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Geological characteristics and tectonic setting of Proterozoic iron oxide (Cu–U–Au–REE) deposits

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ABSTRACT

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Recent work on the Olympic Dam Cu–U–Au–Ag deposit, South Australia, the Wernecke Mountain breccias, Yukon, the Kiruna iron ore district, Sweden, and the southeast Missouri iron ore district, and a review of literature on other iron-rich mineral deposits in Proterozoic rocks, suggest that these occurrences constitute a distinct class of ore deposits characterized by low-titanium, iron-rich rocks formed in extensional tectonic environments. Other examples of this class may include the mineral deposits of the Great Bear magmatic zone of northwest Canada, the Bayan Obo district of China, and perhaps the Redbank breccia pipes of the Northern Territory, Australia. We designate this class of deposits as Proterozoic iron oxide (Cu–U–Au–REE) deposits, and propose that the ore deposits generally referred to as 'Kiruna-type' should be considered a subset of this larger class. Salient characteristics of this class of deposits are as follows:

(1) *Age.* The majority of known deposits, particularly the larger examples, are found within Early to mid-Proterozoic host rocks (1.1–1.8 Ga).

(2) *Tectonic setting.* The deposits are located in areas that were cratonic or continental margin environments during the late Lower to Middle Proterozoic, and in many cases there is a definite spatial and temporal association with extensional tectonics. Most of the districts occur along major structural zones, and many of the deposits are elongated parallel to regional or local structural trends. The host rocks may be igneous or sedimentary; many of the deposits occur within silicic to intermediate igneous rocks of anorogenic type. However, mineralization in many deposits is not easily related to igneous activity at the structural level of mineralization.

(3) *Mineralogy.* The ores are generally dominated by iron oxides, either magnetite or hematite. Magnetite is found at deeper levels than hematite. CO_3, Ba, P, or F minerals are common and often abundant. The deposits contain anomalous to potentially economic concentrations of REEs, either in apatite, or in distinct REE mineral phases.

(4) *Alteration.* The host rocks are generally intensely altered. The exact alteration mineralogy depends on host lithology and depth of formation, but there is a general trend from sodic alteration at deep levels, to potassic alteration at intermediate to shallow levels, to sericitic alteration and silicification at very shallow levels. In addition, the host rocks are locally intensely Fe-metasomatized.

In spite of these similarities, many variations occur between and within individual districts, particularly in deposit morphology. Individual deposits occur as strongly discordant veins and breccias to massive concordant bodies. Both the morphology and the extent of alteration and mineralization appear to be largely controlled by permeability along faults, shear zones and intrusive contacts, or by permeable horizons such as poorly welded tuffs. Thus, the variations of morphology are explicable in terms of local wall-rock and structural controls. Similarly, local variations in mineralogy and geochemistry may be largely attributable to wall-rock composition, and to P, T, and f_{O_2} controls related to depth of formation.

We believe that these deposits formed primarily in shallow crustal environments (< 4–6 km), and that they are expressions of deeper-seated, volatile-rich igneous–hydrothermal systems, tapped by deep crustal structures. The global occurrence of this type of deposit at approximately 1.8 to 1.4 Ga suggests a relation to a global rifting events effecting continental

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crust, possibly the break-up of a Proterozoic supercontinent. Secular cooling of the Earth insured that subsequent rifting and mineralizing events might generate deposits similar in kind but smaller in magnitude.

