



Iron Oxide Copper-Gold Deposits: Geology, Space-Time Distribution, and Possible Modes of Origin

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Abstract

Many diverse ore systems are classified together as iron oxide copper-gold (IOCG) deposits based on an empirical definition arising primarily from geochemical features that do not specify tectonic setting, geologic environment, or sources of ore-forming fluid, metals, or other ore components. Such deposits have (1) Cu, with or without Au, as economic metals; (2) hydrothermal ore styles and strong structural controls; (3) abundant magnetite and/or hematite; (4) Fe oxides with Fe/Ti greater than those in most igneous rocks and bulk crust; and (5) no clear spatial associations with igneous intrusions as, for example, displayed by porphyry and skarn ore deposits.

IOCG deposits commonly have a space-time association with Kiruna-type apatite-bearing oxide Fe ores and many examples of the latter contain sulfide minerals, Cu, and Au. Most IOCG deposits display a broad space-time association with batholithic granitoids, occur in crustal settings with very extensive and commonly pervasive alkali metasomatism, and many are enriched in a distinctive, geochemically diverse suite of minor elements including various combinations of F, P, Co, Ni, As, Mo, Ag, Ba, LREE, and U. Iron oxide Cu-Au systems are numerous and widely distributed in space and time; they occur on all continents and range in age from the present at least back into the Late Archean. In economic terms, the most important IOCG deposits are those in the Carajás district, Brazil (Archean, Amazon craton); in the Gawler craton and Cloncurry districts, Australia (late Paleoproterozoic to Mesoproterozoic debated intracratonic or distal subduction-related settings), and in the Jurassic-Cretaceous extended continental margin arc of the coastal batholithic belt in Chile and Peru. IOCG deposits and associated features define distinct metallogenic belts in which other types of Cu and Au deposits are rare or absent. The largest deposits include Salobo, Cristallino, Sossego, and Alemão (Carajás), Olympic Dam (Gawler craton), Ernest Henry (Cloncurry district), and Candelaria-Punta del Cobre and Manto Verde (Chile), and have resources greater than 100 million metric tons (Mt), ranging up to more than 1,000 Mt with metal grades that exceed those in most porphyry-style Cu ± Au deposits.

A comparison of larger and well-described IOCG deposits illustrates the geologic diversity of the class as a whole. They occur in a wide range of different host rocks, among which plutonic granitoids, andesitic (meta)volcanic rocks, and (meta)siliclastic-metabasic rock associations are particularly prominent. Host rocks may be broadly similar in age to the ore (e.g., Olympic Dam, Candelaria-Punta del Cobre, Raul-Condestable) but in other cases significantly predate mineralization such that ore formation relates to a quite separate geologic event (e.g., Salobo, Ernest Henry). Mineralization is interpreted to have occurred over a wide depth range, from around 10 km (e.g., several deposits in the Cloncurry district) to close to the surface (e.g., Olympic Dam); where systems have been tilted and exposed in cross section (such as at Raúl-Condestable in Peru), they can display strongly zoned mineral parageneses. Structural and/or stratigraphic controls are pronounced, with deposits characteristically localized on fault bends and intersections, shear zones, rock contacts, or breccia bodies, or as lithology-controlled replacements.

Host rocks in the vicinity of orebodies display intense hydrothermal alteration. In the immediate vicinity of the ore, the variable pressure-temperature conditions of alteration and mineralization are reflected in a spectrum of deposits ranging from those in which the dominant Fe oxide is magnetite and alteration is characterized by minerals such as biotite, K-feldspar, and amphibole through to hematite-dominated systems in which

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the main silicate alteration phases are sericite and chlorite. Where present, Na and Na-Ca alteration tends to be developed deeper or more distal from ore, is more extensive, and commonly predates K-Fe alteration and mineralization. Carbonates are commonly abundant, particularly in association with, or postdating, Cu-bearing sulfides that tend to be paragenetically late and postdate high-temperature silicate alteration in the deeper seated deposits. Independent variation in f_{O_2} - f_{S_2} (T) conditions during mineralization produced deposits ranging from pyrite-poor examples, with complex Cu mineral associations, including chalcopyrite, bornite, and chalcocite (e.g., Salobo, Olympic Dam), to others in which pyrite and chalcopyrite are the main sulfides (e.g., Ernest Henry, Candelaria).

Fluid inclusion evidence suggests that geochemically complex brines, commonly with a carbonic component, were involved in IOCG genesis. However, the ultimate sources of water, CO₂, metals, sulfur, and salinity have yet to be well constrained, and it is possible that these components may have different origins from deposit to deposit. Brines and metals may be sourced directly from underlying magmas, indirectly by interaction of magmatic fluids with country rocks or other fluids, or independently through modification of basinal or metamorphic fluids. Ore deposition may primarily involve interaction of voluminous fluid with wall rocks and cooling. However, several studies have emphasized the role of mixing sulfur-poor, metal-rich brines with sulfur-bearing fluids at the site of ore deposition, although characterization of the causative fluids has proven problematic. Uncertainty also exists about the original tectonic settings of several major IOCG districts, and considerably more research is needed before it will be clear whether these deposits are linked by a single family of related genetic mechanisms or whether they can form in a range of fundamentally different geologic environments from fluids of different sources.

Introduction

ONE OF THE world's great ore deposits was discovered at Olympic Dam in South Australia in 1975 by Western Mining Corporation. This event was remarkable not only for the size and grade of the resource but also for the unpredicted style and geologic environment of the deposit, which were unlike those of any other major ore system known at the time (Roberts and Hudson, 1983). One of the most notable features was the geochemical signature, in which large quantities of Fe oxides are associated with Cu, Au, Ag, U, Ba, F, and light rare earth elements (LREE). Since then, the term "Fe-oxide Cu-Au (or IOCG) deposit" (Hitzman et al., 1992) has come into general use (e.g., Porter, 2000) and has become associated with one of the most challenging fields of research in contemporary economic geology. The classification has been applied to deposits displaying great diversity in their age, geochemical signature, mineralogy, host rocks, and local geologic settings. Many occur in early Precambrian terranes where the tectonic setting during ore formation is unclear or controversial. The sources of fluids, ligands, and ore metals are vigorously debated and may not be the same in all cases. Various affinities with other groups of ore deposits have been speculated upon but remain undemonstrated, and it is quite possible that examples of what have so far been called IOCG deposits will ultimately prove to be products of several fundamentally different processes and/or environments. This paper provides an overview of the time-space distribution of IOCG deposits and a descriptive geologic framework covering the larger and best studied deposits.

Historical Perspective

The emergence of the concept of an IOCG ore deposit class stemmed directly from the discovery of Olympic Dam, though mining of what might now be considered IOCG deposits goes back at least as far as the seventeenth century when Cu production was recorded from northern Sweden (Bergman et al., 2001). The Olympic Dam deposit occurs immediately below an unconformity that separates early Mesoproterozoic basement from Neoproterozoic sedimentary rocks. Its

discovery was essentially unexpected, as it resulted from a regional exploration program designed to assess the strata-bound Cu deposit potential of the cover sequence. Early observations from drill cores led to an initial misinterpretation that the hydrothermal breccias that host the orebody were sedimentary rocks (Roberts and Hudson, 1983; Reeve et al., 1990). Meyer (1988) included "Olympic Dam-type" ore deposits as one category in a larger group, also including "Kiruna-type" apatite-bearing Fe oxide ores, all proposed to be affiliated with alkalic granites, porphyries, and carbonatites. The omission of the diorite-associated and distinctly Kiruna-like apatitic Fe ores of the Andes from that classification (cf. Sillitoe, 2003) and a better appreciation of the true petrogenetic character of igneous rocks at Kiruna (Bergman et al., 2001) have exposed flaws in the basis for that grouping. However, possible roles for alkali-carbonatite magmatism in IOCG genesis, and origin of the affinities with Kiruna-type Fe ores, still remain major themes in the literature.

Two further discoveries, namely those of the La Candelaria (1987, Mesozoic Coastal Cordillera, Chile), and Ernest Henry (1991, Proterozoic eastern Mount Isa block, Australia) deposits, had immediate impacts on the perceived Cu-Au prospectivity of their regions, but it was several years before the first descriptions of these deposits were available in the public domain. In the meantime, the true breccia-hosted hydrothermal character of the Olympic Dam deposit had become apparent (Oreskes and Einaudi, 1990, 1992), and Hitzman et al. (1992) proposed that Olympic Dam, various other Fe oxide-rich hydrothermal Cu-Au deposits, together with Kiruna-type apatite-bearing Fe oxide ores, could be regarded as a related group of Proterozoic hydrothermal deposits that formed in extensional tectonic settings. This was particularly contentious in the sense that the Kiruna-type ores were long argued as originating from immiscible Fe-rich melts (see Frietsch, 1978, for a summary). Fundamentally different perspectives on the origin of apatite-Fe oxide rocks remain to this day (Nyström and Henriquez, 1994; cf. Sillitoe and Burrows, 2002), and despite spatial associations and some striking similarities in alteration mineralogy, a close genetic relationship of these deposits to IOCG deposits remains enigmatic.

Leading up to the discovery of Ernest Henry, two smaller metasedimentary rock-hosted deposits had previously been discovered in the 1980s in the eastern Mount Isa block (Cloncurry district), namely Starra, which has also been known as Selwyn, and Trough Tank, which was subsequently renamed Osborne. Both have close spatial associations with Fe oxide-rich rocks and had been interpreted previously to be syngenetic exhalative ore deposits (Davidson et al., 1989). Perhaps because of this, the concept that ore deposits with Olympic Dam affinities might exist in northern Australia received little international attention for some time, despite the possibility being promoted within the country itself (e.g., Wyborn, 1992; Williams and Blake, 1993).

Large tracts of the eastern Mount Isa block are well exposed, allowing the recognition of extensive alkali metasomatism around the batholithic granitoids. This alteration system bears a strong resemblance to that associated with apatite-bearing Fe ores in northern Sweden and the Andes and is evidently due to large-scale circulation of saline fluids (Phillips et al., 1994; Williams, 1994; De Jong and Williams, 1995; Williams et al., 1995). Iron, leached during this metasomatism, could have been redeposited in structural sites to form Fe oxide bodies and associated breccias such as those that host the Starra Au-Cu deposit (Williams, 1994). Independently, Barton and Johnson (1996) invoked the circulation of fluids with evaporite-sourced salinity in the formation of igneous-related Fe oxide-(REE-Cu-Au-U) deposits based on literature studies, geochemical considerations, and their own experience in the arc and extensional terranes of southwestern North America (Barton et al., 1988; Battles and Barton, 1995).

By the late 1990s, basic geologic data for previously undescribed, large IOCG deposits were beginning to appear, along with detailed metallogenic data for a number of deposits other than Olympic Dam (in part reviewed by Partington and Williams, 2000). The Starra and Osborne deposits were subsequently reinterpreted as epigenetic (Rotherham, 1997; Ad-shead et al., 1998) and the strongly mineralized, comparatively well-exposed, and easily accessible Cloncurry district was established as an important location for IOCG metallogenic studies (e.g., Hitzman, 2000). By this time it was also becoming apparent that the extraordinarily well endowed Carajás minerals province in Brazil contains Cu-Au deposits of IOCG affinity, including the large Salobo deposit that had been discovered in 1977 (e.g., Lindenmayer and Teixeira, 1999). This subsequently became even more significant as geochronologic data emerged to show that at least some of the Carajás deposits are Archean in age, thus considerably expanding the known temporal range of IOCG metallogenic processes (Requia et al., 2003).

Two compilations of papers (Porter, 2000, 2002) and a recent review of occurrences in the Andes (Sillitoe, 2003) have enhanced the general descriptive framework for IOCG deposits. They also reveal the tendency for an increasingly diverse range of deposits to be discussed in this context and exposed gaps in the general understanding of IOCG ore-forming processes, as discussed in later sections of this paper. Some key questions hark back to the original classification of Meyer (1988), including whether IOCG deposits are indeed part of a continuum that includes Kiruna-type Fe ores

and whether some of them, at least, may be related directly to alkali-carbonatite magmas (cf. Groves and Vielreicher, 2001).

Classification

The gaps in our genetic understanding obligate the use of a rather empirical combination of features to provide a definition of the IOCG deposit class. The key features include the following: (1) The presence of Cu with or without Au as economic metals; (2) hydrothermal vein, breccia, and/or replacement ore styles, characteristically in specific structural sites; (3) abundant magnetite and/or hematite though some districts contain cogenetic deposits in which host-rock influences appear to have suppressed the formation of Fe oxides (e.g., Williams, 2001; Knight et al., 2002); (4) iron oxides which have low Ti contents compared to those in most igneous rocks; and (5) absence of clear spatial associations with igneous intrusions, such as those which characterize porphyry and skarn ore deposits. This part of the definition excludes deposits directly associated with carbonatite intrusions (e.g., Phalabowra; cf. Groves and Vielreicher, 2001).

Other common features of IOCG deposits and their geologic environments that may not be present in all cases include: a broad time-space association with batholithic granitoids; crustal settings with exceptionally voluminous, generally pervasive alkali metasomatism; and ores enriched in a distinctive, geochemically diverse suite of minor elements, including various combinations of F, P, Co, Ni, As, Mo, Ag, Ba, LREE, and U (in contrast to Zn and Pb, which in most cases are not enriched or are depleted compared to host rocks).

Kiruna-type Fe oxide-apatite ores and Fe skarns are not IOCG deposits but may share certain characteristics with IOCG deposits, including: (1) coexistence in some metallogenic provinces where other types of ore deposit are rare; (2) common association with large-scale alkali, particularly sodic alteration systems; and (3) overlapping minor element associations (e.g., Cu, Au, P, F, REE, U). These similarities have stimulated much discussion about relationships between these various Fe-rich ore deposits, including the possibility that they represent a continuum (Meyer, 1988; Hitzman et al., 1992; Barton and Johnson, 1996, 2000; Hitzman, 2000). Consequently, the following two sections that address the space-time distribution and size-composition relationships of IOCG deposits also consider these other types of Fe-rich ore deposits.

Global Distribution in Space and Time

Hydrothermal systems with IOCG-related characteristics are numerous and widely distributed in space and time; they occur on all continents and they range in age from the present at least back into the Late Archean (Figs. 1, 2). Although Cu-Au-rich deposits of economic interest are uncommon and may be restricted to a minority of regions, each province shown in Figure 1 contains multiple systems that have deposits with abundant Fe oxide, accessory Cu, Au, Co, and/or REE, and typically extensive alkali-rich hydrothermal alteration (App. Table A1).

Within most regions, occurrences with these distinctive characteristics number in the tens to hundreds. Many regions are incompletely characterized due to combinations of

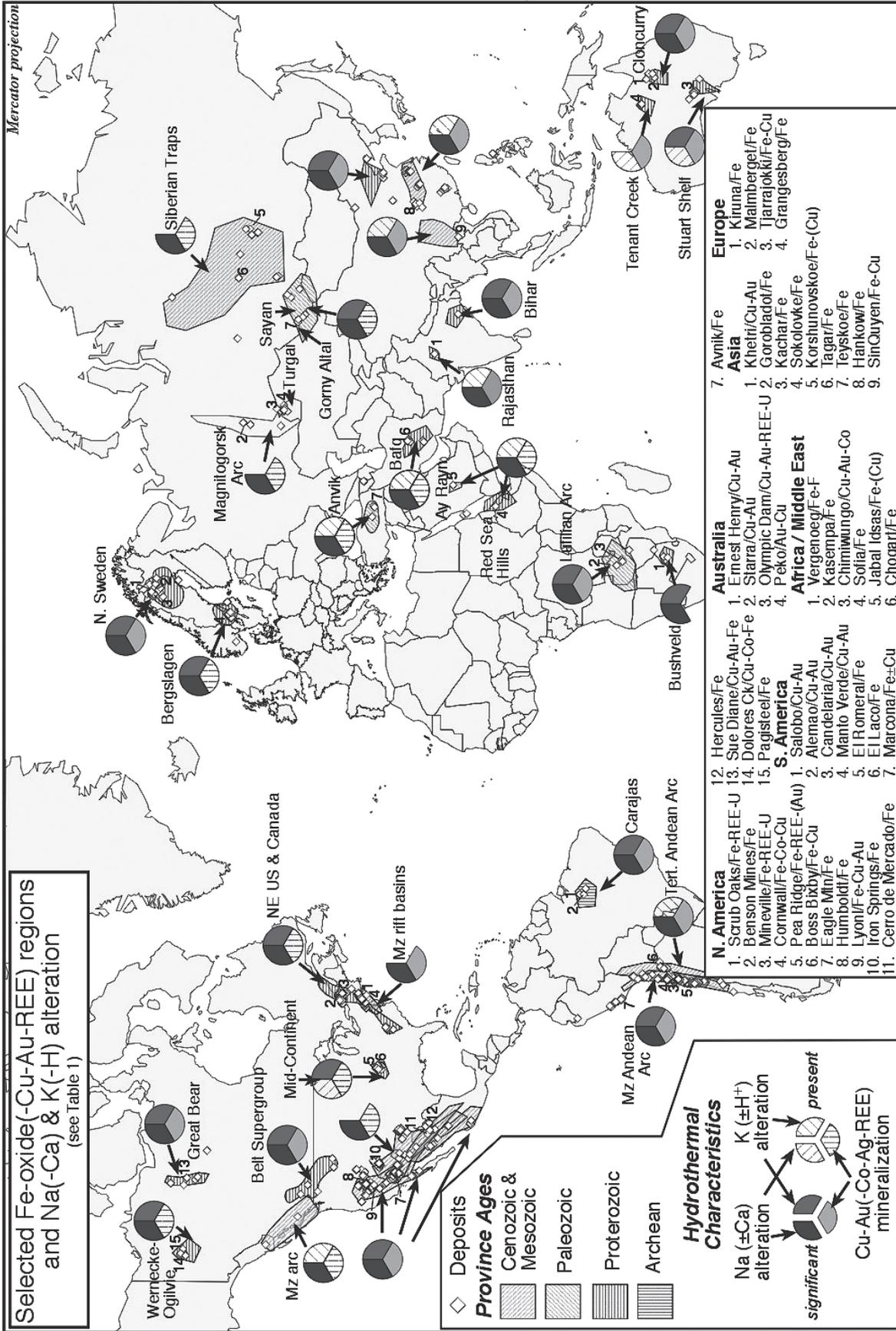


FIG. 1. Location of principal IOCG and other hydrothermal Fe oxide provinces with selected deposits, some of which are discussed in the text. Provinces are indicated by the lined fills and differentiated by age (cf. Fig. 2). For each province, the three-part circular symbols indicate the presence and known significance of metals and major types of hydrothermal alteration. Information and key references for each province are summarized in Appendix Table A1.

