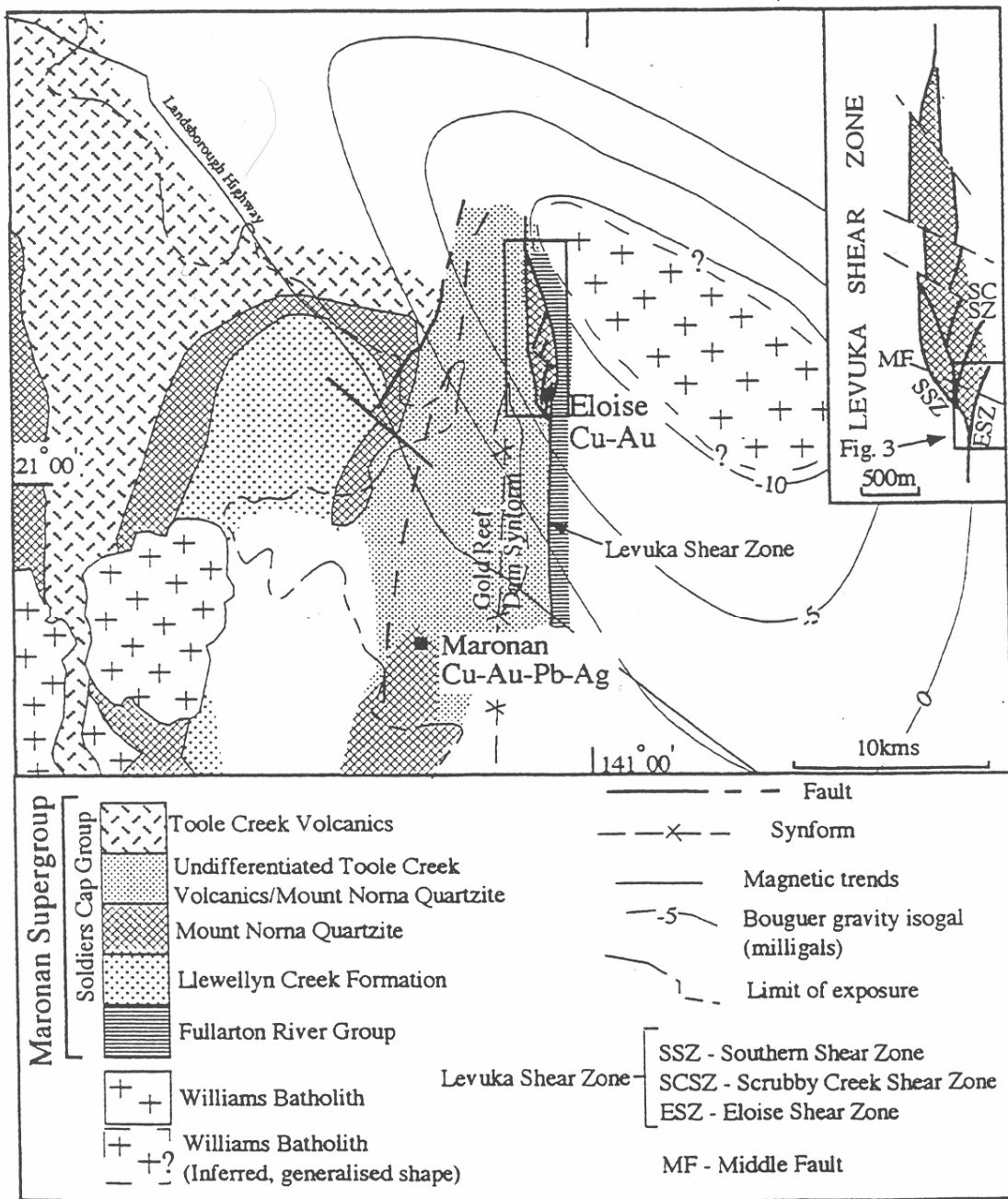
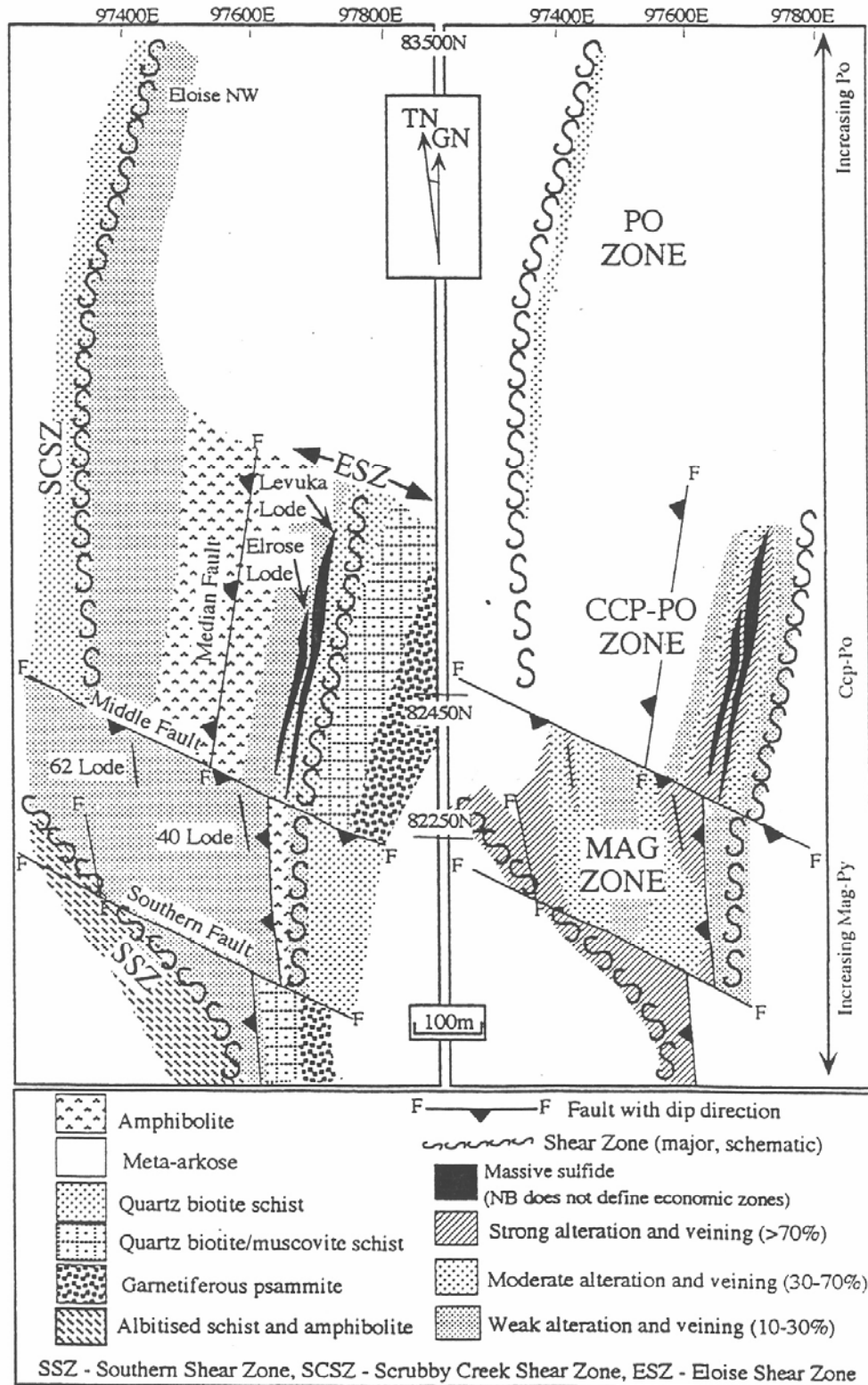