Introduction

The discovery in 1976 of the giant Olympic Dam deposit in the Stuart Shelf region of South Australia (2000 million metric tons of 1.6% Cu, 0.06% U_3O_8, 0.6 g/metric ton Au, and 3.5 g/metric ton Ag; Roberts and Hudson, 1983; Scott, 1987) has stimulated renewed interest in iron-rich deposits of the Early to mid-Proterozoic. The unusual features of this deposit—viz., the occurrence of disseminated Cu–Fe sulfides and uraninite within highly hematitic breccias, and the unusual textures of these breccias—at first led to an emphasis on its distinctive characteristics more than on its similarities with other ore deposits (Roberts and Hudson, 1983, 1984). Some workers even suggested that Olympic Dam might represent a new environment of mineralization (e.g. Nash et al., 1980). However, C. Meyer (pers. commun., 1981) recognized textural and mineralogical similarities between the rocks at Olympic Dam and ores of the southeast Missouri iron province, while Bell (1982) and Youles (1984) noted similarities to breccias in the Wernecke Mountains, Yukon, and the Mount Painter district, South Australia (Coats and Blisset, 1971; Lambert et al., 1982), respectively. Enumeration of similarities between Olympic Dam and other deposits was furthered by Hauck and Kendall (1984), Hauck et al. (1989) and Hauck (1990) who extended the analogy to include a variety of iron-rich deposits, mostly Proterozoic in age, such as the giant iron ore deposits of the Kiruna district, Sweden, and the Fe–REE deposits of Bayan Obo, China. Recently, a number of workers have accepted this general grouping and have suggested that all of these deposits may be genetically related, perhaps constituting a new class of ore deposits (Sims et al., 1987; Meyer, 1988a, b; Oreskes et al., 1989; Einaudi and Oreskes, 1990; Gandhi and Bell, 1990b). Most recently, Hauck (1990) has presented a broader overview of these deposits with reference to the exploration potential of the United States mid-continent; however, his genetic model differs considerably from that presented here.

The purpose of this paper is to demonstrate—by providing a detailed account of the geology—that Olympic Dam, Kiruna, and other low-titanium iron deposits of the Proterozoic do indeed constitute a distinct class of ore deposits, and to propose a preliminary model for their genesis. We refer to this class as Proterozoic iron oxide (Cu–U–Au–REE) deposits, and we suggest that ore deposits heretofore referred to as 'Kiruna-type' should be considered a subset of this larger class. We further suggest that these deposits formed *primarily* by shallow-level hydrothermal processes, albeit probably related to deep-seated magmatism. The data we present are based on our own recent work on the Olympic Dam deposit and nearby Fe occurrences on the Stuart Shelf, the Wernecke Mountain breccias, the Kiruna district, and the southeast Missouri iron district, and on a review of literature on similar deposits. Table 1 summarizes the deposits upon which our study is based.

The recognition of this class of deposits is important for two reasons, one scientific and one economic. First, the genesis of Kiruna and similar iron oxide deposits has been the subject of extensive controversy (Geijer and Ödman, 1974; Parak, 1975a, b; Frietsch, 1978; Wright, 1986). Genetic hypotheses include liquid immiscibility of an iron-rich melt (Daly, 1915; Geijer, 1930a; Amstutz, 1960; Park, 1961; Kisvarsanyi and Proctor, 1967; Philpotts, 1967; Murrie, 1973; Frutos and Oyarzun, 1975; Badham and Morton, 1976;

Wracher, 1976; Frietsch, 1978; Henriquez and Martin, 1978; Lundberg and Smellie, 1979; Smellie, 1980), hydrothermal exhalative processes (Parak, 1973, 1975a, b, 1984, 1985; Anderson, 1976; Roberts and Hudson, 1983; Nold, 1988) and hydrothermal replacement (Crane, 1912; Geijer, 1915, 1930a; Singewald and Milton, 1929; Meyer, 1939; Ridge, 1957, 1972; Geijer and Ödman, 1974; Bookstrom, 1977; Knutson et al., 1979; Laznicka and Edwards, 1979; Panno and Hood, 1983; Hildebrand, 1986; Hildebrand et al., 1987; Chao et al., 1989; Oreskes and Einaudi, 1990). This range of proposed processes suggests a need for a fresh examination of the available data. If the Olympic Dam and Kiruna deposits are related, then information from Olympic Dam and related deposits may help to resolve this long-standing genetic question.

The second important reason for the recognition of this class of deposits is that these Proterozoic iron deposits have been viewed traditionally solely as sources of iron ores. The addition of Olympic Dam, Bayan Obo, and the Wernecke breccias to this class of deposits indicates economic potential for Cu, U, Au, Ag, and REE ores as well.

Regional setting

Proterozoic iron oxide (Cu-U-Au-REE) deposits are located in areas that were cratonic or continental margin environments during the time of mineralization. Nearly all the districts share a geological setting suggestive of intracratonic extensional tectonics coeval with host rock deposition. Accordingly, the host rocks are generally upper crustal igneous or sedimentary rocks. In the Northwest Territory of Canada, in South Australia, in northern and central Sweden, and in southeastern Missouri, iron oxide deposits occur within silicic-alkalic volcanic and plutonic terrains. In the Northern Territory of Australia, Yukon Territory of Canada, and Inner Mongolia, China, iron oxide deposits occur within sedimentary sequences (Fig. 1).