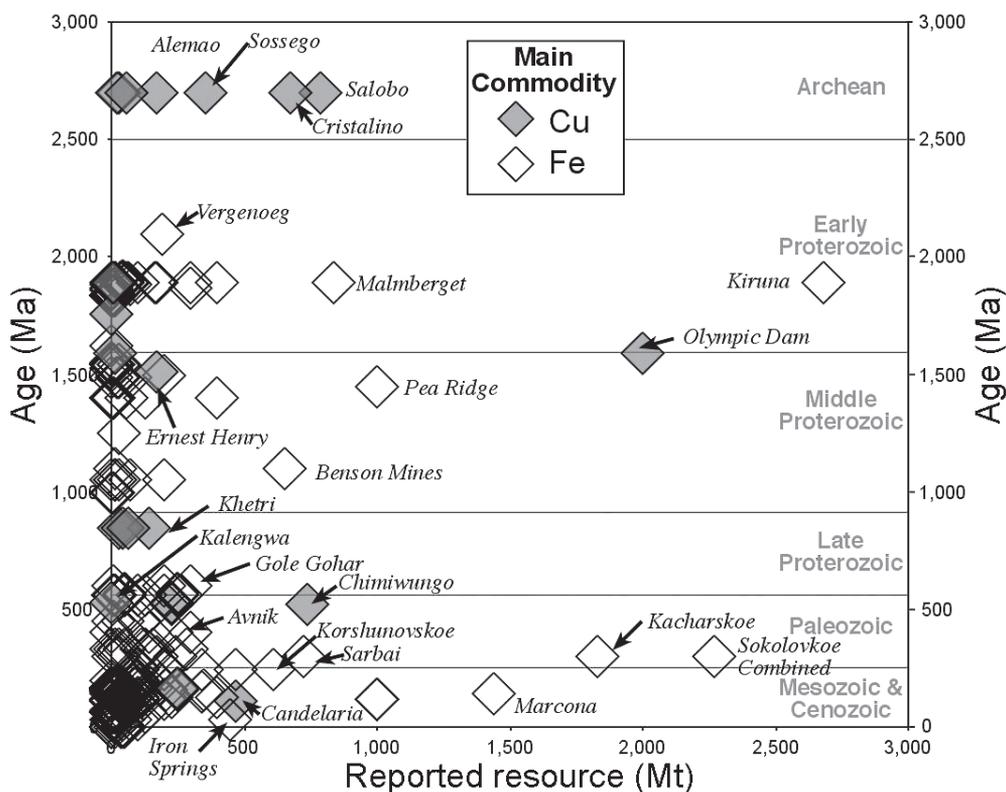


FIG. 2. Age distribution of IOCG deposits and Fe oxide deposits (mainly magnetite-apatite and Fe skarn; cf. Barton and Johnson, 1996), divided on the basis of the main commodity recovered (Cu or Fe). Ages for provinces are summarized in Table A1; deposit data are derived from publicly available data (D.A. Johnson and M.D. Barton, unpub. compilation).

postmineral cover, complex geologic histories that can obscure the features of the IOCG systems, and the relatively recent interest in these systems as a defined group. Nevertheless, they can be easily recognized because of their distinctive bulk physical and chemical characteristics that survive even intense overprinting by younger events. Associated igneous rocks and tectonic settings (Table A1) are quite varied. Igneous rocks, which are almost invariably present, range from felsic, variably alkalic, granitoids to intermediate arc-related rocks to mafic intrusions. Large Cu and Au deposits are associated with the intermediate to felsic compositions.

Neither Fe- nor Cu-dominated deposits show an obvious time dependence apart from the perhaps singular abundance of Cu-rich deposits in the Late Archean Carajás province (Fig. 2). Ages in many areas are poorly constrained because of difficulties in dating multiply deformed and altered rocks. Local maxima in the temporal distribution represent individual or spatially related provinces. Namely, Carajás for the Late Archean, northern Laurentia (northern Sweden, Great Bear) for the Early Proterozoic, most Australian and the United States midcontinent systems for the Middle Proterozoic, Pan-African events for the Late Proterozoic, the Altaiides for the middle to late Paleozoic, and the American Cordillera for the late Mesozoic to Cenozoic. The Middle Proterozoic suite recognized long ago by Meyer (1988) is but one of these episodes and gave a misleading impression that IOCG systems were distinctively Proterozoic in age, a concept now known to be incorrect (Fig. 2).

Age relationships within individual provinces are complex, as discussed below. Commonly, but not universally, Cu-Au mineralization occurred relatively late in the prolonged development of well-mineralized terranes (e.g., Cloncurry; Williams and Skirrow, 2000). In other areas, Cu was deposited relatively early, or repeatedly, within particular magmatic episodes (e.g., Humboldt system Nevada; Dilles et al., 2000; Punte del Cobre district, Chile; Marschik and Fontbote, 2001a).

Late Archean and Early Proterozoic districts are spread across most continents, with a temporal concentration around 1900 Ma (Baltic Shield, Great Bear, Tennant Creek). The Late Archean Carajás province is exceptionally Cu-Au rich. These regions contain numerous systems that share alteration and metal associations with younger examples. The Norrbotten, Sweden, ore province in the northern Baltic Shield is notable for its regional association of IOCG deposits and apatite-bearing Fe oxide bodies, including the archetypal example of the latter at Kiruna, and regional Na-dominated alkali alteration (Frietsch et al., 1997; Bergman et al., 2001). The province also contains a large Fe oxide-bearing Cu-Au deposit at Aitik that has debatable IOCG versus porphyry affinities (Wanhainen et al., 2003). Australia and North America host the best known Middle Proterozoic systems—an association that may reflect proximity of these continents during that time (e.g., Thorkelson et al., 2001). Occurrences in the Wernecke Mountains, Yukon Territory, Canada, contrast with similar-aged examples in Australia in that there is

currently no evidence for synchronous magmatic activity. Deposits and occurrences in southern Australia (Stuart Shelf, South Australia, and Curnamona craton, New South Wales) are of approximately the same age (1500–1600 Ma) as deposits of the Cloncurry region of the Mount Isa inlier, Queensland. Broadly similar ages (ca. 1500 Ma) typify the granite-rhyolite terrane of the United States midcontinent and metamorphosed equivalents along the Eastern seaboard (Table A1). Sodic alteration is present in all these Middle Proterozoic areas. K silicate alteration occurs widely and can be recognized even in granulite facies terranes in the eastern United States where it is represented by microcline-magnetite gneisses.

Possible latest Proterozoic and earliest Paleozoic IOCG systems occur predominantly in Africa and adjacent continents, notably in southern Africa in the Lufilian arc and within and on the margins of the Arabian Shield (Figs. 1, 2). Although precise dating has only recently begun, mineralization in these regions may have taken place intermittently over several hundred million years and may overlap with Gondwanan occurrences in southwestern Asia (Iran and western India; see Table A1). Moderate- to high-temperature sodic (-calcic) alteration is widespread and K silicate assemblages are reported from most areas. Possible mid to late Paleozoic IOCG provinces occur principally along the Altai-related belts from central Asia through Kazakhstan, the Ural Mountains, and into Turkey (Fig. 1; Zonenshain et al., 1990). These include the large (some >1 billion tons), Cu-bearing (up to 0.6 wt %), diorite-related volcanic-hosted and skarn deposits of the Turgai and Magnitogorsk provinces in Kazakhstan and Russia, respectively (Smirnov, 1977; Herrington et al., 2002). All these districts contain abundant sodic-calcic alteration; widespread potassic alteration occurs with more felsic igneous suites in Central Asia.

Mesozoic-Cenozoic occurrences occur along the margins of modern continents in eastern China, the American Cordillera, and the eastern United States. They include the Fe (-Cu-Co-Au) skarns and veins of southeastern Pennsylvania, which are distinctive in that they are related primarily to basaltic intrusions. Analogous, but larger breccia dominated, hydrothermal Fe oxide systems with extensive sodic-calcic and/or skarn alteration, but lacking economic Cu, occur with the Permo-Triassic Siberian traps (e.g., Korshunovsk, Vetka). The Chilean and Peruvian coastal belts contain many magnetite-apatite deposits such as Marcona, Peru, and El Romeral, Chile, as well as a significant number of Cu ± Au deposits, including Candelaria and Manto Verde, Chile, and Raul-Condastable, Peru. This latter group and their Chinese and North American counterparts are associated with arc to back-arc environments dominated by intermediate magmatism (Table A1). The “detachment-type” massive hematite-hosted Cu-Au occurrences of the southern Basin and Range province are a variant that lacks any clear association with magmatism (Wilkins et al., 1986; Ilchik and Barton, 1997). Cenozoic systems of possible IOCG affinity are best known for their Fe production or Fe potential (e.g., Cerro de Mercado, Durango, Mexico; Iron Springs, Utah; El Laco, Chile). Only a few of these deposits have produced Cu or Au, notably minor historical production from detachment-related hematite Cu-Au deposits of the southwestern United States

(Wilkins et al., 1986), although high Andean Cu-Au prospects of likely IOCG affinity such as Arizaro, Argentina (Dow and Hitzman, 2002), have attracted considerable attention.

Metal Contents and Sizes of IOCG Systems

The largest apatite-bearing Fe ore and IOCG systems contain more than 1 billion tons (Gt) of Fe ore or Cu-Au ore and commonly have significant quantities of accessory elements including P, REE, Co, Ag and U (Table A1, Fig. 3, App. Fig. A1). Relatively few deposits have a full suite of chemical data; Fe deposits generally have reported Fe, P, S, and Ti, rarely Cu, and in almost no cases Au or other trace elements. Conversely, in Cu deposits, only Cu and Au are widely reported, other trace element compositions are sparse, and data on Fe contents are rare. Mineralogy and rock descriptions consequently provide the main basis for recognizing similarities among these deposits, although there are sufficient Cu data from Fe deposits to suggest these could all represent a continuum (Fig. 3).

Iron ore deposits average between 30 and 70 wt percent Fe, corresponding to 50 to nearly 100 wt percent hematite or magnetite (Fig. A1). In the few Cu-rich deposits for which Fe data are available, concentrations range from 15 to 25 wt percent Fe in the Cu ores. The difference in values reflects the metals of interest, e.g., in the Candelaria Cu(-Au) deposit large volumes of rock contain >30 percent Fe but do not coincide with the Cu orebodies. The amount of magnetite contained in the Candelaria system (i.e., not restricted to Cu-Au ore) likely exceeds the amount in the largest Fe deposits of the Chilean Fe belt, and the contained Fe in the Cu ore at Olympic Dam approaches that of the largest Fe systems known. Among the other ferrous metals, available data indicate that Ti is scarce (<1 wt % TiO₂; Fe/Ti > crustal Fe/Ti) in both Cu and Fe ore types; Co and V can be enriched relative to Fe, but Ni, Mn, and Cr generally are depleted relative to Fe (Kisvarsanyi and Proctor, 1967; Frietsch, 1970; Hauck, 1990).

Copper grades from >60 Cu-Au deposits mostly fall between 0.5 and 4 wt percent, averaging about 1 wt percent; they may form an approximate continuum with Cu contents in Fe-dominant deposits, where reported concentrations are not uncommonly on the order of 0.1 wt percent (Fig. 3). Most Fe deposits lack Cu assays, even though chalcopyrite is commonly described as an accessory mineral. Typically, Cu and Au contents represent hypogene grades, although weathering can remove Cu, leaving residual Au-only resources, as in the upper part of the Igarape Bahia-Alemão system in the Carajás district.

Copper/Fe and Cu/Au vary significantly among deposits and show major breaks, principally as a function of the major commodity produced. Many Fe deposits have Cu/Fe ~1/500 to 1/50, similar to, or greater than, the crustal ratio for these elements, whereas the Cu-Au deposits have Cu/Fe ~1/15 (D.A. Johnson and M.D. Barton, unpub. compilation). Copper/Au ratios are available only for Cu-rich systems, and the ratios also are within a factor of 5 of crustal values and overlap with the ranges found in Au-bearing porphyry Cu systems (D.A. Johnson and M.D. Barton, unpub. compilation). Interestingly, in most deposits and districts, Cu/Au varies by less than a factor of 3 unless affected by supergene processes.

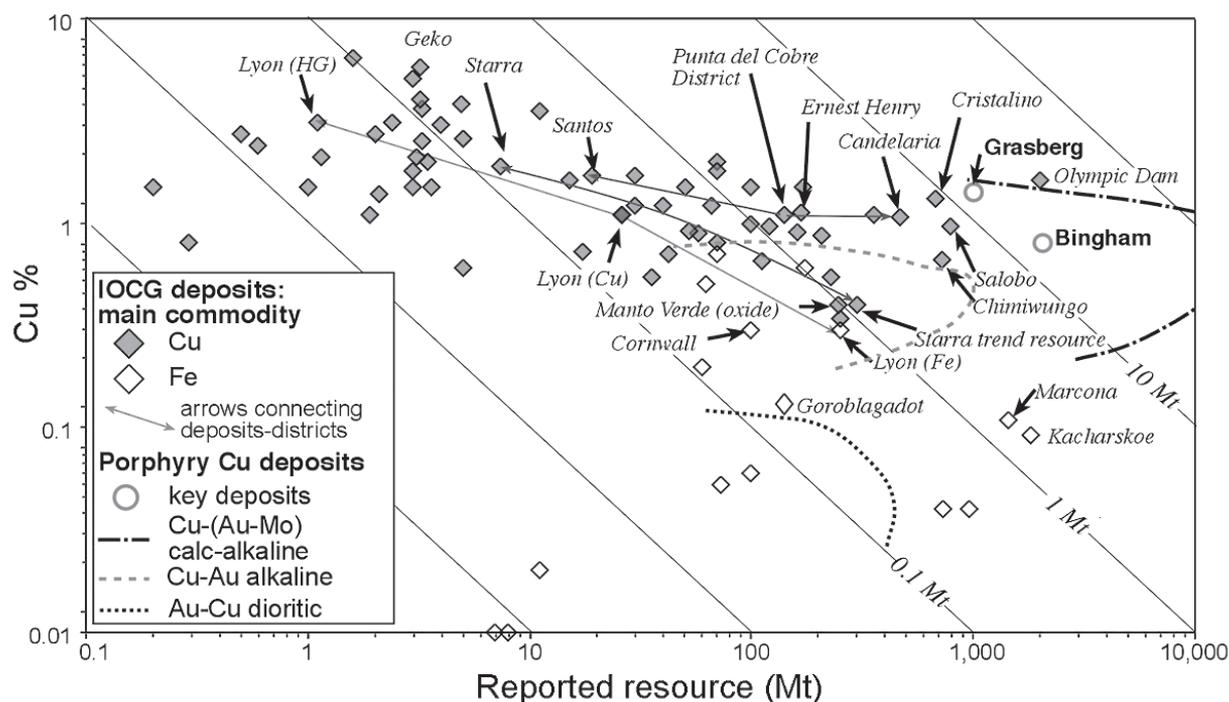


FIG. 3. Grade-tonnage data showing Cu grades for Fe oxide (mainly magnetite-apatite and skarn) and Fe oxide-Cu-rich hydrothermal systems (D.A. Johnson and M.D. Barton, unpub. compilation). Note the arrows connecting various resource categories (deposits and ore types) within the same districts. These data are skewed by the principal commodity produced and by the scarcity of representative Cu data for Fe-producing deposits. Also shown are fields for several types of porphyry Cu-Au-rich systems (modified from Seedorff et al., 2005).

Some of the Tennant Creek deposits and Starra in the Cloncurry district stand out in their unusually high Au contents. Silver is widely mentioned but without grade information—Ag/Au is 10 ± 5 (by wt) for 11 districts where numbers have been published. Uranium minerals are commonly reported, but quantitative U data are virtually nonexistent.

Few data are available for other base metals; Mo, as molybdenite, is commonly reported but published assays are lacking. Zinc and Pb are absent in most systems, although a few, such as Candelaria and Monakoff, Queensland, have significant Zn contents that may exceed 0.1 wt percent in large portions of the resource (Ryan et al., 1995; Davidson et al., 2002). Rare earth element concentrations vary considerably. Reported total RE₂O₃ concentrations average about 0.5 wt percent, with values locally reaching about 10 percent in small bodies, whereas estimated bulk contents in apatite-bearing Fe ores are about 0.1 percent (Johnson and Barton, in press).

Among other elements, S and P in Fe deposits range from a few wt percent to as low as a few hundredths of a percent. Reported values in Fe ores likely underestimate the bulk content, given the deleterious effect of these elements. For example, apatite-rich veins are commonly reported in many systems but are not included in the material budgets. Skarn-type Fe ores of possible IOCG affinity (Table A1) typically have significantly lower P contents (<0.2 wt % P₂O₅) than ores in volcanic, plutonic, or clastic hosts (e.g., data in United Nations, 1970). Bulk P₂O₅ values have not been published for Cu-rich systems, although the lack of described phosphates in most deposits indicates that bulk contents are lower than in

most Fe ores. Reported sulfur contents of Fe ores average 0.7 wt percent, whereas modal data and Cu grades indicate that Cu-Au-rich deposits likely average 4 to 10 wt percent total sulfide (i.e., about 2–5 wt % S).

Most IOCG deposits have low total sulfide contents and high proportions of Cu-bearing sulfides compared to pyrite. As such they are generally less susceptible to supergene enrichment than other types of sulfide Cu deposits (e.g., porphyry deposits and volcanic-hosted massive sulfide deposits). Nonetheless, significant supergene effects may be present, as at Ernest Henry in Australia, where a well-developed weathering profile was present prior to mining (Ryan, 1998). Several IOCG resources contain a large proportion of oxidized (acid-leachable) ore as at Manto Verde in Chile, described in a later section.

District and Deposit Characteristics

General

The remainder of this paper is devoted to the few districts that have been significant economically for their Cu-Au content, as opposed to Fe endowment. IOCG deposits are unusually diverse in comparison to deposits within most other recognized classes of hydrothermal ores. Globally, they vary in age, size, Cu/Au, host rocks, alteration associations, conditions of mineralization, geochemical signature, and physical properties. Even within single districts, broadly coeval deposits can be quite different from one another in these respects (e.g., Skirrow, 2000; Williams and Skirrow, 2000). The

reader requiring more information about deposits not dealt with here is directed to many relevant papers in the compilations by Porter (2000, 2002). The deposits described here by region illustrate some of this variability and also include some of the largest (Fig. 2) and most intensively studied examples. The deposits are (1) Salobo, the best documented example from the Carajás district in Brazil and one in that region now interpreted to be Archean in age (Requia et al., 2003); (2) Ernest Henry, a large and currently mined Proterozoic deposit in the Cloncurry district of Australia where the predominant Fe oxide is magnetite and the broad-scale mineralogical-geochemical zoning is well documented; (3) Olympic Dam, a Proterozoic deposit in which the predominant Fe oxide is hematite, located in the Gawler craton of South Australia and containing the world's largest known IOCG resource; (4) Candelaria-Punta del Cobre, located in the Mesozoic Coastal belt of Chile, in which the dominant Fe oxide is magnetite and encompasses the Candelaria mine which has the largest resource of any Phanerozoic IOCG deposit; (5) Manto Verde, another large deposit from the Chile Coastal belt that has both hematite- and magnetite-dominant zones; and (6) Raul-Condostable, a relatively small deposit from the Mesozoic Coastal belt in Peru, showing a similar mineralogy and alteration pattern as Candelaria and included here because it has an unusually large paleodepth section preserved.

The selected deposits reflect the wide distribution of IOCG systems in geologic time and emphasize uncertainty and debate about the original tectonic context of IOCG systems in general, particularly in respect to the Precambrian examples. Salobo and several other large IOCG deposits appear to have developed in association with granitoid magmatism during

the Late Archean evolution of the Carajás province within the Amazon craton (Requia et al., 2003). As with all geologic associations of this age, there is inherent uncertainty about the extent to which the settings of these deposits can be compared to those of modern-day plate tectonic styles. Proterozoic systems in Australia, including the eastern Gawler craton (hosting Olympic Dam) and Mount Isa block in the North Australian craton (hosting Ernest Henry), have been widely interpreted as products of intracratonic processes (following Etheridge et al., 1987). However, it has been argued recently that the critical rock associations in these cratons evolved in distal subduction-related settings (Betts et al., 2002) and further research is needed to resolve this matter. It should also be noted that IOCG metallogeny in the eastern Gawler craton and Mount Isa block appears to have been related to two separate orogenic events in the period 1.6 to 1.5 Ga (see below), and that the IOCG endowment of the Australian continent cannot be a product of unique circumstances created in a single setting. In contrast, the younger deposits of the Andean coastal belt in Chile and Peru are clearly related to magmatism in transient extensional settings in an archetypal continental margin arc (e.g., Sillitoe, 2003).