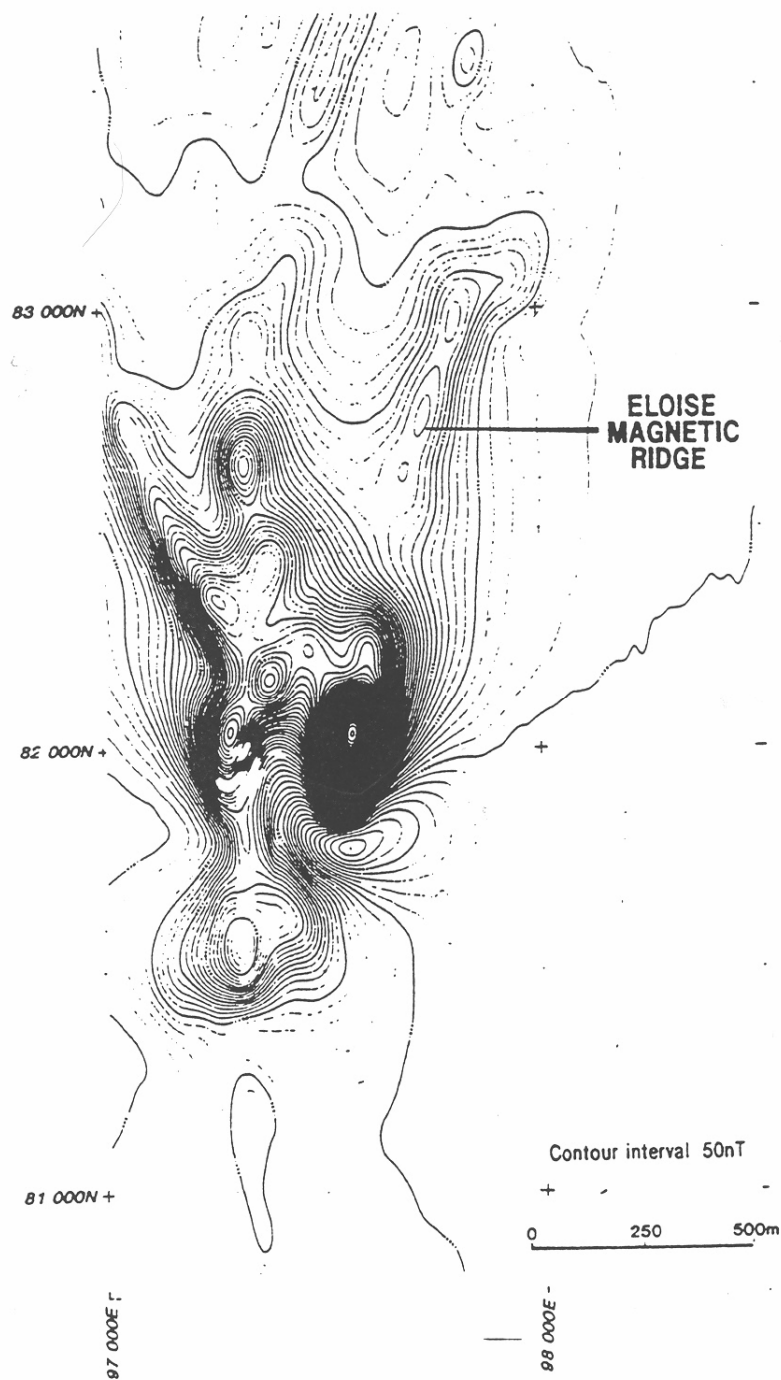