Deposits hosted by igneous rocks

Great Bear magmatic zone, Northwest Territory, Canada

In the Great Bear magmatic zone of the Northwest Territory, Canada, a linear zone 500 km long which probably marks a former continental margin, iron-copper-uranium deposits are hosted by silicic and intermediate volcanic and plutonic rocks. Mineralization occurs within the younger magmatic belt (1.875-1.84 Ga) of the 2.1 to 1.8 Ga Wopmay Orogen (Hoffman, 1980; Hoffman and Bowring, 1984; Gandhi 1988, 1989; Gandhi and Bell 1989, 1990a). The volcanic sequence in the northern portion of the magmatic zone is up to 10 km thick and has been divided into three groups: the Labine Group in the west, the Sloan Group in the central area and the Dumas Group in the east (Barager, 1977; Hoffman and McGlynn, 1977; Hoffman 1980, 1984; Hildebrand et al., 1987). The Labine Group is subdivided into a lower part consisting primarily of basalts and rhyolites and an upper part dominated by andesites which are thought to have formed as stratovolcanoes (Hildebrand, 1981, 1982; Hoffman et al., 1976; Hoffman and McGlynn, 1977). It lies unconformably on an older, deformed basement complex (Hildebrand et al., 1983; Hildebrand 1984) and has been intruded by a number of synvolcanic, intermediate composition plutons, dated by U-Pb analysis of zircons at 1.87-1.86 Ga (Hildebrand, 1986). The andesites and intermediate plutonic rocks host much of the iron oxide mineralization described by Hildebrand (1986) and Reardon (1989, 1990) in the Echo Bay-Camsell River area. The Sloan Group overlies the Labine Group and consists dominantly of dacite-rhyodacite-rhyloite flows and ignimbrites with minor mafic volcanic rocks. The Dumas Group is a lateral equivalent of the Labine Group (Hildebrand et al., 1987) and

TABLE 1

Proterozoic Fe oxide (Cu–U–REE–Au) deposits

Region/District	Deposits	Tonnage/Grade	References
South Australia Stuart Shelf	Olympic Dam Acropolis, Wirda Well, Bopeechee, Oak Dam, Dromedary Dam	2 billion tonnes, 35% Fe, 1.6% Cu, & 0.06% U3O8	Conan-Davies, 1987; Creaser, 1989; Mortimer et al., 1988, Oreskes and Einaudi, 1990; Parker, 1987; Paterson, 1966a, b; Reeve et al., 1990; Roberts and Hudson, 1983, 1984
Northern Territory, Australia Redbank district	Bluff Sandy Flat Redbank, Eagles Nest, Tom Springs, Packsaddle, Bauhinia 50 pipes known in region	2 million tonnes, 5–10% Fe, 2% Cu 1.5 million tonnes, 8% Fe, 2% Cu	Knutson et al., 1979; Orridge and Mason, 1975, Roberts et al., 1963
Yukon, Canada Wernecke and Richardson Mountains	Pagisteel Igor Dolores Creek, Irene, Nor, Bond, Pike (Eaton), Demon (Gremlin) over 90 breccia bodies known	1 million tonnes, 29% Fe 0.5 million tonnes, 1% Cu	Archer, and Schmidt, 1978; Archer et al., 1977; Archer et al., 1986; Bell, 1986a, b; Laznicka, 1977a, b; Laznicka and Edwards, 1979; Laznicka and Gabory, 1988; Morin, 1977, 1979; Parrish and Bell, 1987; Stevens et al., 1982; Templemann-Kluit, 1981
Great Bear Magmatic Zone N.W.T., Canada	Terra Sue-Diane Damp - at least 20 other prospects	1 million tonnes, 50% Fe 8 million tonnes, 0.8% Cu	Badham, 1978; Badham and Morton, 1976; Cannuli et al., 1990; Gandhi, 1988; 1989; Gandhi and Bell, 1989, 1990a, b; Hildebrand, 1981, 1982, 1984, 1986; Reardon, 1989, 1990; Sawiuk et al., 1990
Inner Mongolia China Bayan Obo district	Main (Central) Eastern Western	20 million tons, 35% Fe, 6.19% REO 15 million tons, 33% Fe, 5.71% REO 10(?) million tons, 31% Fe, 1% REO	Argall, 1980; Bai and Yuan, 1985; Chao et al., 1989; Drew et al., 1989; Jishun et al., 1987; Philpotts et al., 1989
Northern Sweden Kiruna district - magnetic orebodies	Kiirunavaara Luossavaara Tuolluvaara Rakkurijoki	2.6 billion tonnes, 60% Fe 37 million tonnes, 60% Fe 43 million tonnes, 60% Fe	Daly, 1915; Ekstrom, 1973; Erikson and Hallgren, 1975, Forsell and Godin, 1980; Frietsch, 1978, 1979a, b, 1980a; Geijer, 1915, 1930a, 1960, 1968; Geijer and Magnusson, 1952; Geijer and Odman, 1974; Gilmour, 1985; Magnusson, 1970; Offerberg, 1967; Parak, 1973, 1975a, b, 1984, 1985; Vollmer et al., 1984; Welin, 1971
Kiruna District–Per Geijer magnetite/hematite and Hauki hematite orebodies	Lappmalmen Nokutusvaara Henry Rektorn Haukivaara	100 million tonnes, 40% Fe 8 million tonnes, 42% Fe 7 million tonnes, 45% Fe 11 million tonnes, 33% Fe 8 million tonnes, 50% Fe	