Carajás district, Brazil

General: The Carajás district in Brazil might be judged the world's outstanding IOCG locality based on the presence of several Cu-Au resources larger than 100 Mt (Table A1; Figs. 3, 4). Of these, the as yet unmined Salobo deposit has one of the highest Cu contents among IOCG deposits worldwide and, to date, Salobo is the most intensively studied due in part to its relatively early discovery in 1977 to 1978 (e.g., Souza and Vieira, 2000).

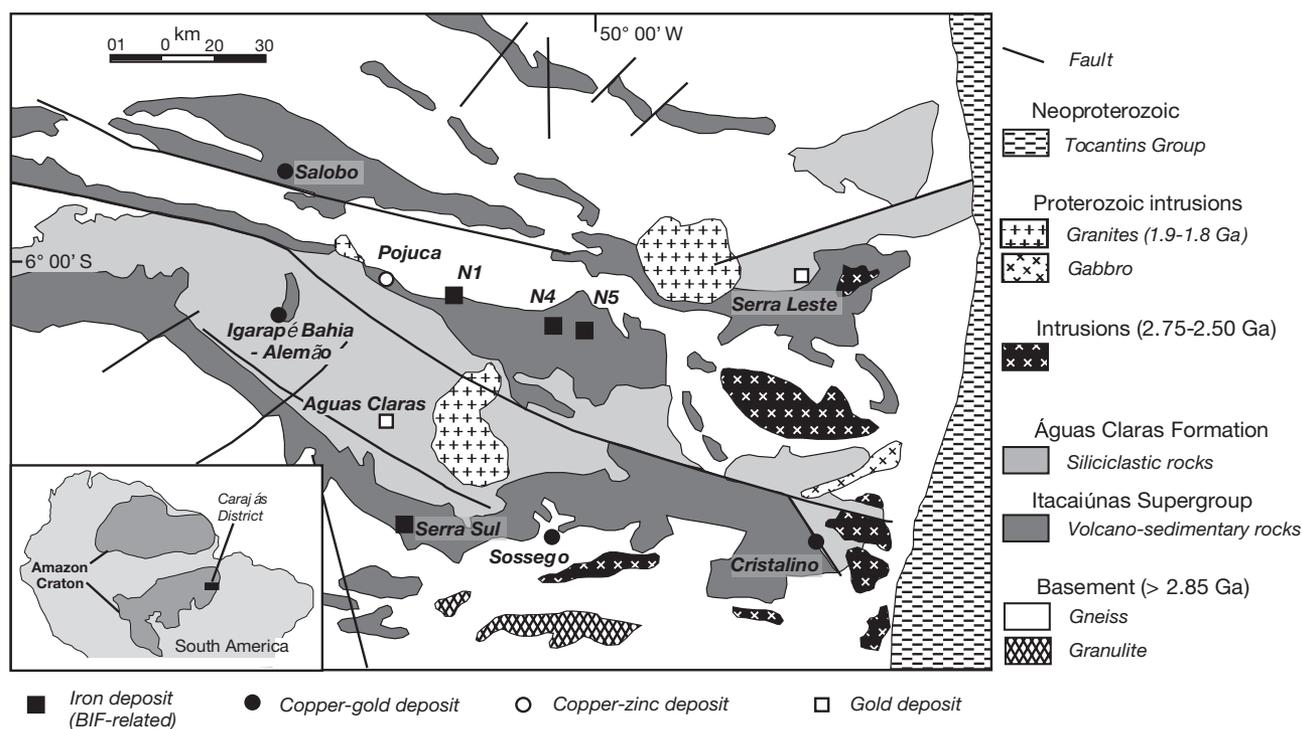


FIG. 4. Geology and major ore deposits of the Carajás mineral district, as adapted by Tazava and de Oliveira (2000) from Docego (1988) and Barros et al. (1997).

The region lies in Amazonia in the southeastern part of the Amazon craton. Archean basement (>2.85 Ga; Machado et al., 1991), older volcano-sedimentary ($> ca. 2.7$ Ga Itacaiúnas Supergroup) and younger siliciclastic (2.7–2.6 Ga Águas Claras Formation) cover sequences are variously intruded by ca. 2.6 to 2.7 Ga granitoids (Plaquê Suite and Estrela Complex), ca. 2.65 Ga gabbro and diabase sills and dikes, ca. 2.57 Ga granite (Old Salobo Granite) and 1.8 to 1.9 Ga anorogenic granites (Young Salobo Granite and other intrusions); see sources of geochronological data summarized by Requia et al. (2003). Basement and Itacaiúnas Supergroup rocks are highly deformed and metamorphosed at grades ranging up to granulite facies.

The main IOCG deposits of Carajás are hosted by various units of the Itacaiúnas Supergroup, which consists of bimodal metabasalt-metarhyolite associations, intermediate to felsic pyroclastic rocks, metasiliclastic rocks, carbonates, and Fe formation that are inferred to have been deposited on older continental crust in an intracratonic rift or rifted margin (e.g., Docegeo, 1988; Olszewski et al., 1989).

Salobo—an Archean, very low sulfur, magnetite-dominated system: The Salobo deposit occurs in a partly shear-bounded 300- to 600-m-thick sequence of metagraywacke, quartzite, iron formation, and amphibolite known as the Salobo Group. This is intercalated with basement gneisses and cut by intrusions of both Old (2573 ± 2 Ma; Machado et al., 1991) and Young (ca 1880 Ma) Salobo Granite (Fig. 5). In the vicinity of the deposit, the Salobo Group includes a large proportion of K- and Fe-rich rocks containing K-feldspar, biotite, cummingtonite-grunerite series amphiboles, almandine garnet, and magnetite (Lindenmayer and Teixeira, 1999; Souza and

Vieira, 2000; Requia et al., 2003) that are interpreted to be altered rocks (Requia and Fontboté, 2000). The principal hosts of the orebody are exceptionally Fe rich rocks composed of magnetite, fayalite, Fe-rich amphiboles, almandine, and biotite that contain disseminated bornite, chalcocite, and chalcopyrite. The ore has a complex B-F-Co-As-Mo-Ag-LREE-U minor element association and some of it is graphitic (Lindenmayer and Teixeira, 1999; Requia and Fontboté, 2000; Souza and Vieira, 2000; Requia et al., 2003).

Re-Os dating of molybdenite associated with Salobo Cu minerals produced two slightly different age groupings with weighted means of 2576 ± 8 and 2562 ± 8 Ma, interpreted to reflect primary mineralization and a later disturbance associated with shearing, respectively (Requia et al., 2003). Further support for an Archean age of mineralization was obtained using a Pb-Pb step leaching of Cu sulfides that produced an imprecise isochron of 2579 ± 71 Ma (Requia et al., 2003). Primary mineralization therefore appears to have occurred at about the time of emplacement of the Old Salobo Granite.

Early studies of the Salobo deposit led to the suggestion that it formed by exhalative processes related to the deposition of the Salobo Group (Lindenmayer, 1990). However, reinterpretation of the Fe-rich hosts as products of replacement, the distinctive minor element geochemical signature, alteration styles, and new geochronological data all point to an epigenetic hydrothermal origin and classification as an IOCG deposit (Lindenmayer and Teixeira, 1999; Requia et al., 2003), as defined in the introduction to this paper. The limited amount of published data for other large Carajás Cu-Au resources in the district indicates that these vary in their style and mineralogy. The Alemão deposit, for instance, has a

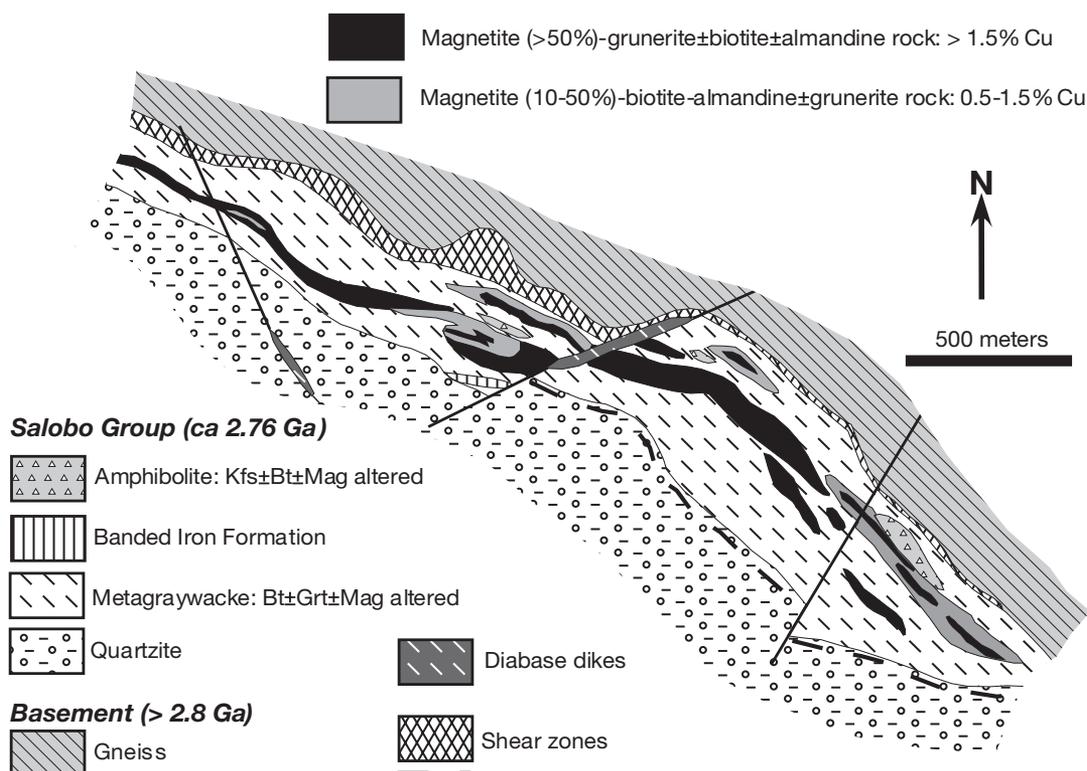


FIG. 5. Geology of the Salobo deposit, Carajás district, at level 250 (adapted from Requia et al., 2003).

distinctive minor element signature that is similar to Salobo, although the ore is quite different in character, being composed of brecciated and altered metavolcanic rocks with magnetite, carbonate, pyrite, and chalcopyrite (Ronzê et al., 2000). In these respects, Alemão ore resembles that at the Ernest Henry IOCG deposit in Australia (see below). The Sossego deposit, which was recently brought into production as the first Cu mine in the Carajás district, consists of bodies of veinlet, vein, and breccia-style chalcopyrite-magnetite

within extensively Na- and Ca-metasomatized igneous rocks (Marschik and Leveille, 2001).

Cloncurry district, Australia

General: Geologically, this district occupies the Eastern fold belt of the largely Paleoproterozoic Mount Isa inlier in northwestern Queensland, along with its extensions beneath shallow Palaeozoic and Mesozoic cover (Fig. 6; Williams and Skirrow, 2000; Williams and Pollard, 2003). IOCG deposits

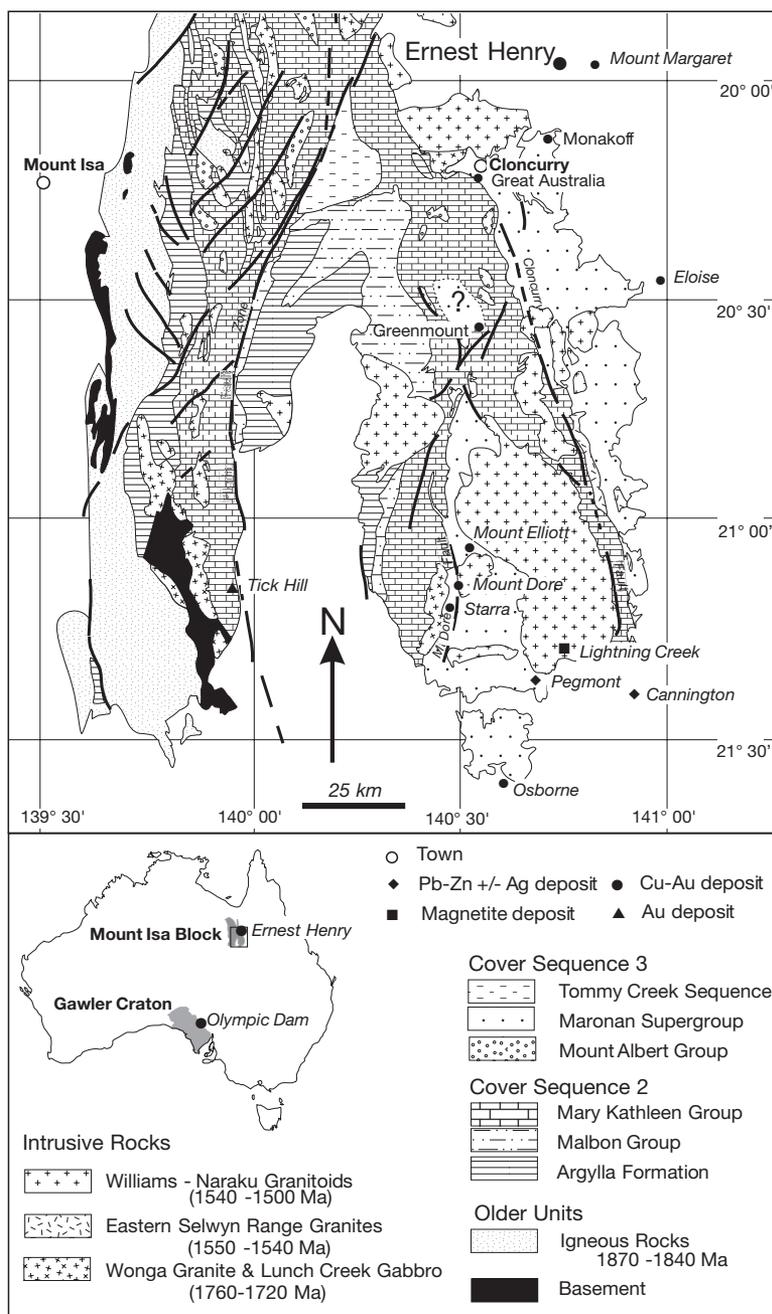


FIG. 6. Location of the Ernest Henry deposit and geology of the adjacent parts of the eastern Mount Isa block in northern Australia (geology as adapted by Williams (1998) from maps published by the Australian Bureau of Mineral Resources and other sources). Metavolcanic rocks in an isolated outcrop west of Ernest Henry are similar in age to the Wonga Granites, suggesting that the host rocks of the deposit belong to a supracrustal unit that may represent a younger component of cover sequence 2 in the main Proterozoic inlier (Page and Sun, 1998).

occur in rocks belonging to two major supracrustal sequences known as cover sequences 2 and 3 (Blake, 1987; Fig. 6). Cover sequence 2 was extensively intruded by granitoids at 1760 to 1720 Ma in the western part of the district, and diorites were emplaced near Ernest Henry at around 1660 Ma (Pollard and McNaughton, 1997; Page and Sun, 1998). Both cover sequences were affected by two orogenic events with extensive deformation and metamorphism up to upper amphibolite facies at ~1590 Ma (Diamantina orogeny) and from 1550 to 1500 Ma (Isan orogeny; MacCready et al., 1998). The only known intrusions related to the earlier event are anatectic pegmatites, whereas the differentiated, predominantly potassic and magnetite-bearing Williams-Naraku batholiths were emplaced in the eastern part of the district during the latter event (Pollard and McNaughton, 1997; Page and Sun, 1998; Giles and Nutman, 2002). Iron oxide Cu-Au deposits seem to have been formed during both events, but only the Osborne deposit thus far provides evidence that it is related to the earlier Diamantina event as implied by the correspondence of U-Pb (titanite) ages from pegmatites and Re-Os (molybdenite) ages from the ore (Gauthier et al., 2001). Metamorphic ^{40}Ar - ^{39}Ar ages are in the 1590 to 1550 Ma range (Perkins and Wyborn, 1998). All other published radiometric ages in the district are consistent with mineralization during the latter part of the Isan orogeny, synchronous with the Williams-Naraku batholiths, including two U-Pb (titanite) ages from the Ernest Henry alteration system (Mark et al., in press), a U-Pb (SHRIMP-rutile) age from Ernest Henry ore (Gunton, 1999), and more than 20 ^{40}Ar - ^{39}Ar (mica/amphibole) ages from Ernest Henry and other deposits (Pollard and Perkins, 1997; Twyerould, 1997; Perkins and Wyborn, 1998).

The Ernest Henry mine exploits what is by far the largest of the several economic IOCG deposits in the Cloncurry district. Like Olympic Dam, it is a breccia-hosted deposit. Other economic and subeconomic deposits near Cloncurry have a variety of styles, including (1) quartz-rich lodes at Osborne and (part of) Eloise (Adshead et al., 1998; Baker, 1998); (2) replaced hornblende-biotite-altered rocks at Eloise (Baker, 1998); (3) replaced magnetite ironstones at the Starra mine (Rotherham, 1997); (4) skarn at Mount Elliott (Wang and Williams, 2001); (5) carbonate-rich veins at Great Australia (Cannell and Davidson, 1998); (6) veins and disseminations in selectively mineralized carbonaceous rocks at Greenmount (Krcmarov and Stewart, 1998); and (7) magnetite-barite-fluorite-manganosiderite lodes at Monakoff (Davidson et al., 2002).

Ernest Henry—a Proterozoic magnetite-dominated deposit: The Ernest Henry deposit was originally concealed by cover rocks ~12 km from the nearest Proterozoic basement outcrop. The deposit was discovered beneath 35 to 60 m of Mesozoic to Recent sedimentary material by drilling coincident magnetic and electrical anomalies by Western Mining Corporation in 1991 (Webb and Rowston, 1995). The published premining resource was 166 Mt at 1.1 wt percent Cu and 0.54 g/t Au and was open at depth (Ryan, 1998). An area of approximately 15 km² around the deposit was systematically drilled with cored holes on 100- to 200-m centers in a sterilization program prior to mine development. Along with the associated geochemical database, this has provided a

unique opportunity to study the mineralogical and geochemical characteristics of the alteration system around this major IOCG deposit (Fig. 7).

Strongly altered basic to felsic metavolcanic rocks (mostly meta-andesite) predominate in the drilled area and probably correlate with outcrops of ca. 1745 Ma metavolcanic rocks farther west (Page and Sun, 1998; Mark et al., 2000). Subordinate metasedimentary rocks include diopside- and scapolite-bearing calc-silicate and metasiliclastic rocks with localized cordierite, andalusite, and garnet. The regional structural and geophysical grain trends steeply north but bends to the northeast in the vicinity of the deposit. The main shear zone foliations and faults at the mine dip moderately southeast, and the breccia-hosted orebody plunges downdip to the southeast within these fabrics (Webb and Rowston, 1995; Twyerould, 1997). Dioritic intrusions, emplaced at ca. 1660 Ma, occur both north and south of the deposit and generally lack tectonic fabrics despite the fact that they predate the main regional deformation events (Pollard and McNaughton, 1997; Twyerould, 1997; Ryan, 1998; Mark et al., 2000). Detailed logging of drill holes in mine section 39080N has allowed the immediate host rocks of the orebody to be differentiated on the basis of the presence and/or absence and grain size of plagioclase phenocrysts (Fig. 8). Along with subordinate bodies of metasedimentary rocks, these petrographic variations suggest a complex pattern of discontinuous and lenticular rock units, which mine geologists have interpreted to be influenced by faults at low angles to the tectonic foliation.