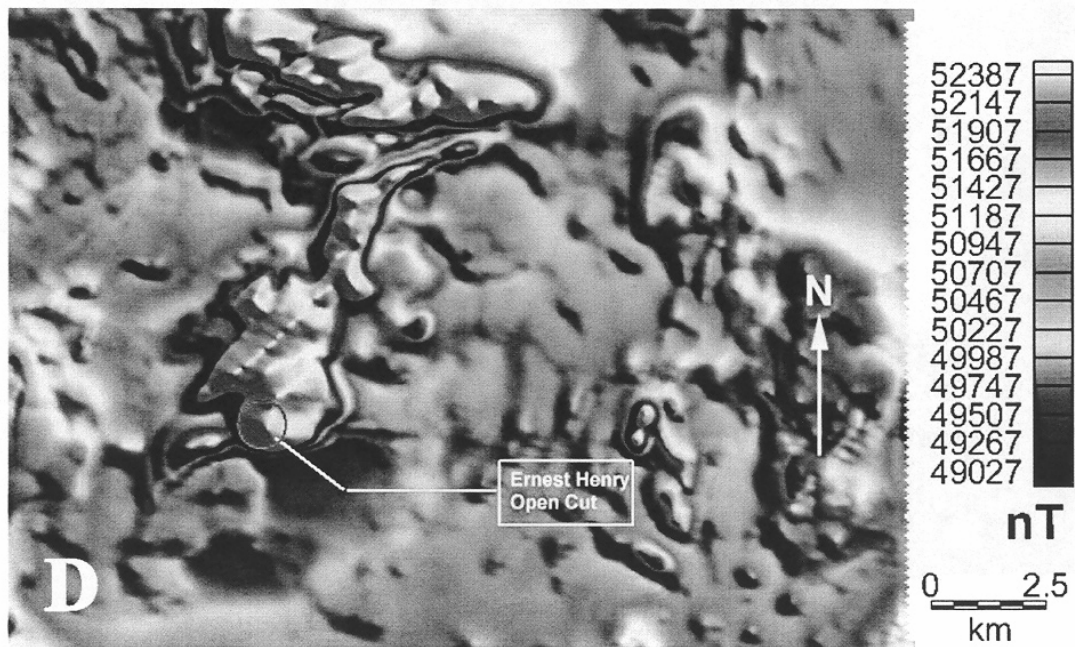
Figure 11. Geological setting of the Eloise Cu-Au deposit (from Baker and Laing, 1998).