Svapparaara district 35 km SE of Kiruna	Gruvberget Leveaniemi Mertainen Painirova Altavaara Tansari, Kuosanen, Ainsjarvi, Ylipaasnjaska	74 million tonnes, 55% Fe 300 million tonnes, 60% Fe 165 million tonnes, 34% Fe 36 million tonnes, 30% Fe 35 million tonnes, 26% Fe	Erikson and Hallgren, 1975; Frietsch, 1980a, b, c; Lundberg and Smellie, 1979
Stuor-Ratek district 20 km SW of Kiruna	Ekstromberg Tjdorika Renhagen Harrejaure Skokimjokk Tjarrojakka (50 km SW of Kiruna) Pattock (40 km SW of Kiruna) Piedjastjakka, Patovare, Fustiljakk	37 million tonnes, 55% Fe 3 million tonnes, 57% Fe 5 million tonnes, 40% Fe 1 million tonnes, 30% Fe 65 million tonnes, 36% Fe 68 million tonnes, 45% Fe	Frietsch, 1974; Offerberg, 1967
Gallivare (Malmberget) district	18 deposits	total of approximately 930 million tonnes, 55% Fe	Geijer, 1930b; Geijer and Magnusson, 1952; Parak, 1973
Central Sweden Bergslagen district	Grängesberg deposits-Exportfeld-Risbergsfaltet-Blotberget Idkerberget Lekonberg-total of 16 deposits in 30×5 km belt	400 million tonnes, 55% Fe 10 million tonnes, 62% Fe	Aronsson, pers. commun., 1987; Frietsch, 1974; Geijer and Magnusson, 1952; Lundstrom and Papunen, 1986; Magnusson, 1970
Southeast Missouri, USA St. Francois Mountains outcrop area	Iron Mountain Pilot Knob (surface) Pilot Knob (subsurface) Cedar Hill, Shephard Mountain, Iron Knob, Greasy	30 million tons, 30–38% Fe 2 million tons, 50% Fe 22 million tons, 50% Fe	Anderson, 1976; Bickford et al., 1981; Bickford and Mose, 1975; Brandom et al., 1985; Cullers et al., 1981; Crane, 1912; Emery, 1968; Einaudi and Oreskes, 1990; Kisvarsanyi, E., 1980, 1981; Kisvarsanyi, 1975; Kisvarsanyi and Proctor, 1967; Kisvarsanyi and Smith, 1988; Marikos et al., 1989; Meyer, 1939; Murphy and Ohle, 1968; Murrie, 1973; Nold, 1988; Nuelle et al., 1989; Panno and Hood, 1983; Ridge, 1957, 1972; Sides et al., 1981; Singewald and Milton, 1929; Snyder, 1968, 1969; Spurr, 1927; Wracher, 1976
Northern subsurface area	Pea Ridge Bourbon Boss Bixby Kratz Springs (North Sullivan), Camels Hump	136 million tons, 56% Fe 200 million tons, 40% Fe 50–80 million tons, 20% Fe	