The majority of the rocks in the drilled area are strongly altered as are those in the nearest surface exposures some 12 km away from the deposit. The alteration system evolved through a series of distinct stages, causing a series of overprints that produced complex mineralogical and geochemical distribution patterns (Twyerould, 1997; Mark et al., 2000). A classification of alteration and veining products simplified from Mark et al. (2000) is adopted here for brevity, in their order of development (1) sodic and/or sodic-calcic, (2) premineralization, spatially associated with the deposit, (3) mineralization, and (4) postmineralization.

The earliest alteration produced sodic and sodic-calcic assemblages broadly similar to those that characterize regionally extensive alteration systems in the exposed Proterozoic basement farther south (cf. Williams, 1994; De Jong and Williams, 1995; Oliver et al., 2004). This apparently affected rocks throughout the area although later events, especially in the vicinity of the deposit, were too strong to have preserved the sodic and calcic minerals. This stage typically involved pervasive, texturally preservative and/or destructive replacement of the host rock by albite-rich plagioclase associated with veins and breccia zones containing infill with various combinations of diopside, actinolite, magnetite, and albite. Rocks that preserve the effects of this stage commonly have Na₂O >5 wt percent and very low K/Na ratios (Mark et al., in press).

Premineralization veins and alteration spatially associated with the deposit overprinted the sodic-altered rocks with a range of K-, Fe-, and Mn-bearing minerals, including biotite (in part manganiferous), magnetite, almandine-spessartine garnet, and K-feldspar. Hairline cracks containing fine-grained

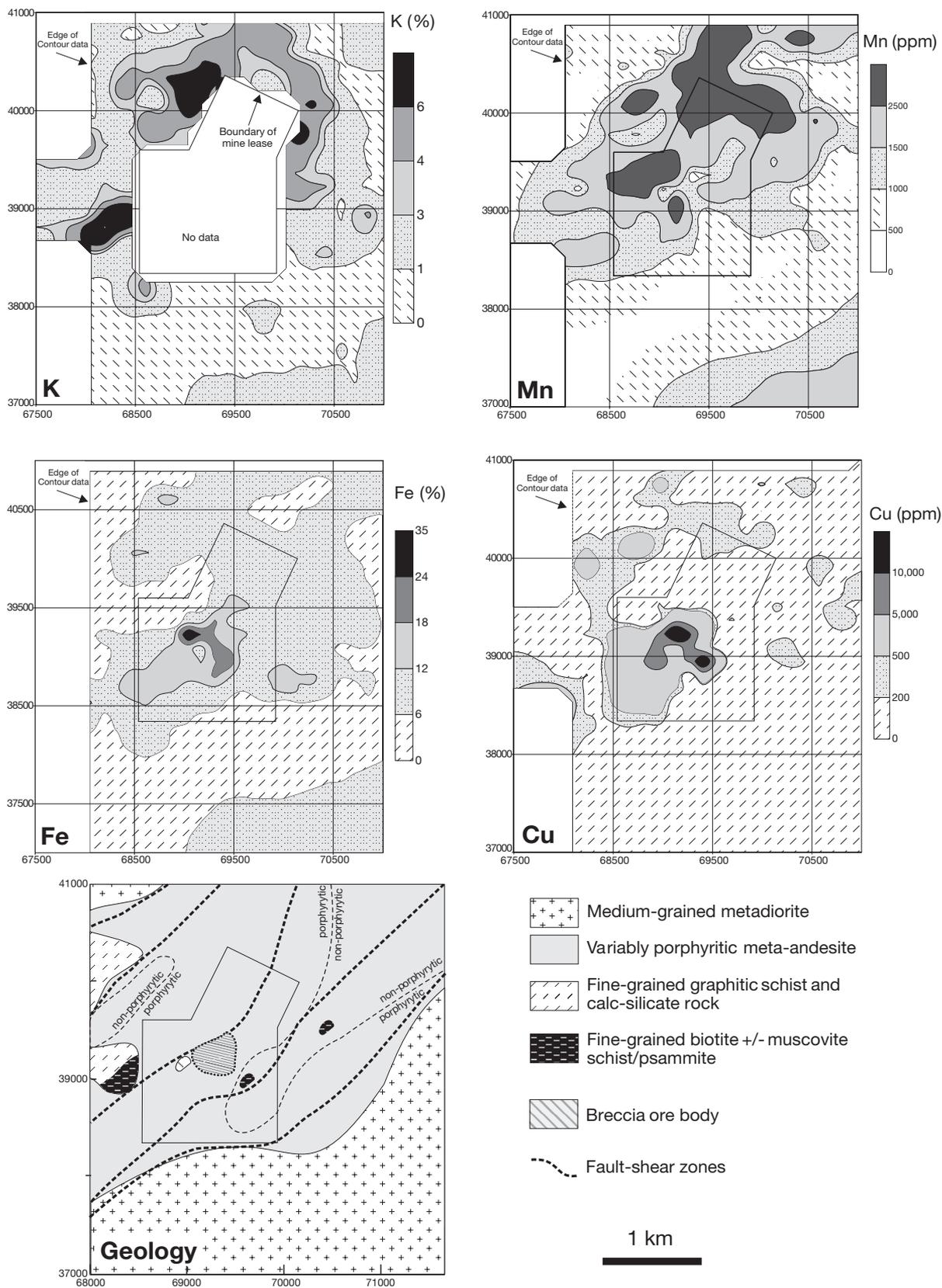


FIG. 7. Geology (logged in drill holes) and selected element distributions (from term-lease sterilization drilling geochemical database) at relative elevation 1,947 m (cf. Figs. 8, 9) in the vicinity of the Ernest Henry mine (adapted from Mark et al., in press). Grid in meters.

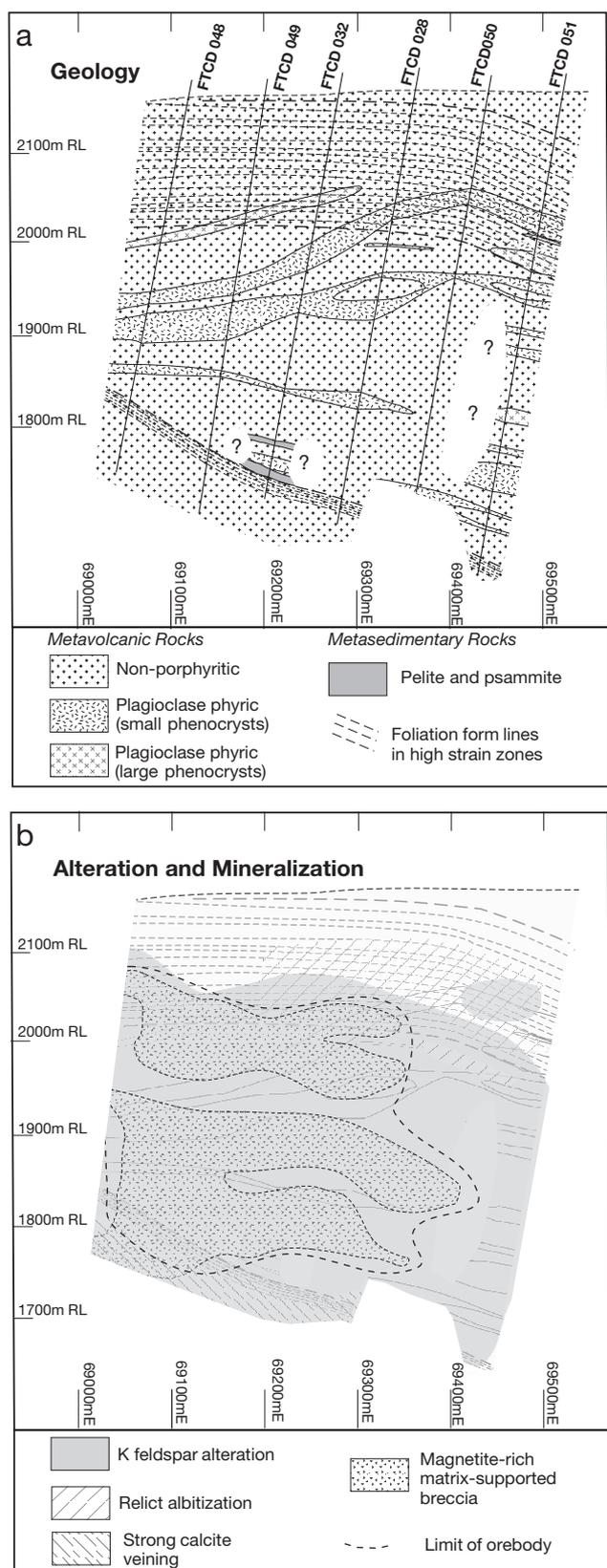


FIG. 8. (a). Lithological variation and tectonic fabric form lines in Ernest Henry mine section 39080N (Fig. 9) as logged from cored drill holes. (b). Distribution of Cu concentration and key alteration minerals in Ernest Henry mine section 39080N (adapted from Mark et al., in prep).

magnetite and biotite, interpreted to belong to this stage, are recorded up to several kilometers away from the deposit. Near-pervasive effects in the mine area extend 1 to 2 km from the orebody (Fig. 7) and include fine-grained biotite-magnetite alteration and garnet-K-feldspar-biotite-(quartz) alteration/veins, with the latter concentrated in the footwall (i.e., north) of the orebody. Biotite-rich rocks commonly display strong tectonite fabrics, particularly in two shear zones associated with the upper and lower contacts of the orebody (Fig. 8). These types of alteration were overprinted by biotite-destructive hematitic K-feldspar alteration that was pervasive in the immediate host rocks of the orebody (Fig. 8). The associated, and only partly coincident, enrichments of K, Fe, and Mn define marked lithochemical anomalies around the orebody (Fig. 7).

Mineralization was associated with a strong brecciation event that selectively affected the hematitic K-feldspar-altered rocks. Breccias display a range of textures from incipient crackling, grading to matrix-dominated breccias and locally contain imbricated elliptical clasts, suggesting that they may have developed in a brittle-ductile regime. Copper and Au are strongly coenriched in two main plunging lenses separated by weakly brecciated and mineralized rocks (Figs. 8, 9). Grade largely corresponds to the proportion of magnetite (20–25% of the ore), variably arsenian and cobaltian pyrite (ca. 9 wt % of the ore), and chalcopyrite-rich matrix material in the breccias (Ryan, 1998). The orebody is crudely zoned, with increased pyrite/chalcopyrite ratios toward the periphery and in localized coarse-grained semimassive pyrite bodies (Ryan, 1998; Brodie, 2001). Significant matrix gangue minerals include calcite, quartz, biotite, garnet, and chlorite. Chalcopyrite and native Au are the only significant ore minerals, although the ore is geochemically complex and contains a host of minor minerals, including epidote, allanite, titanite, tourmaline, fluorite, REE fluorocarbonates, monazite, barite, apatite, scheelite, arsenopyrite, bismuthinite, cobaltite, glaucodot, greenockite, galena, molybdenite, native bismuth, hessite, sylvanite, Bi telluride, uraninite, rutile, coffinite, and brannerite (Twyerould, 1997; Ryan, 1998; Mark et al., 2000; Brodie, 2001). Subordinate late-stage mineralization formed discordant veins that commonly contain distinctive magnetite pseudomorphs after hematite, similar to those at Candelaria (cf. Marschik et al., 2000).

Carbonate deposition appears to have become progressively greater as the style of mineralization evolved from breccia matrix to veins. Calcite is also generally more abundant in the lower ore lens and, along with subordinate amounts of dolomite, predominates in an extensive postore vein system developed below the footwall contact of the orebody that constitutes what has previously been referred to as “marble matrix breccia” (Ryan, 1998).

Gawler craton, Australia

General: The Gawler craton, which underlies a large area of southern Australia, is separated from another cratonic block to the east, the Curnamona province, by Neoproterozoic continental supracrustal rocks preserved in the early Paleozoic Adelaide fold belt (Fig. 10). The giant Olympic Dam Cu-Au-U deposit occurs beneath 300 to 400 m of Neoproterozoic and Cambrian sedimentary rocks near the northeastern

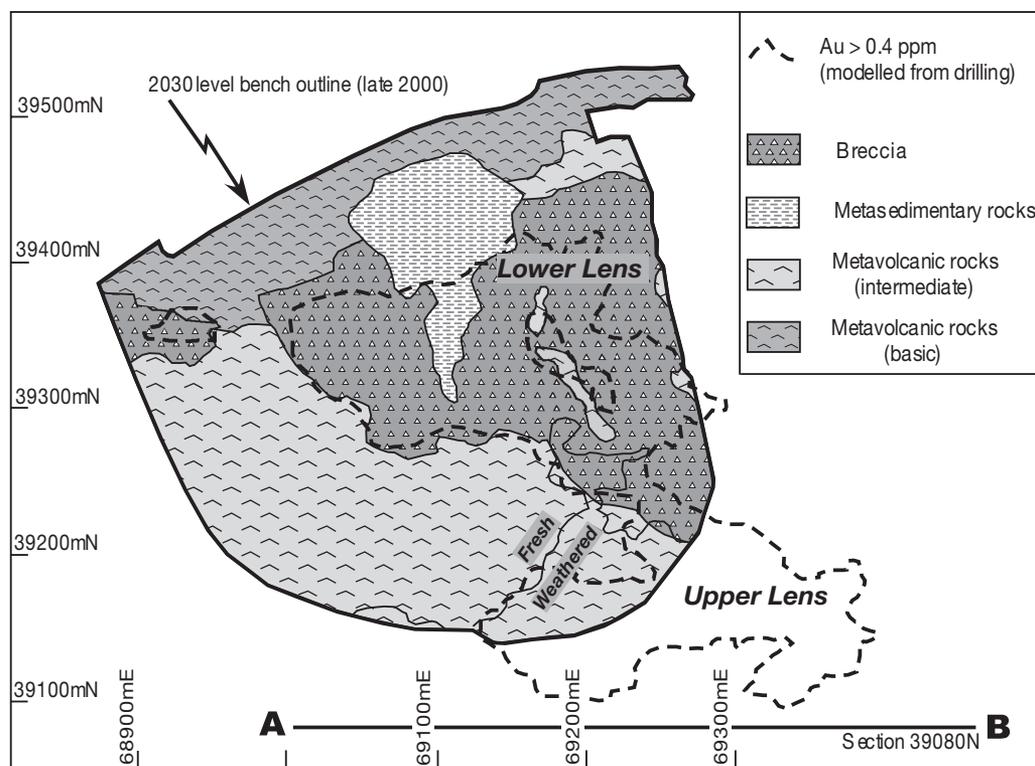


FIG. 9. Plan geology of the 2030 level bench in the Ernest Henry open cut in late 2000, showing the location of section 39080 (Fig. 8), as adapted by Brodie (2001) from mapping by Ernest Henry Mining geologists. Gold-grade distribution, which is largely undisturbed in the weathering profile, is used to illustrate the relationship between brecciated rocks and ore at this elevation.

margin of the Gawler craton (Roberts and Hudson, 1983; Reeve et al., 1990; Reynolds, 2000). The basement here is dominated by the youngest cratonic rock associations, as represented by the ca. 1590 Ma Gawler Range Volcanics, broadly contemporaneous Hiltaba Suite granitoids, and slightly younger terrestrial siliclastic sequences of the Pandurra Formation (e.g., Flint et al., 1993). Several other Fe oxide bodies with subeconomic amounts of Cu, Au, and U occur within about 50 km of Olympic Dam (Cross, 1993; Gow et al., 1994; Skirrow et al., 2002; Fig. 10). This is a distinctive association in the Australian context in that the IOCG deposits are similar in age to their host rocks and appear to have been emplaced at shallow crustal levels (cf. Williams and Pollard, 2003). IOCG deposits also occur in other parts of the craton where Hiltaba Suite granitoids intrude deformed and metamorphosed Paleoproterozoic sequences (Skirrow et al., 2002). One such setting is the Mount Woods inlier approximately 150 km west northwest of Olympic Dam, where a significant IOCG resource was recently discovered at Prominent Hill (Belperio and Freeman, 2004).

Olympic Dam—a Proterozoic hematite-dominated system: The Olympic Dam deposit (Fig. 11) has a global resource of 3,810 Mt with 1.0 wt percent Cu, making it one of the world's largest Cu deposits. The deposit also contains 0.5 g/t Au, and 400 g/t U_3O_8 (WMC Ltd. press release, 31 October 2004). A previously published underground mining reserve was 605 Mt with 1.8 percent Cu, 0.5 g/t Au, 3.6 g/t Ag, and 500 g/t U_3O_8 (WMC Ltd. Annual Report, 1999). The resource also

contains significantly elevated Ag, F, Ba, and LREE and its average Fe content is 26 wt percent, predominantly in the form of hematite (Reynolds, 2000). The deposit occurs in a ~7- × 5-km zone of brecciated and altered rock developed entirely within the Hiltaba-Suite Roxby Downs granite which has been dated at 1588 ± 4 Ma (Johnson and Cross, 1995). The breccias contain large blocks of sedimentary rocks and are interpreted to have formed within a few hundred meters of the surface (Oreskes and Einaudi, 1990; Reeve et al., 1990). The breccias are intruded by many ultramafic, mafic, and felsic dikes, which are temporally related to the hydrothermal activity. The association has been interpreted to represent a diatreme-maar volcanic setting (e.g., Haynes et al., 1995). A postore dike has been dated at 1592 ± 8 Ma (Johnson and Cross, 1995), implying that the deposit formed soon after emplacement of the Roxby Downs Granite.

The system is cored by a pipelike body of barren hematite-quartz breccia surrounded progressively by a complex zone of heterogeneous multiphase breccia and an outer zone of granite-rich breccia and hematized granite (Fig. 11; Reeve et al., 1990; Reynolds, 2000). Clasts include variably altered granite, rocks dominated by various different textural types of hematite, and a range of subordinate lithologies, including porphyritic volcanic rocks, sedimentary rocks, various dike rock types, and mineralized vein fragments. Hematite-rich breccias that host much of the ore are generally matrix supported with clast sizes mostly less than 20 cm but ranging up to tens of meters (Reeve et al., 1990; Reynolds, 2000).