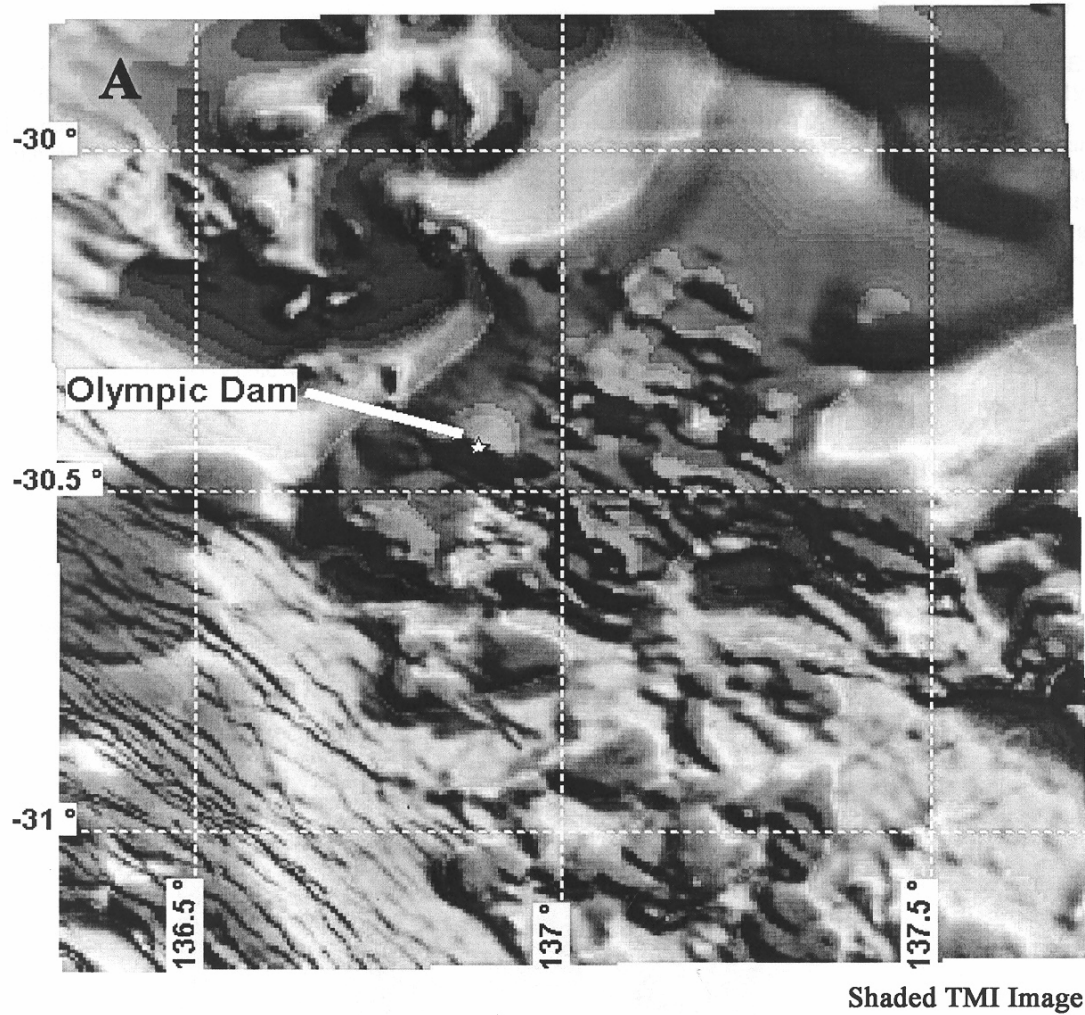
**Figure 12.** Geology and alteration map of the Eloise deposit at -100RL (from Baker and Laing, 1998).



**Figure 13.** Total magnetic intensity image of the Eloise Cu-Au deposit (from Brescianini et al., 1992).



**Figure 14.** Total magnetic intensity image over the Ernest Henry Cu-Au deposit (from Smith, 2002).



**Figure 15.** Shaded total magnetic intensity image of Olympic Dam (from Smith, 2002).



**Figure 16.** Airborne magnetics (reduced to pole) for the Candelaria Cu-Au deposit (from Smith, 2002).

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## Conclusions

The magmatic, structural and physicochemical controls on Fe-oxide Cu-Au mineralization outlined above have important implications for exploration. The different characteristics of individual deposits appear to be a reflection of various interactions of these factors. In the most simplified case, deposit formation involves: 1) escape of magmatic-hydrothermal fluids from crystallizing granitoids, 2) channelized flow of these fluids along pre-existing, or newly formed structures, and 3) precipitation of metals due to processes such as unmixing, mixing and/or fluid interaction with wall rocks. The large size of some deposits may reflect metal precipitation in favourable structural sites via fluid mixing (e.g. Olympic Dam, Candelaria). The high grade character of other deposits may have been controlled by chemical interactions with the host rocks. The gold-rich character of some parts of the Tennant Creek and Starra deposits may reflect oxidation of earlier-formed magnetite accompanied by progressive reduction of the fluid phase. In rock buffered situations such as this, it is unlikely that very large deposits will be formed unless there is intense fracturing of the rocks to provide fluid access to a very large surface area of the reactive rock.

Geophysical prospecting needs to take account of the highly variable relationships between magnetite, sulphides and Cu-Au mineralization. Exploration for iron-oxide Cu-Au deposits therefore needs to be carefully thought out, as poor initial results may reflect drilling in the wrong part of the system rather than a lack of economic potential (e.g. Prominent Hill, see Carter et al., 2003).

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