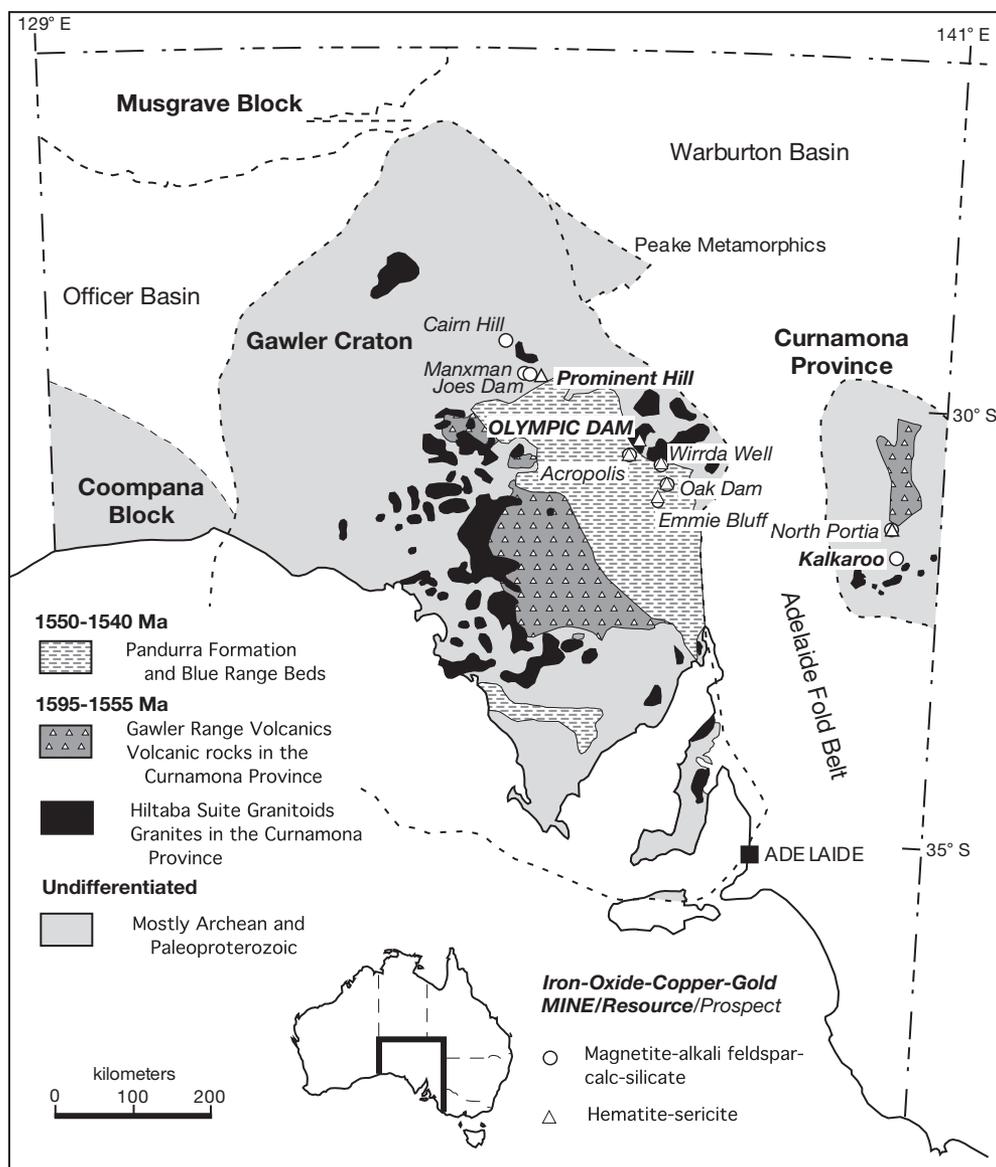


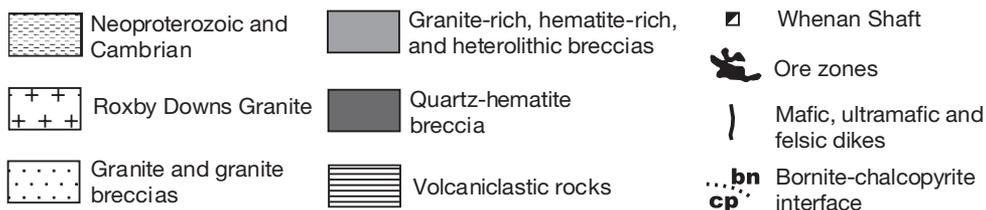
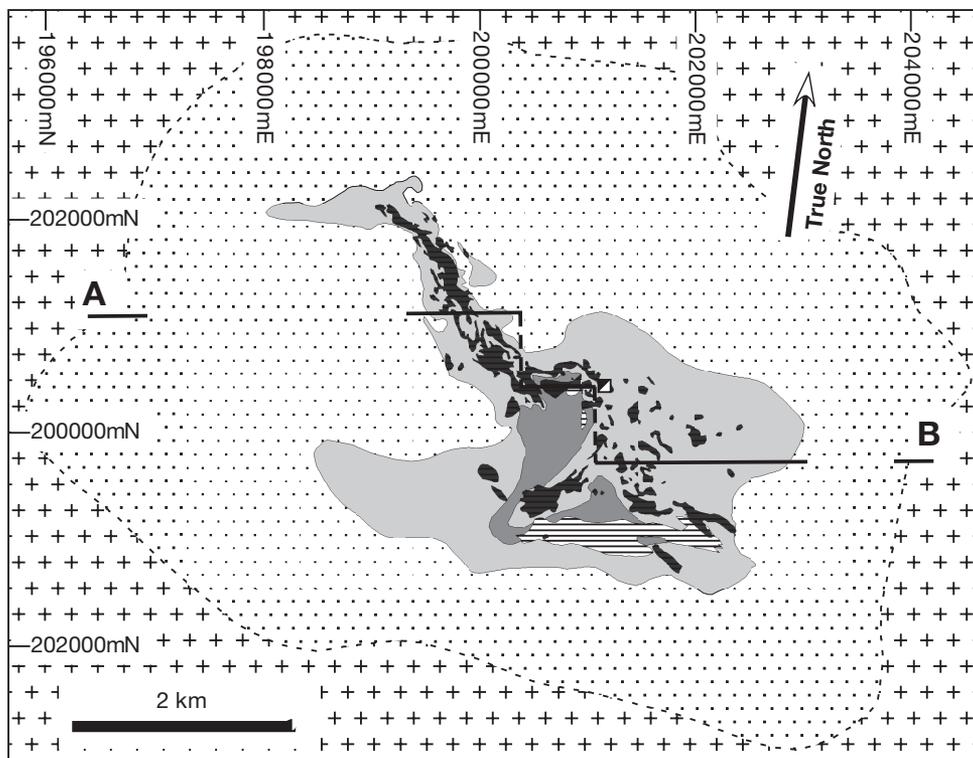
FIG. 10. Simplified geology of South Australia, showing the distribution of IOCG deposits and 1595 to 1555 Ma igneous rocks (adapted from Flint et al., 1993, with additional information from Williams and Skirrow, 2000, and Skirrow et al., 2002). Neoproterozoic and younger cover has been omitted in the cratonic areas. Note that the Gawler Range Volcanics are known to persist extensively under the Pandurra Formation and host several IOCG occurrences south of Olympic Dam, including Acropolis and part of the Emmie Bluff prospect.

The dominant alteration assemblage is sericite-hematite \pm chlorite \pm quartz \pm siderite. Magnetite is subordinate, paragenetically earlier than the hematite-phyllsilicate alteration, and concentrated in deeper parts of the system (Reynolds, 2000). Ore occurs in hematite-rich zones near the center of the complex. Barite and fluorite are abundant and La and Ce are present at high concentrations in bastnaesite, florencite, and monazite (Oreskes and Einaudi, 1990). Uranium is closely associated with Cu and occurs mostly as fine-grained to amorphous uraninite (pitchblende; Reeve et al., 1990). Mineralization was synchronous with multiple brecciation events and produced a marked hypogene zonation from barren hematite at the top, and in the core of the deposit,

through native Cu, chalcocite, bornite, and chalcopyrite to pyrite with minor chalcopyrite and magnetite (Reeve et al., 1990; Haynes et al., 1995; Reynolds et al., 2000). Economic grades occur close to the bornite-chalcopyrite interface (Fig. 11).

Fluid inclusions associated with paragenetically early magnetite, pyrite, and siderite were trapped at significantly higher temperatures than those of main-phase mineralization (Oreskes and Einaudi, 1992). Magnetite-quartz oxygen isotope pairs suggest temperatures of 400° to 500°C and fluid $\delta^{18}\text{O}$ of 7 to 10 per mil. Conan-Davies (1987) described further high-temperature inclusions with very high salinities (40–70 wt % salts), multiphase daughter salt assemblages

a



b

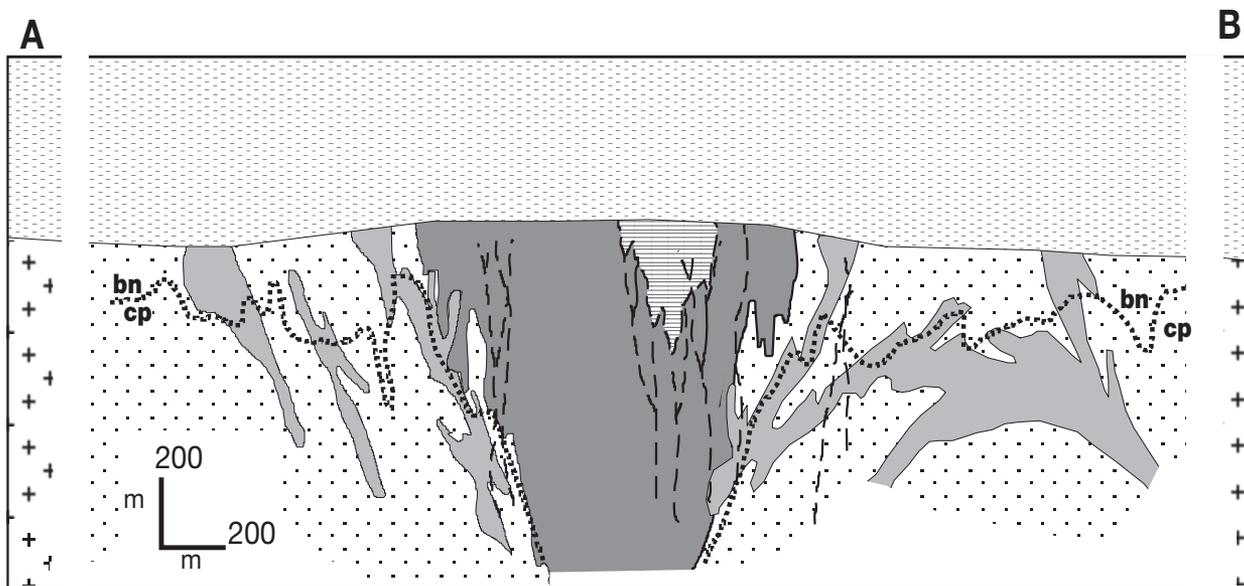


FIG. 11. (a). Geology and ore distribution at around 400 m below land surface at Olympic Dam (adapted and simplified from Reynolds, 2000). (b). Olympic Dam cross section, showing generalized geologic relationships and location of the bornite-chalcopyrite interface (adapted and simplified from Reeve et al., 1990).

including Fe-bearing solids, homogenization temperatures up to 580°C, and phase behavior suggesting entrapment at pressures of 50 to 100 MPa. These coexist with three-phase H₂O-LCO₂-VCO₂ inclusions estimated to contain up to 26 mol percent CO₂. This is a similar fluid inclusion assemblage to that recorded at the Ernest Henry deposit where complex brine inclusions containing carbonate daughter minerals coexist with liquid CO₂-dominated inclusions (Mark et al., 2001). Main-phase mineralization at Olympic Dam was from cooler fluids with variable salinity and low $\delta^{18}\text{O}$, ranging from -2 to +6 per mil (Oreskes and Einaudi, 1992). Johnson and McCulloch (1995) showed that Nd isotope data imply a mantle component of REE in the ore and suggest a genetic connection to the mafic and/or ultramafic dikes, and perhaps to a larger mafic and/or ultramafic body at depth; the latter could be partly responsible for the gravity and magnetic anomalies associated with the deposit (Roberts and Hudson, 1983; Campbell et al., 1998).

Central Andean coastal belt

General: The youngest economically significant IOCG province occurs discontinuously for some 2,000 km along the coasts of Peru and northern Chile, forming a distinctive component of the world's premier Cu-endowed region (Fig. 12;

Sillitoe and Perello, 2005). It is also an important Fe ore province, where many large apatite-bearing magnetite \pm hematite deposits are associated with extensively alkali-metasomatized rocks (Bookstrom, 1977; Espinoza, 1990; Hawkes et al., 2002). These Fe oxide-rich ore systems lie west of, and are present at generally lower elevations than, the many large Andean Tertiary porphyry Cu deposits (Fig. 12). However, the coastal IOCG province also contains porphyry-style deposits that are commonly Au rich compared to most of the younger porphyry deposits in the Andes. The province is further characterized by numerous Cu and Cu-Ag deposits in breccia bodies and mantos, some of which may be genetically related to the IOCG deposits (Sillitoe, 2003).

The IOCG and apatite-bearing Fe ore systems are diachronous, and both appear to have accompanied the eastward migration of arc magmatism in the Late Jurassic through the Late Cretaceous (Sillitoe, 2003). The arc magmas were predominantly calc-alkaline and to a large extent emplaced during extension of the old continental margin that also produced shallow marine basins with carbonate and evaporite sedimentary rocks. The Coastal Cordillera is characterized by extensive arc-parallel shear zones and brittle faults, such as the Atacama fault zone in northern Chile. Such structures were active during arc magmatism, controlled basin development, and had a strong controlling influence on the distribution of both IOCG and apatite Fe ore deposits (e.g., Espinoza, 1990; Atherton and Aguirre, 1992; Brown et al., 1993; Sillitoe, 2003). IOCG mineralization styles include veins, hydrothermal breccias, replacement mantos, and calcic skarns, with the larger deposits generally exhibiting several of these in combination (Sillitoe, 2003). Most of the apatitic Fe orebodies have Cu- and Au-bearing hypogene sulfide parageneses that typically overprint the main-stage Fe oxides (e.g., Vivallo et al., 1995; Hawkes et al., 2002).

Candelaria-Punta del Cobre, Coastal Belt, Chile—a Phanerozoic magnetite-rich system: The Candelaria-Punta del Cobre mining district (e.g., Marschik and Fontboté, 1996; Marschik et al., 1997; Ullrich and Clark, 1999; Marschik and Fontboté, 2001a; Mathur et al., 2002) is in northern Chile, near the city of Copiapó (Fig. 12, App. Fig. A2). The IOCG deposits define an approximately 5- × 20-km discontinuously mineralized belt along the eastern margin of the Copiapo batholith. The Candelaria mine is the largest deposit with reserves in 2000 of 470 Mt at 0.95 wt percent Cu, 0.22 g/t Au, and 3.1 g/t Ag, whereas several medium- and small-size mines of the Punta del Cobre district contain an estimated combined reserves plus production of >120 Mt with 1.5 wt percent Cu, 0.2 to 0.6 g/t Au, and 2 to 8 g/t Ag (Marschik et al., 2000). The latter mainly lie a few kilometers northeast of Candelaria (App. Fig. A2).

The layered rocks in the Candelaria-Punta del Cobre district record a Late Jurassic (Berriasian) transgression on a subaerially exposed volcanic basement (Marschik and Fontboté, 2001b). The volcanic and volcanoclastic rocks of the Bandurrias Group, representing a Lower Cretaceous volcanic arc, interfinger to the east with the marine limestones of the Chañarcillo Group (up to 2,000 m thick), which were deposited in a back-arc environment (Fig. A2). Marine conditions in the back-arc region ceased in the late Aptian or Albian. Calc-alkaline diorite, granodiorite, tonalite, monzodiorite,

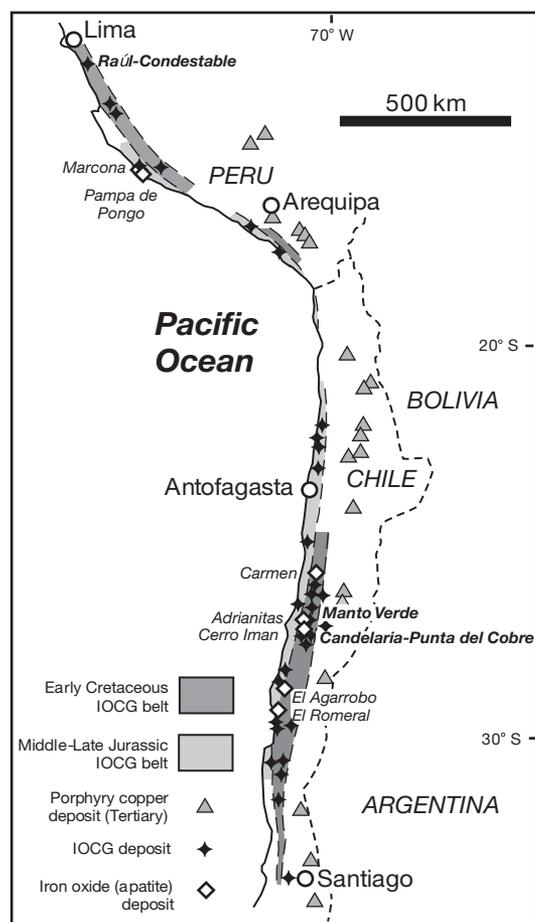


FIG. 12. Distribution of IOCG and large Fe oxide (apatite) ore deposits in Peru and northern Chile (adapted from Sillitoe, 2003, with additional information from Espinoza, 1990, and Hawkes et al., 2002).

and quartz monzonite plutons of the Copiapó Plutonic Complex (119–97 Ma; Zentilli, 1974; Arévalo, 1994, 1995, 1999; Ullrich et al., 2001) intrude the volcanic and sedimentary rock in the western part of the district (Fig. A2). These subalkaline to marginally alkaline metaluminous plutons intruded a volcanic arc and are magnetite series and I type. They have initial $^{87}\text{Sr}/^{86}\text{Sr}$ from 0.7031 to 0.7032, initial $^{143}\text{Nd}/^{144}\text{Nd}$ from 0.51273 to 0.51278, and $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ from 18.43 to 18.77, 15.55 to 15.60, and 38.13 to 38.40, respectively (Marschik et al., 2003a, b). The isotope signatures and trace element distributions suggest that the magmas were derived from a subduction fluid-modified mantle source without significant crustal contamination through assimilation.

The Cu-Au ore occurs as massive sulfides in veins, in the matrix of hydrothermal breccias, discontinuous stringers or veinlets in hydrothermally altered rocks or superposed on massive magnetite replacement bodies, and as lens-shaped, bedding-concordant bodies in which the sulfides are replacements or pore space fillings in the clastic rocks (Marschik and Fontboté, 2001a). The orebodies are hosted mainly in volcanic and volcanoclastic rocks in the upper part of the Punta del Cobre Formation (Bandurrias Group) or, locally, also in the lower part of the overlying calcareous late Valanginian Abundancia Formation (Chañarcillo Group). The intersection of northwest-to north-northwest-trending brittle faults with the contact of massive volcanic and volcanoclastic rocks controls most of the larger orebodies in the district (Fig. 13; Marschik and Fontboté, 1996, 2001a). In the Candelaria mine, portions of the ore occur in foliated biotitized rocks of a

broadly north-northeast-trending shear zone with an average 50° west dip (Candelaria shear zone).

Copper-Au ore contains magnetite and/or hematite, chalcopyrite, and pyrite. Gold occurs mainly as tiny inclusions in chalcopyrite, filling microfractures in pyrite, and as Hg-Au-Ag alloy (Hopf, 1990; Ryan et al., 1995; Marschik and Fontboté, 2001a). Locally, there are pyrrhotite, sphalerite, traces of molybdenite, arsenopyrite, and elevated concentrations of LREE. Ore formation was a multistage event associated with a complex sequence of widespread pervasive and more localized fracture-controlled alteration (Marschik and Fontboté, 2001a). The rocks in the district were affected by various superposed alteration types with variable intensities, and there is a lateral and vertical zonation at district scale and within the deposits. The most impressive feature is the voluminous sodic \pm calcic alteration in volcanic, sedimentary, and intrusive rocks at district scale (App. Fig. A3). The alteration is manifest as albite or sodic plagioclase and/or sodic scapolite with or without calcic amphibole, pyroxene, and/or epidote. The rocks outside the large sodic alteration zone are affected by propylitic alteration and/or contact metamorphism (Marschik et al., 2003a). Within the pervasively Na-metasomatized volcanic and volcanoclastic rocks, there are locally rock volumes with intense potassic \pm calcic alteration. Both rocks with sodic \pm calcic or potassic \pm calcic alteration assemblages may contain economic Cu-Au deposits and both alteration types are common in a mine. Larger Cu-Au orebodies tend to be hosted in rocks with potassic alteration assemblages, although well-mineralized Na-metasomatized rocks are common in the upper portions of the mines (see below). Some ore zones occur exclusively within the latter rocks.

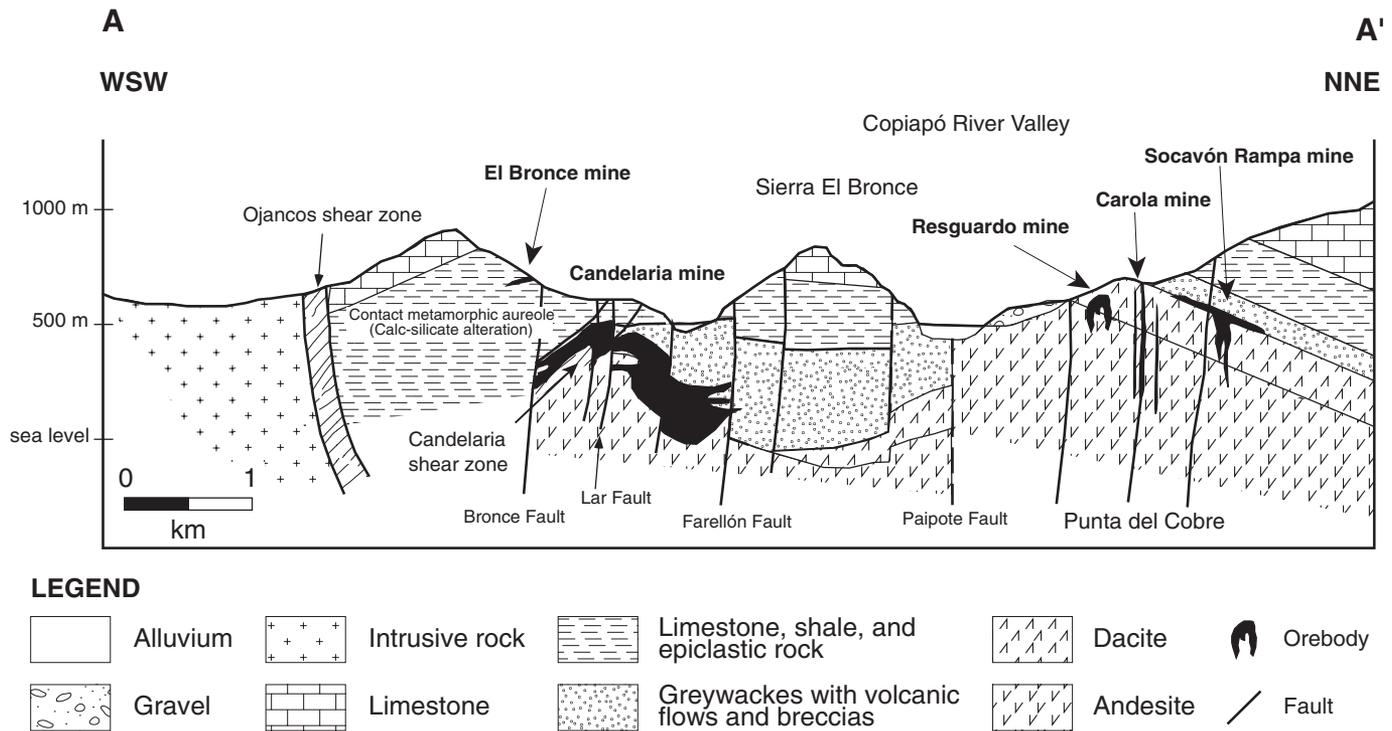


FIG. 13. Schematic section through the Punta del Cobre district (from Marschik and Fontboté, 2001a). For the location of the section, see the geologic map of Figure A2.

Outcrop and open-pit mining combined with mine workings and underground drilling in the district provide an understanding of the overall system and variations with depth. At deeper levels, chalcopyrite shows a close spatial association with calcic amphibole at Candelaria (Fig. 14). Calcic amphibole ± epidote, biotite, potassium feldspar ± sodic plagioclase are commonly associated with the ore in these parts of the system, depending on the type of host rocks. Magnetite is ubiquitous and occurs as massive magnetite bodies with or without superimposed sulfide mineralization, whereas hematite is scarce. Scattered quartz veins are widespread. Calcite is uncommon at depth, whereas anhydrite occurs locally in veins and veinlets.

The intermediate levels of the system, represented by the central upper portions of the Santos and Carola mines in the Punta del Cobre district, are characterized by potassic alteration (biotite and/or potassium feldspar). Calcic amphibole ± epidote, sodic plagioclase and/or anhydrite may be present locally.

Toward the distal (shallower) portions of the system, chlorite forms at the expense of biotite and amphibole, and albitization, chloritization, and carbonatization increase in intensity. The distal portions, represented by the Socavón Rampa mine in the Punta del Cobre district, are characterized by albite-chlorite ± sericite-rich assemblages. Pervasive carbonatization and calcite veining are common and may be intense,

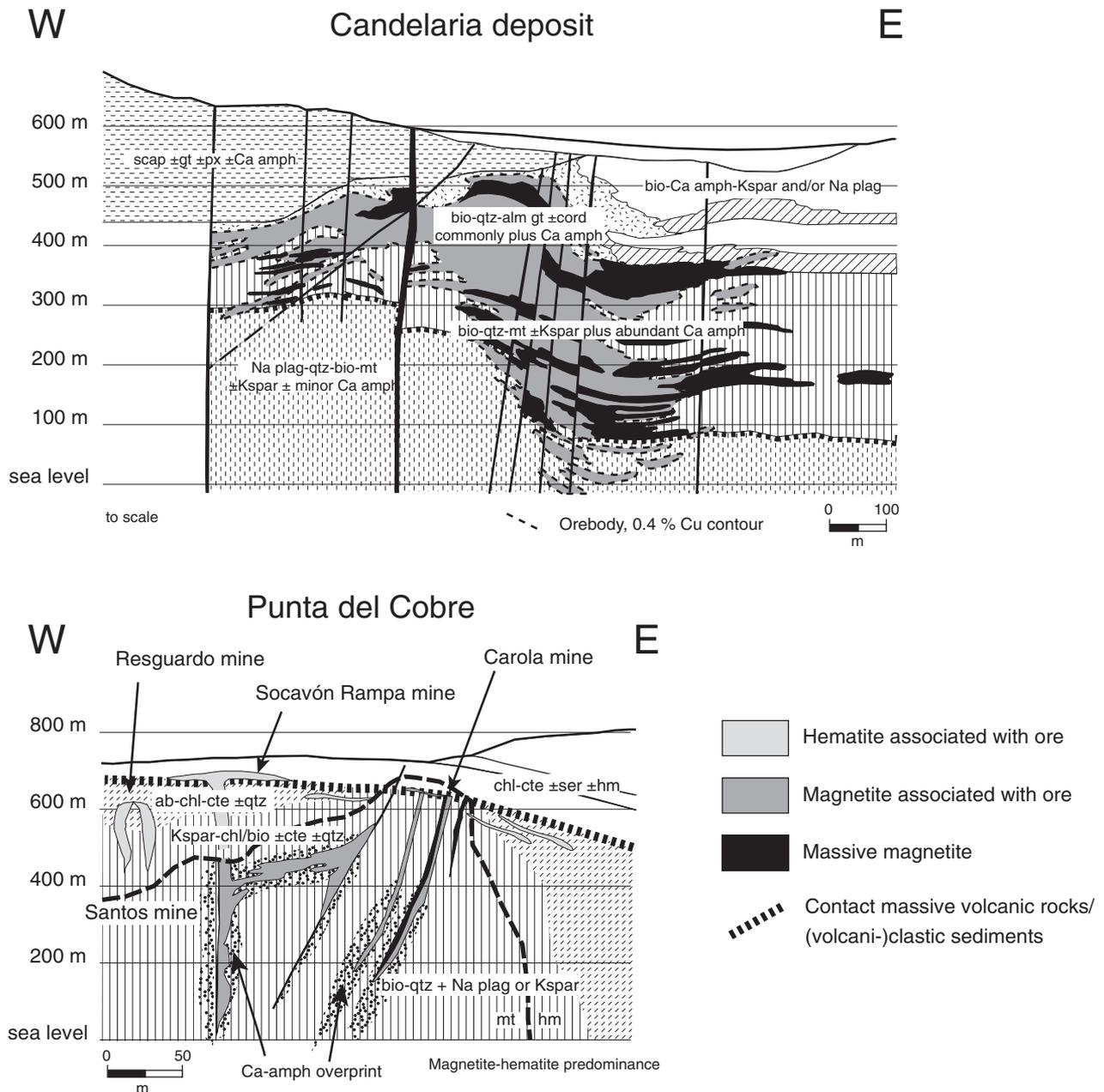


FIG. 14. Schematic sections through the Candelaria deposit and selected deposits at the Punta del Cobre (from Marschik and Fontboté, 2001a).

whereas anhydrite is absent. Hematite is the predominant Fe oxide species, whereas magnetite is minor or only locally significant.

Mineralization can be schematically described as a retrograde process, although indications for the prograde evolution (replacement of hematite by magnetite) and several mineralization pulses are recorded (Marschik et al., 2001a). Marschik and Fontbote (2001a) reported that the paragenetic sequence is subdivided into a high-temperature Fe oxide stage (600°–500°C), characterized by pervasive magnetite-quartz-biotite alteration; a main sulfide stage (500°–300°C) with chalcopyrite-pyrite; and late-stage mineralization (<250°C) with hematite-calcite and locally minor sulfides. Sulfides from Candelaria and several other deposits in the nearby Punta del Cobre district have $\delta^{34}\text{S}_{\text{CDT}}$ values largely between –3.2 to +3.1 per mil (e.g., Rabbia et al., 1996; Ullrich and Clark, 1999; Marschik and Fontboté, 2001a), with some higher values for late-stage mineralization (up to 7.2‰; Ullrich and Clark, 1999) or in the marginal parts of the system (up to 6.8‰; Rabbia et al., 1996). These data are consistent with sulfur derived largely from a magmatic source with minor contributions from the evaporite-bearing sedimentary host sequence in the peripheral portions of the orebody and during late-stage mineralization (Ullrich and Clark, 1999; Marschik and Fontboté, 2001a). Oxygen isotope compositions, combined with preliminary microthermometric data of fluid inclusions, suggest that magmatic dominated fluids (or alternatively silicate-equilibrated nonmagmatic fluids) were involved in the main Cu mineralization at Candelaria (Marschik et al., 2000; Marschik and Fontboté, 2001a). Isotopic and fluid inclusion data indicate that external nonmagmatic fluids were clearly present during the later stages of hydrothermal activity (Ullrich and Clark, 1999; Marschik et al., 2000; Ullrich et al., 2001) and the volumes of the early sodic alteration and isotopic results suggest a partial nonmagmatic source in the early events as well (M.D. Barton and E.P. Jensen, unpub. data).

Alteration ages cluster around 116 to 114 and 112 to 110 Ma, suggesting prolonged hydrothermal activity in the area (e.g., Marschik et al., 1997; Ullrich and Clark, 1999; Marschik and Fontboté, 2001a). Re-Os ages of 115.2 ± 0.6 and 114.2 ± 0.6 Ma of molybdenite from Candelaria are interpreted to record the time of the main mineralization (Mathur et al., 2002). These dates indicate that ore formation was coeval with the emplacement of the plutons of the Copiapó Complex, and a calculated initial $^{187}\text{Os}/^{188}\text{Os}$ of 0.36 ± 0.1 from an isochron, based on data of hydrothermal magnetite and sulfide from Candelaria, falls in the same range as calculated initial $^{187}\text{Os}/^{188}\text{Os}$ of 0.21 to 0.41 for igneous magnetite in nearby plutons (Mathur et al., 2002). The Chañarcillo Group reached its full thickness of 2 km in the late Aptian, and since the Cu-Au orebodies formed at stratigraphic levels close to the base of the Chañarcillo Group, the depth of ore formation corresponds approximately to the thickness of the latter.

Manto Verde, Coastal Belt, Chile—a zoned Phanerozoic magnetite-hematite system: The Manto Verde deposit lies in the Los Pozos district, in the Chilean Coastal Cordillera about 100 km north of Copiapó. The district hosts a number of other IOCG and small magnetite deposits (App. Fig. A4). Mining commenced in this area in the early 1800s and

continued intermittently until the present. Manto Verde is the largest mine with a resource of 120 Mt of supergene Cu oxides with an average grade of 0.72 wt percent Cu (at 0.2 wt % Cu cutoff) estimated before startup in 1995. Mineable reserves were 85 Mt with 0.82 wt percent total Cu (Vila et al., 1996). In the mid to late 1990s, exploration led to a substantial addition of the identified resources of Cu oxides, on the order of 180 Mt with an average grade of 0.5 wt percent Cu overlying a sulfide resource of more than 400 Mt with 0.52 wt percent Cu (Zamora and Castillo, 2001).

The Los Pozos district lies in a structural segment that is bound to the east and west by two branches of the north-south-trending Atacama fault zone (AFZ; Fig. A4), which is a subduction-related, arc-parallel wrench fault system that extends for over 1,000 km along the Chilean coast (e.g., Scheuber and Andriessen, 1990). The Manto Verde fault, a prominent N15° to 20°W-trending, 40° to 50°E-dipping brittle structure that cuts this segment, controls the main portion of the Cu-Au ore (App. Figs. A4, A5). The area is composed mainly of andesitic flows and breccias of Jurassic and/or possibly Early Cretaceous age, which may correlate with either the Jurassic La Negra Formation or the Early Cretaceous Bandurrias Group. The volcanic rocks are intruded by Cretaceous granitoids of the Las Tazas plutonic complex (~130 Ma; Berg and Breitzkreuz, 1983; Wilson et al., 2000) occur in the western part of the Manto Verde deposit and diorites, monzodiorites, granodiorites, and tonalites of the Remolino Plutonic Complex (~127 Ma; Berg and Breitzkreuz, 1983; Dallmeyer et al., 1996) lie to the east (Fig. A4). The Las Tazas plutonic complex has $^{87}\text{Sr}/^{86}\text{Sr} = 0.7033$ to 0.7034 and ϵNd (130) values of 5.1 to 6.4 (Berg and Baumann, 1985; Hodgkinson et al., 1995), suggesting a mantle provenance of the magmas without significant crustal contamination. $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology for mylonites, Al in hornblende barometry, and concordance of magma emplacement and hornblende cooling ages suggest that the plutons were emplaced syntectonically at shallow crustal levels above the brittle-ductile transition (Dallmeyer et al., 1996). Shear zones in the wall rocks along some of the contacts of the plutonic complexes are explained by heat transfer from these plutonic magmas to permit the shallow-level ductile deformation.

The Cu-Au ore is hosted mainly in specularite-dominated tabular breccia bodies (Manto Verde), breccia pipes (Manto Ruso and Manto Monstruo), and stockwork bodies (Monte Cristo; Vila et al., 1996). The Manto Verde orebody lies subparallel to the orientation of the Manto Verde fault (Vila et al., 1996) on a slightly misoriented fault segment. Other Cu-Au deposits also occur on more north-west-trending bends of this fault or the eastern branch of the Atacama fault zone, or at the intersection of the Manto Verde fault and related second-order structures (Fig. A4), implying that active fault dilation most likely localized deposits along these master faults. Ore minerals are mainly chrysocolla, brochantite, antlerite, and atacamite with minor quantities of malachite and cuprite. The Cu minerals are closely associated with Fe oxides (Zamora and Castillo, 2001). Specularite-rich ore occurs mainly in the footwall of the Manto Verde fault in the northern part of the district, whereas in the southern part Cu oxides associated with magnetite occur in both the hanging-wall and footwall

blocks of Manto Verde fault. The main specularite-dominated orebodies are hydrothermal breccias with fragments of volcanic rocks in a specularite-rich matrix. In the periphery of these bodies, there is Cu mineralization associated with specularite stockwork, which represents a transitional facies toward the barren wall rocks (termed Transition zone at Manto Verde; Fig. A5). Magnetite-rich Cu-Au ore, with variable quantities of magnetite and specularite, occurs as breccias, stockwork, and disseminations in tectonized zones at the intersection of the Manto Verde fault and secondary faults (Altavista and Montecristo, Fig. A4).

Primary sulfides are present beneath the level of supergene oxidation at a 40- to 100-m depth at Manto Ruso (Orrego and Zamora, 1991) and up to a 200-m depth at Manto Verde (Vila et al., 1996). At Manto Ruso, the transition between oxidized zone and primary ore is relatively sharp. Chalcopyrite and pyrite occur disseminated in the specularite breccia, in veinlets, and in aggregates. Late calcite gangue occurs down to a depth of 320 m. Copper grades in the sulfide zone of Manto Ruso vary between 0.6 and 0.8 wt percent (Orrego and Zamora, 1991). The situation is similar in the Manto Verde deposit, where chalcopyrite is disseminated in the matrix of the hydrothermal specularite breccia (termed Manto Atacama) and as coarse grains in veinlets and veins intergrown with calcite (Vila et al., 1996). In the central part of the district, south of Manto Verde, chalcopyrite and pyrite are associated with magnetite (Zamora and Castillo, 2001). In the footwall of the Manto Verde fault, the sulfides are hosted by andesitic volcanic and deformed volcanoclastic rocks (termed Manto Verde Breccia at Manto Verde). Grades of the hypogene mineralization along this fault are variable, locally exceeding 1 wt percent Cu with about 0.25 g/t Au (Zamora and Castillo, 2001). Elevated REE concentrations also occur locally in the Manto Verde deposit (R. Marschik, unpub. data).

There is a vertical zonation in the distribution of the Fe oxide species with magnetite at depth and specularite at shallow paleodepth consistent with other IOCG systems (e.g., Hitzman et al., 1992; Marschik and Fontboté, 2001a; Fig. A5). The north-south zonation at district scale of sulfide-associated Fe oxide species, in which magnetite is largely absent in the north and increases in abundance to the south, is possibly the result of differences in the erosion level (Orrego and Zamora, 1991).

The distribution of main ore and alteration types is shown in a schematic vertical section through the Manto Verde deposit (Fig. A5). Potassic alteration, which affected the rocks at district scale, is thought to be related to the intrusion of granite dikes (Vila et al., 1996). K-metasomatized rocks contain K-feldspar-chlorite with minor quartz and hematite assemblages. Hydrothermal biotite is commonly chloritized and only preserved locally (Vila et al., 1996). The original rock texture of pervasively K-metasomatized rocks remained intact. Toward the Manto Verde fault, an increase in quartz and sericite, at the cost of K-feldspar, is notable. Hypogene mineralization is associated with a chlorite-quartz-rich alteration assemblage with moderate to strong sericitization, and calcite, hematite, and/or magnetite (Vila et al., 1996; Zamora and Castillo, 2001). Crosscutting relationships of veinlets observed in the Transition zone at Manto Verde suggest that the chlorite-sericite-quartz phase postdates K-feldspar alteration

(Vila et al., 1996). Tourmaline is disseminated in the fragments of the Manto Atacama breccia together with K-feldspar, chlorite, sericite, quartz, specularite, and clay minerals. Late carbonate occurs in preexisting K-feldspar and quartz veinlets (Vila et al., 1996). Carbonate (calcite, siderite) veins and breccias that may contain Cu oxides occur at several places in the district (Fig. A5; Orrego and Zamora, 1991; Zamora and Castillo, 2001). In the sulfide zone, the carbonate is locally accompanied by coarse-grained chalcopyrite. Whether this chalcopyrite is cogenetic or occurs as fragments, as is the case in some veins at Punta del Cobre, is unclear. At Manto Verde, carbonate veins cut the specularite breccia, indicating that they formed late in the evolution of the system (Vila et al., 1996).

There are preliminary microthermometric data from fluid inclusions in quartz that accompanied main IOCG mineralization and from inclusions in late calcite with and without associated chalcopyrite (Campos, 1994, *in* Vila et al., 1996). Liquid-dominated three-phase inclusions in quartz have salinities between 32 and 56 wt percent NaCl equiv and homogenization temperatures of mostly between 215° and 340°C, with a maximum value of 500°C. Two-phase inclusions are possibly close to NaCl saturation (26 wt % NaCl equiv) and homogenize mainly between 180° and 260°C, reaching a maximum homogenization temperature of almost 400°C (Vila et al., 1996). The coexistence of liquid- and vapor-rich inclusions in quartz suggests boiling (Vila et al., 1996). Fluid inclusions (mainly three-phase) in late calcite without associated chalcopyrite have a mean homogenization temperature of 240° with a maximum at 360°C. Their salinities are between 32 and 40 wt percent NaCl equiv. Inclusions in calcite with associated chalcopyrite are exclusively liquid-dominated two-phase inclusions that homogenize mainly between 210° and 280°C, with a maximum at 310°C. Their salinities range between 14 and 21 wt percent NaCl equiv. The data are compatible with ascending saline, oxidized, moderately to relatively high temperature ore-bearing hydrothermal fluids that boiled, cooled, and became diluted, probably by mixing with surficial meteoric water. K-Ar ages of 117 ± 3 Ma of sericite from an andesite of the transition zone and of 121 ± 3 Ma of sericite from a granite dike are interpreted as minimum ages for the mineralization (Vila et al., 1996; Fig. A5).

Raúl-Condestable, Coastal Belt, Peru—a Phanerozoic magnetite(-hematite) system: The IOCG deposit of Raúl-Condestable is located 90 km south of Lima, on the Peruvian coast (Fig. 12), on the western side of the Peruvian coastal batholith (~107–37 Ma; Pitcher et al., 1985, and references therein). It has a cumulative production of >32 Mt at 1.7 wt percent Cu, 0.3 g/t Au, and 6 g/t Ag. The ore occurs as mantos and veins containing chalcopyrite-pyrite-pyrrhotite-magnetite with actinolite. Previous work on the Raúl-Condestable deposit includes Ripley and Ohmoto (1977, 1979) and Cardozo (1983), who proposed a volcano-exhalative genesis, whereas Injoque (1985), Atkin et al. (1985), and Vidal et al. (1990) favored it to be a skarn. Recently, de Haller (2000), de Haller et al. (2001, 2002), and Injoque (2002) classified Raúl-Condestable as an IOCG deposit based on its replacement character, structural controls, alteration associations including sodic-calcic styles, and abundant hypogene Fe oxides.

The stratigraphy in the Raúl-Condestable district dips ~40° west-southwest such that the geologic map broadly corresponds to an oblique section within the Berriasian to middle Aptian (Late Jurassic to Early Cretaceous) volcano-sedimentary host sequence, here divided into Unit I to Unit V from

bottom to top. The whole sequence is more than 6 km thick and the main characteristics of each unit are shown in Figure 15. All magmatic rocks in the deposit share typical arc geochemistry. Unit IV is a 2- to 3-km-thick dacite-andesite volcanic package intruded by a subvolcanic quartz-diorite

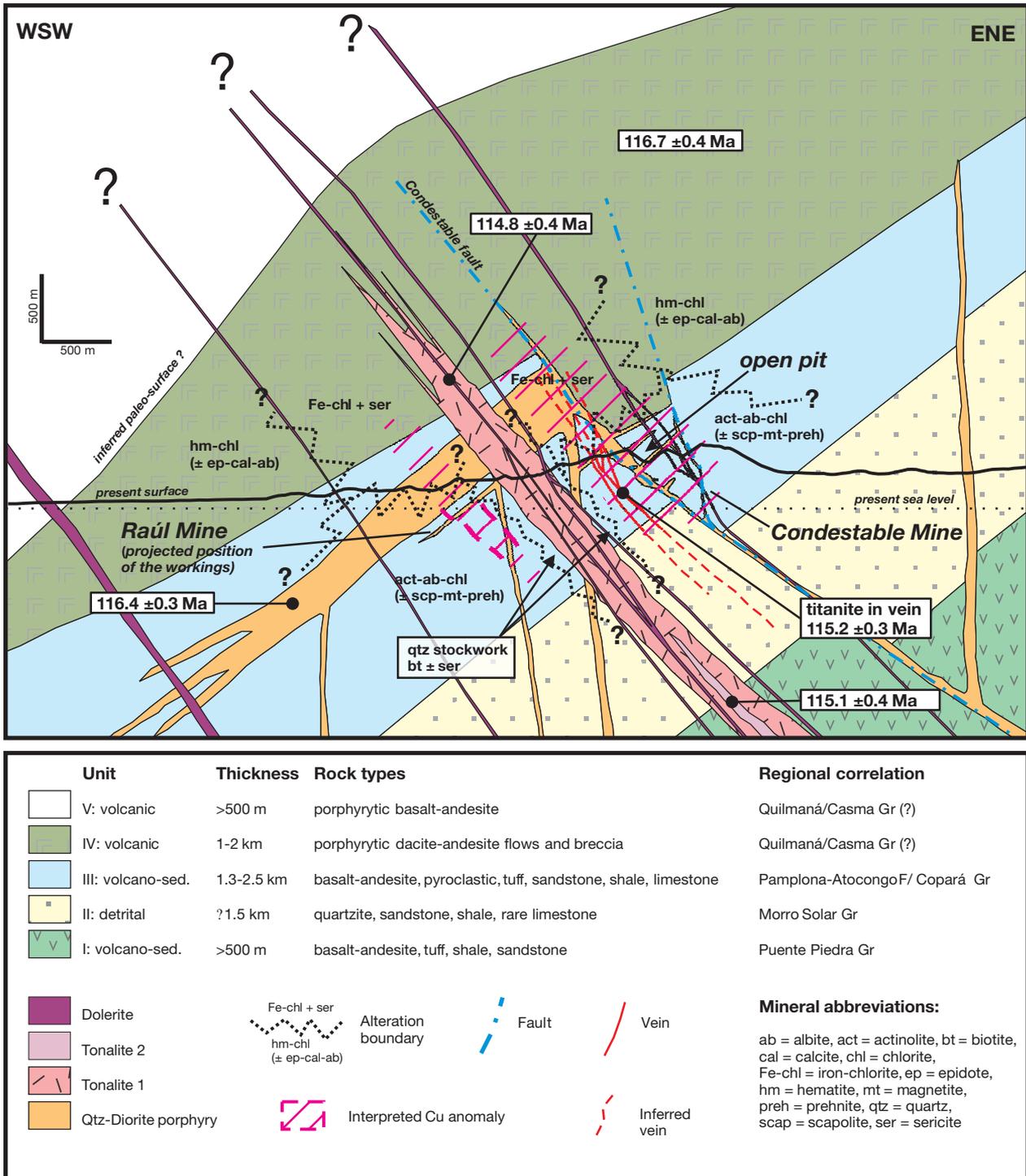


FIG. 15. Raúl-Condestable section 0-BASE. Geology and U-Pb ages are from de Haller et al. (unpub. data) The Condestable mine lies in this section, whereas the Raúl mine workings are located 0.5 to 2 km farther south and have been projected. The shape of the Cu anomaly is interpreted from surface ICP geochemistry, mine data, and extrapolation from other areas of the district.

porphyry sill-dike complex. Porphyry dikes follow northeast, north northeast, and northwest orientations. Northeast- and north-northeast-trending dikes are near vertical, while the northwest-trending Condestable fault dike (Fig. 16) dips 35° to 45° to the east with pre- to synintrusion normal movement of up to 500 m. Most of the sills intrude the upper part of Unit III, while the main and uppermost sill forms a laccolithic structure up to 500 to 600 m thick that extends 4 km along strike (Fig. 16). The center of this porphyritic sill-dike complex is successively cut by two tonalitic intrusions (Fig. 16). Tonalite 1 forms a stock with minor apophyses, 1.6 km long and up to 300 m across. Tonalite 2 is a dike, 200 m long and 30 m across, that cuts Tonalite 1. A late, regional, northwest-trending and east-dipping dolerite dike swarm crosscuts all the intrusions, the volcano-sedimentary sequence, and the IOCG mineralization (Fig. 16) The west-southwest tilting of the host sequence probably took place after dolerite emplacement.

Conventional U-Pb data on zircon and hydrothermal titanite by de Haller et al. (2002), and de Haller et al. (unpub. Data; Fig. 16) indicate that the quartz-diorite porphyry sill-dike complex and the Unit IV dacite-andesite volcanic dome formed at around 116.5 Ma, whereas Tonalite 1 and Tonalite 2 intruded at around 115 Ma. Hydrothermal titanite of the IOCG mineralization gives a U-Pb age (115.2 ± 0.3 Ma) coeval with the tonalitic magmatism and only slightly younger than the overlying volcanic edifice (Unit IV).

The IOCG deposit surrounds the Tonalite 1 stock and is mainly developed in Unit III. The deposit forms mantos and

disseminations around feeder veins in carbonate rocks, tuffs, pyroclastic deposits, and volcanic breccias. Feeder veins trend northeast, northwest, and east-west and are broadly perpendicular to the bedding. The veins cut the volcano-sedimentary sequence from Unit II to Unit IV, the quartz-diorite porphyry, and locally the Tonalite 1. Mineralization in units II and IV occurs only as veins. The alteration pattern (Fig. 16) has an early biotite core cut by a quartz stockwork surrounding, but almost exclusively outside, Tonalite 1. The alteration is zoned outward to actinolite-albite-chlorite (\pm magnetite-scapolite) and upward to Fe chlorite and sericite assemblages. Actinolite veinlets cut the biotite alteration. Late prehnite and pumpellyite locally overprint the actinolite-albite-scapolite assemblage. An upper distal alteration halo, consisting of hematite-chlorite (\pm epidote-calcite-albite) laterally surrounds the Fe chlorite + sericite and actinolite-albite-chlorite assemblages.

A composite paragenetic sequence based on vein and coarse-grained open-space fillings within the main mineralized area is shown in Figure 17. Actinolite metasomatism was coeval with the quartz stockwork and the Fe oxides that occur as open-space fillings and massive replacement bodies. Hematite was deposited first, and then completely pseudomorphous and eventually overgrown by magnetite. Scapolite (the Na end-member marialite) occurs as crystals up to tens of centimeters long, in or close to feeder veins. The dated hydrothermal titanite precipitated together with K-feldspar in veinlets crosscutting scapolite. The subsequent sulfide stage started with minor molybdenite, then pyrrhotite, pyrite with

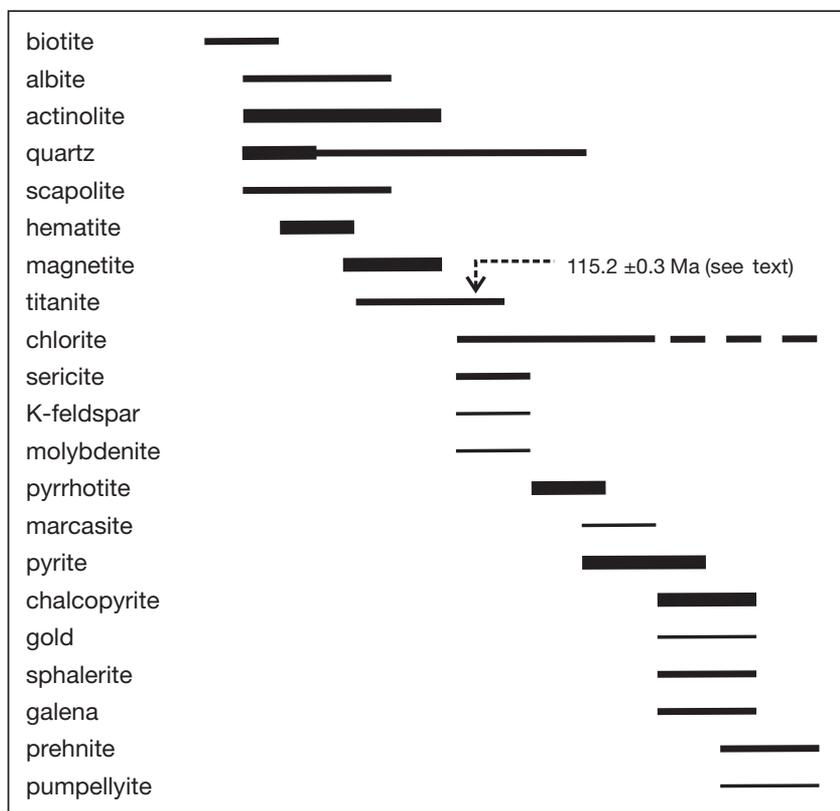


FIG. 16. Composite paragenetic sequence within the main IOCG mineralized area at the Raul-Condestable mines.

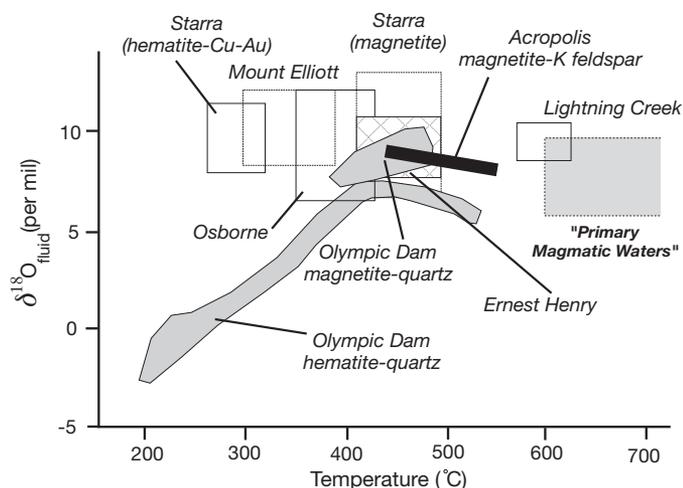


FIG. 17. Summary $\delta^{18}\text{O}_{\text{fluid}}$ -temperature diagram for various Australian Proterozoic IOCG systems (as adapted by Partington and Williams, 2000, from Oreskes and Einaudi, 1992, with additional data from Adshear, 1995, Twyrould, 1997, Rotherham et al., 1998, Perring et al., 2000, and Wang and Williams, 2001).

minor marcasite, and ended with chalcopyrite, Au, and minor sphalerite and galena. Gold occurs as inclusions within chalcopyrite. Prehnite and pumpellyite formed late close to feeder veins. Zoning both in veins and mantos from proximal to distal relative to the feeder veins (but also late to early within the paragenetic sequence) is chalcopyrite, pyrite, pyrrhotite, and magnetite.

Copper-Fe sulfides are characterized by $\delta^{34}\text{S}$ values ranging from 2.7 to 26.3 per mil (CDT), with a peak at around 5 per mil ($n = 20$) (de Haller et al., 2002) that might indicate a significant contribution of reduced marine or evaporite sulfate (Ripley and Ohmoto, 1977). A significant contribution of magmatic-sourced sulfur is proposed but cannot be proved (or dismissed) by the sulfur isotope data. Hypersaline fluid inclusions with halite and Fe chloride daughter crystals (SEM-EDS determination) occur in stockwork quartz, together with vapor-rich inclusions.

Multielement geochemistry from grade-control samples from four crosscuts of veins and six samples of blasted manto ore shows Cu values up to 20 wt percent in veins and 7 wt percent in mantos. Gold and Ag correlate with Cu and can respectively reach values over 2 and 100 ppm. Zinc, Pb, and Mo are typically less than 2,500, 500, and 100 ppm, respectively. Cobalt and Ni are anomalous but less than 400 ppm. Uranium and Th are below 5 ppm, and F can be up to 0.17 wt percent. Locally, REE can be enriched, with La up to 260 ppm, and Ce up to 500 ppm.

The zoned pattern of the alteration, with potassic alteration near the tonalitic stock grading outward to sodic-calcic and upward to phyllic assemblages, suggests a direct connection between the tonalite intrusions and mineralization. This is also supported by the U-Pb ages, which show that the mineralization is coeval with this magmatism. Moreover, the high Fe chloride content of the hypersaline fluid inclusions in the quartz stockwork, which closely surrounds the tonalite, suggests that the latter could be the source of the Fe. Sulfide deposition was subsequent to extensive magnetite formation, which largely replaced early specular hematite, thus following

a similar evolution to that described by Marschik and Fontboté (2001a) at La Candelaria.

Only about 1.5 m.y. separated the build up of the Unit IV dacite-andesite volcanic edifice and related quartz-diorite porphyry sill-dike complex from the tonalitic intrusions. The close time and space relationships existing between these two magmatic events strongly suggest they are part of the same magmatic system, the tonalite intrusions corresponding to the last magmatic events that occurred in the Raúl-Condestable volcanic center. The top of Unit IV is interpreted as broadly corresponding to the paleosurface at the time of the deposit formation (Fig. 16), thus implying that mineralization occurred at a paleodepth of 2 to 3 km in a subvolcanic setting.

Discussion: Origin of Iron-Oxide Copper-Gold Deposits

Synopsis of evidence

Most IOCG deposits have mineral assemblages that imply the ore fluids were oxidized and sulfide poor, although host rocks locally appear to have influenced the formation of more reduced, pyrrhotite-bearing assemblages as at Eloise in the Cloncurry district (Baker, 1998) and at Raúl Condestable (de Haller et al., 2002). Fluid inclusions indicate that the ore fluids were saline and the overall impression is that these deposits formed from fluids with high Cl/S (cf. Barton and Johnson, 1996).

The geologic evidence presents a complex picture with key features including the following:

1. There is evidence for both shallow and deeper crustal settings and the associated implications for the involvement of surficial fluids in the former but their likely absence from deposits in ductile regimes.
2. There is generally strong, but not universal, evidence for coeval magmatism with different regions having different compositional intrusive suites. IOCG deposits are associated with igneous rocks that range from diorites to granites and there is no clear association with alkaline magmatism. Mineralization and associated alteration occurred at submagmatic temperatures.
3. There are large volumes of spatially associated and typically sodic-altered rocks. The elemental depletions in these may, among others, include Fe, Cu, and Au, but this in itself does not indicate that the alteration systems represent the source of the diagnostic ore components.
4. The lack of any clear universal tectonic control, although some (e.g., Barton and Johnson, 1996) have argued for evaporitic or other nonmagmatic brine sources that would be consistent with tectonic and in some cases climatic controls.

Stable isotope evidence includes variable $\delta^{34}\text{S}$, which is typically but not universally close to 0 ± 5 per mil. $\delta^{18}\text{O}$ values are largely consistent with rock equilibration, although surficial water contributions appear to be significant in some cases (e.g., Olympic Dam). Radiogenic isotope data are consistent with rock sources of Nd and Os that in some cases must have been away from the site of deposition.

Mineralizing processes

Fluid inclusion data (see summaries in Pollard, 2000; Partington and Williams, 2000; Williams and Pollard, 2003)

demonstrate a ubiquitous role for brines in IOCG genesis. However, the variable characteristics of IOCG deposits imply there were fundamental differences in the conditions and mechanisms of mineralization among deposits. Mineral parageneses, fluid inclusion data, and stable isotope geothermometry (as reviewed by Partington and Williams, 2000) suggest that Cu-Au precipitation occurred at various different temperatures between 200° and 500°C (Fig. 17). Features of IOCG deposits that influence interpretation of mineralization processes include (1) distinct to absent specific host rock controls, (2) presence or absence of Cu and/or Fe mineral zoning, (3) variable minor element chemistry, including the inconsistent but distinctive F-Ba-LREE-U association, (4) occurrence of different examples displaying either uniform or variable stable isotope signatures, (5) varying relative significance of vein, breccia, and replacement mineralization styles, (6) limited amounts of vein quartz compared to many other types of hydrothermal ore deposit, and (7) the common evidence for significant carbonic fluid components. Factors other than cooling of the ore fluid were important and may have been predominantly responsible for mineralization in many cases.

Mineralization controlled by wall-rock reactions; Some IOCG deposits are localized by specific lithologies and display textural evidence of fluid-rock reactions during sulfide deposition. The host rocks in these cases can be divided into two main categories, namely, magnetite ± hematite ironstones and carbonaceous rocks.

In ironstone-hosted deposits a geochemical influence of the host rock can be inferred where sulfides selectively replaced Fe minerals in situ or are restricted to structural sites where magnetite was altered to hematite during sulfide deposition in open spaces. The ironstones need not be products of the same events as those responsible for Cu-Au deposition, and the distribution of Cu and Au may be partly independent from ironstone at district to mine scale. In some cases, as documented from several Proterozoic examples in Australia, the driving mechanism appears to have been sulfate reduction by magnetite to form hematite (Gow et al., 1994; Rotherham, 1997; Skirrow and Walshe, 2002). In others, comparatively reduced fluids sulfidized the Fe oxides in a process that formed pyrrhotite-bearing and hematite-free ore (Huston et al., 1993; Skirrow and Walshe, 2002).

In cases where distinct reaction textures have not been described, it should not be assumed that ironstone hosts have exerted a direct chemical control on ore deposition. The association may be more influenced by structure if ironstones were more susceptible to fracturing than surrounding rocks during mineralization. It has also been suggested that other processes such as fluid mixing may have acted in conjunction with wall-rock reactions in some ironstone-hosted deposits (Williams et al., 2001; Skirrow and Walshe, 2002). A corollary is that, where present, ironstones cannot automatically be assumed to be the main loci of Cu-Au enrichment. A good illustration is the Osborne deposit in the Cloncurry district, where quartz-magnetite-chalcopryrite-pyrite ± pyrrhotite lodes occur close to, but independent of, thick but weakly mineralized magnetite-rich ironstones (Adshead et al., 1998).

Possible IOCG-related, economic Cu-Au deposits localized by carbonaceous (meta)sedimentary rocks occur in northern

Scandinavia (Lindblom et al., 1996), the Cloncurry district (Krcmarov and Stewart, 1998; Williams and Skirrow, 2000), and Rajasthan, India (Knight et al., 2002). Graphite destruction in favor of carbonate, as well as the presence of hematitic alteration, which is commonly better developed in nearby graphite-free lithologies, all suggest that redox reactions between the ore fluid and graphite contributed to sulfide deposition. In some cases, preexisting pyrite or pyrrhotite in these host rocks may also have contributed sulfur (Krcmarov and Stewart, 1998). Iron oxides are typically minor and may display an antipathetic relationship to graphitic rocks in these systems. They can be absent from the ore zones, even in cases where apparently coeval Fe oxide-rich orebodies occur in related structural settings nearby (Krcmarov and Stewart, 1998).

Mineralization without specific host-rock control: The majority of IOCG deposits, including most of the largest ones, are hosted by strongly altered normal crustal lithologies such as andesites, granitoids, and various common sorts of (meta)sedimentary rocks. In higher temperature, magnetite-dominated examples (e.g., Ernest Henry, Candelaria, Raúl-Conditable), the ore metals appear to have been deposited following complex earlier histories of high-temperature alteration. These histories may have involved early phases of albitization, although evidence for this may only be preserved outside the ore environments, where plagioclase feldspars tend to be absent, having been replaced by K-feldspar, biotite ± magnetite ± garnet, skarn, and amphiboles.

Chalcopryrite is the only economically significant hypogene Cu mineral in most magnetite-dominated IOCG deposits, with the notable exception of Salobo. Sulfides, with or without codeposited magnetite, occur predominantly as infill in veins and veinlets or as matrix in breccias. Replacement of earlier formed ferromagnesian silicates consistent with sulfidation is locally significant, as the Eloise deposit in the Cloncurry district (Baker, 1998). However, in other cases, including Candelaria and Ernest Henry, there is very little evidence for sulfidation, and as noted above, the limited amounts of codeposited quartz suggest that mechanisms other than cooling should be considered. Hematite-dominated IOCG systems include some examples in which chalcopryrite is the only significant hypogene Cu sulfide (e.g., Manto Verde) and others that have distinct chalcopryrite-bornite-chalcocite zoning (e.g., Olympic Dam, Prominent Hill).

Haynes et al. (1995) suggested that various lines of evidence imply ore deposition at Olympic Dam as the result of mixing of at least two fluids, one higher temperature and saline, of deep-seated, magmatic and/or deeply circulated meteoric origin, and the other, meteoric and oxidized. This evidence includes the Cu mineral zoning pattern, the covariation of isotopic equilibration temperatures and deduced $\delta^{18}\text{O}_{\text{fluid}}$ (Fig. 17), and the variable salinities of fluid inclusions and occurrence of inclusions with higher salinities and homogenization temperatures at depth associated with chalcopryrite. In addition, there are high overall abundances and antipathetic distributions of fluorite and barite, with the latter, highly insoluble mineral, being abundant in the upper and/or inner barren hematite-quartz breccias. Haynes et al. (1995) were also able to numerically simulate the ore mineral zoning and parageneses at Olympic Dam using a mixing scenario, with modeled end-member fluid compositions.

Detailed geochemical models of ore deposition are currently lacking for other larger IOCG systems. The observation that CO₂-rich fluid inclusions are associated with brine inclusions in many IOCG deposits has led to the suggestion that carbonic phase separation and associated pH increase could also be a significant process in mineralization (e.g., Xu and Pollard, 1999). Whereas this might be anticipated in mixed volatile fluids affected by reduced pressures during fracturing and brecciation events, its relevance remains unclear in the absence of convincing evidence for coentrapment of brine and CO₂ inclusions at sites of mineralization (Fu et al., 2003).

Although the relative roles of cooling, fluid-phase separation, and fluid mixing in IOCG deposits are poorly understood, there are persistent indicators of more than one fluid being involved in ore deposition at most deposits. Even in deposits where all stable isotope data are most consistent with ultimately magmatic derivation for oxygen, carbon, and sulfur (e.g., Ernest Henry, Candelaria), there is evidence that fluid mixing could have had an important role in ore deposition: This includes: (1) the presence of distinctive hypersaline fluid inclusions containing Fe or Ba chlorides, which given the oxidized mineral parageneses, implies very low concentrations of sulfur in one of the likely ore fluids (e.g., Starra in the Cloncurry district; Williams et al., 2001; cf. Barton and Johnson, 1996); (2) the local presence of sulfides with highly elevated $\delta^{34}\text{S}$ values (e.g., Raúl Condestable) in systems temporally associated with intrusions, implying nonmagmatic sulfur contributions but nevertheless consistent with a magmatic metal source; and (3) highly complex internal zoning of pyrites (As, Co, Ni; Brodie, 2001) and K-Ba-feldspars (Twyerould, 1997; Mark et al., 2000), and the presence of barite in the latest synore stages at Ernest Henry.

Relationship to other deposit types

Iron oxide Cu-Au deposits display spatial associations with, and/or have, similar mineral associations to several other types of Fe and Cu ore deposits. Preeminent among other deposit types that need to be considered in the context of IOCG genesis are (1) apatitic Fe oxide deposits, (2) the predominantly diorite-associated Au-rich members of the porphyry Cu deposit family and their associated Fe \pm Cu skarns that characteristically contain large amounts of hydrothermal magnetite (e.g., Sillitoe, 1997), (3) the Phalabowra carbonatite-hosted, magnetite-rich Cu deposit, (4) manto and breccia-hosted Cu \pm Ag deposits such as occur widely in the Andean IOCG belt (Sillitoe, 2003), and (5) (meta-)sedimentary-hosted syntectonic Cu deposits such as giant ore system at Mount Isa that may be coeval with the Cloncurry district IOCG deposits (cf. Haynes, 2000).

The broad-scale spatial and compositional affinities of IOCG and apatitic Fe oxide deposits have already been described and, albeit in the face of sparse geochemical evidence, could be taken to indicate a continuum. Historically this has been argued to indicate a genetic link (Hitzman et al., 1992; Barton and Johnson, 1996). However, within regions, the two deposit types display a degree of independent distribution, to a certain extent forming distinct clusters with separate structural controls (e.g., Hitzman, 2000; Bergman et al., 2001; Sillitoe, 2003). In cases where Cu and Au occur within,

or spatially overlapping with, apatitic Fe ores, the sulfide parageneses are typically younger (Barton and Johnson, 2000). Apatitic Fe oxide and IOCG deposits could therefore be products of systems that share the same general geologic environment and sources of some components, but they may also have some fundamental genetic differences between them.

Copper-Au-(Fe) porphyry and skarn deposits have distinct space-time associations with basic to intermediate, oxidized (magnetite-bearing) igneous rocks; they are widely believed to form directly from coeval magmatic fluids (Seedorff et al., 2005). One important implication is that the low sulfur mineral parageneses of IOCG deposits do not present a prima facie case for a nonmagmatic ore-fluid source. Sodic alteration is rarely significant in the mineralized parts of porphyry systems but may be present at depth, potentially extending well away from porphyry centers due to circulation of regional fluids (e.g., Dilles and Einaudi, 1992; Battles and Barton, 1995). Notably, however, some diorite-associated Cu-Au porphyry deposits display strong albitization in the ore environment, and they are spatially associated with magnetite-apatite bodies (in part overprinted by Cu-Au mineralization) and distinctive REE-enriched Cu-Au-magnetite skarns (Cann and Godwin, 1983; Meinert, 1984; Lang et al., 1995; Dilles et al., 2000; Ray and Webster, 2000). Many apatitic Fe oxide and IOCG deposits also have a regional association with oxidized and more primitive igneous rocks but lack the intimate association with subvolcanic stocks and extensive quartz-bearing vein stockworks of porphyry deposits. In the Andes, it is clear that IOCG deposits have a much closer spatial and temporal association with apatitic Fe oxide ores than with porphyry systems (Fig. 12). One contributing factor to this separation could be the depth of formation, with porphyry-related deposits tending to be preserved in less deeply eroded and typically younger belts. However, other factors must also be involved, given the inferred shallow settings and meteoric water contributions in hematite-sericite IOCG systems such as Olympic Dam, a number of the Andean deposits (see earlier section), and the 7-km-vertical section with comparable or shallower levels for IOCG compared to porphyry deposits in the Yerington district, Nevada (Dilles et al., 2000).

Groves and Vielreicher (2001) have suggested that the striking geochemical similarities between the Phalabowra carbonatite-hosted Cu deposit and the Olympic Dam deposit imply that the former should be regarded as an end member of the IOCG family. It was proposed that volatile-enriched alkaline magmas at depth could be the source of Olympic Dam mineralization. Samarium-Nd systematics at Olympic Dam suggest a mantle source for the highly enriched REE values (Johnson and McCulloch, 1995), although the sources of Cu, Au, and U are not known with any certainty. On the other hand, many IOCG deposits, as defined in the current paper, have no known association with carbonatites or any form of alkaline magmatism. Furthermore, many do not display a marked LREE-U-F signature and, perhaps significantly, these include examples in northern Sweden and the Andes that are spatially associated with archetypical apatitic Fe oxide deposits. Consequently, contributions from alkaline-carbonatite magmas could explain features of some IOCG deposits but presently cannot be invoked as essential (i.e., as noted in the

introduction, fundamentally different types of IOCG deposit may exist that could one day require separate classification).

Contrasting with possible alkaline-carbonatite affinities, the spatial associations of IOCG deposits with miscellaneous sedimentary and volcanic-hosted Cu deposits in the Mount Isa block and the Andes point to the possible roles of surficial fluids in the mineralizing systems of these districts. A popular metallogenic concept for such deposits involves the leaching of Cu from basic or intermediate volcanic rocks by oxidized brines linked to evaporitic basins (e.g., Heinrich et al., 1995). Haynes (2000) suggested that IOCG deposits and what he referred to as "Fe sulfide Cu (-Au) deposits" may be related, with the former characteristically occurring in comparatively oxidized and sulfur-deficient host rocks.

Constraints, models, and directions for future research

Critical issues in IOCG genesis that are topics of ongoing debate relate to the sources of metals, ore fluids, and other vital components of the ore-forming systems, such as Cl and S. Insight into these issues must be combined with a better documentation through field studies of the geologic characteristics of IOCG deposits and their causative hydrothermal systems.

Barton and Johnson (1996) proposed that the distinctive geochemical character of the deposits is related to fluids with distinctively high Cl/S, reflecting an evaporitic source for these elements. Others used stable isotope and fluid inclusion evidence to suggest a role for magmatic water (e.g., Gow et al., 1994; Williams et al., 1995; Rotherham et al., 1998). Williams (1994), Barton and Johnson (1996), and Oliver et al. (2004) proposed that Fe and some other elements enriched in the ores (e.g., K, Rb, Ba) were leached via regional alkali metasomatism, although the initial fluid sources may differ between different areas. In contrast, Perring et al. (2000) showed that a giant hydrothermal magnetite deposit at Lightning Creek in the Cloncurry district formed under transitional magmatic hydrothermal conditions, in a system involving at least some fluids that had high Fe and Cu contents consistent with derivation of these metals from the crystallizing magma. Conversely, Johnson and Barton (2000) showed that the tilted Humboldt system in Nevada has IOCG-like zoning and parageneses of minor Cu, Au, Co, and REE, albeit closely linked to >1 billion tons of magnetite mineralization and no evidence for magmatic fluids from the fundamentally basaltic magmas of the complex.

Barton and Johnson (2000) evaluated the similarities and differences that would be expected between nonmagmatic and magmatic brine sources and concluded that this spectrum of fluid sources should yield distinctive geological and geochemical characteristics; they suggested that IOCG characteristics are best explained by a nonmagmatic brine source. Sulfur sources remain enigmatic, partly due to the difficulty of detection in fluid inclusions, but the contrasts between Candelaria and Ernest Henry, with most calculated $\delta^{34}\text{S}_{\text{fluid}}$ values predominantly magmatic, and Raúl-Condestable, with $\delta^{34}\text{S}_{\text{fluid}}$ values reaching in excess of 20 per mil, suggest that both magmatic and near-surface (sedimentary) sources may have contributed sulfur.

Considerable advances have been made in the characterization of IOCG ore fluids using conventional fluid inclusion

studies and stable isotope geochemistry (Fig. 17; see summaries in Partington and Williams, 2000; Pollard, 2000; and subsequent studies by Mark et al., 2001; Marschik and Fontboté, 2001a; Ullrich et al., 2001; Williams et al., 2001, 2003). Fluid inclusions reveal that formation of the deposits involved fluids containing H_2O , with or without carbonic species, and high concentrations of several salts. Where present, the carbonic component is dominantly represented by CO_2 , although CH_4 is also present in some examples, typically those associated with pyrrhotitic ores and/or carbonaceous host rocks, as well as hydrothermal carbonates. The densities of carbonic fluid inclusions are consistent with entrapment pressures ranging up to several 100 MPa, consistent with the deep-seated, brittle to ductile structural settings of some deposits (e.g., Rotherham et al., 1998). Brine and carbonic components typically form discrete fluid inclusion populations and, as noted above, it is not yet clear how these formed (that is, by unmixing of a single homogeneous fluid, CO_2 effervescence resulting from fluid mixing, fluid-rock reactions at the site of entrapment, and/or different fluid inputs to the site of mineralization). Studies near Cloncurry suggest that large variations in the composition of the ubiquitous complex brine-type inclusions existed between locations within a single district, and that the composition of these brines may also have changed during the evolution of individual ore-forming systems (Mark et al., 2001; Williams et al., 2001, 2003).

Stable isotope evidence (Fig. 17), as reviewed by Partington and Williams (2000), suggests that the water in systems like those near Cloncurry was exclusively deep seated (i.e., equilibrated with magmas, igneous rocks, and/or metamorphic rocks at elevated temperatures). However, hematite-sericite systems like Olympic Dam and Prominent Hill in the Gawler craton add another dimension to the problem, as they are likely to have formed in hydrothermal systems containing large quantities of surface-derived water. Fluid inclusion and oxygen stable isotope data from Olympic Dam provide strong support for this concept in this deposit (Oreskes and Einaudi, 1992).

Various genetic models have been proposed for IOCG deposits that can be broadly divided into those involving magmatic and nonmagmatic fluid sources (as summarized by Barton and Johnson, 2004; Fig. 18; Table 1). Magmatic models involve the release of oxidized, sulfur-poor, metalliferous brines from the coeval magmas, with ore deposition subsequently driven by various processes as discussed above. The sources are variously inferred to have been primitive calc-alkaline arc magmas (e.g., Sillitoe, 2003), I-type or A-type granitoids in the debated intracratonic or distal arc settings in Australia and North America (Meyer, 1988; Pollard et al., 1998; Wyborn, 1998), and carbonatite to strongly alkaline magmas (Hauck, 1990; Groves and Vielreicher, 2001). The CO_2 content of the source is an important factor in magmatic models, not only because of the common presence of CO_2 in ore-related fluid inclusions but also because of its role in controlling fluid release from magmas over a wide range of pressures compatible with the depths inferred for IOCG systems (Pollard, 2000). It has also been suggested that the presence of CO_2 may influence alkali partitioning between silicate melts and fluids, potentially generating brines with high Na/K ratios that might be responsible for the widespread sodic alteration present in many IOCG settings (Pollard, 2001).

TABLE 1. Synopsis of Alternative Genetic Models for IOCG Systems (from Barton and Johnson, 2004, cf. Fig. 18).

Fluid source	Nonmagmatic		
	Magmatic	Basin/surface	Metamorphic
Fundamental processes	Release of S ²⁻ -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 ⁺) link to magmas Regional Na(Ca) coincident but not directly related to Cu(-Au)	K (type I), H ⁺ ± Na(Ca) in upwelling zones Na(Ca) ± K (type II) in recharge zones	Primarily K and H ⁺ 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	Arcs 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	Hauck (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)

ore-forming process(es) or an unrelated feature, which only serves to indicate that the tectonic setting, thermal structure, and stratigraphic package are appropriate for IOCG prospectivity.

Most investigators at the present time conclude that IOCG deposits are the products of fault and/or shear zone-controlled distal magmatic hydrothermal systems, some of which incorporated significant amounts of nonmagmatic fluids. If magmas made a major contribution to the deposits, there are important questions to be addressed regarding tectonic setting and petrochemical specialization leading to the generation of an Fe-Cu-rich fluid, as well as fluid evolution under the deep-seated conditions implied for deposits such as those in the Cloncurry district. This implies that the fundamental control on global IOCG distributions may relate to the specifics of magma generation in the lower crust or upper mantle. Alternately, if salts are primarily derived from basinal fluids, evaporites, or metamorphism, and the metals from leaching of rocks along the flow path or at the source (Table

1), then the implication is that prospectivity would relate to the specific paleogeographic environment or stratigraphic succession, as well as to the thermal and permeability history of the host rocks around the deposits. In both cases, a sulfur source is required for precipitation of Cu, and this may be the critical factor for economic productivity, regardless of the source of fluids and metals.

Whether or not fluid mixing or multiple sources might be the defining hallmarks of IOCG deposits remains to be demonstrated. However, recognition of proximity to likely causative intrusions, and the roles of fluid mixing versus wall-rock reaction and cooling, may ultimately permit discrimination of different deposit types within this broader class and a more focused exploration and research understanding.

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