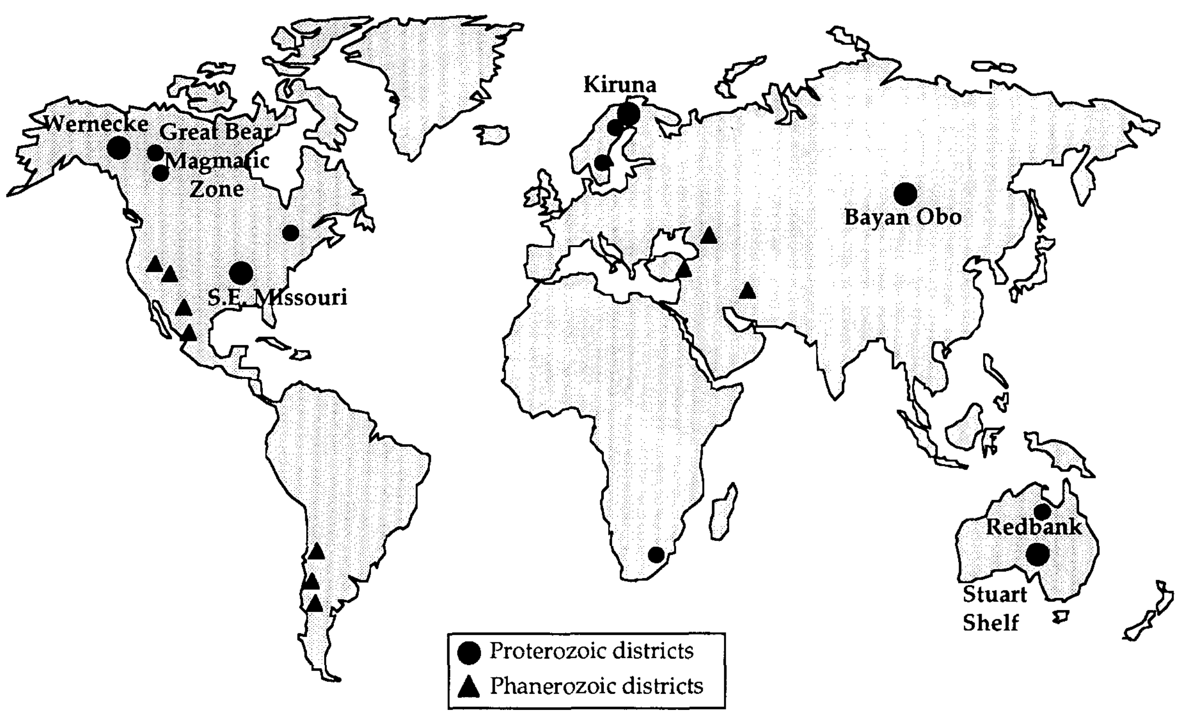

Fig. 1. Location of Proterozoic iron oxide (Cu–U–REE–Au) deposits.

comprises a bimodal assemblage of rhyolites and basalts. Large felsic to intermediate plutons intrude rocks of all three groups. In the southern portion of the Great Bear magmatic zone the volcanic rocks are predominantly rhyodacite and rhyolite and have been intruded by a number of quartz monzonite plutons (Gandhi 1988, 1989; Gandhi and Lentz, 1990). Iron oxide mineralization in the southern area is primarily hosted by felsic volcanic rocks.

Kiruna district, northern Sweden

The Kiruna district, and the satellite iron districts at Svappavaara, Stuor-Ratek and Gällivare in northern Sweden, occur in a mid-Proterozoic continental setting (1.85–1.8 Ga from U–Pb dating, Skiöld and Cliff, 1984; Skiöld, 1987). The host rocks consist of voluminous alkalic rhyolite, trachyte and trachyandesite ash flows and lava flows, with subsidiary cogenetic intrusive rocks that grade upward into a continental sedimentary succession (Geijer, 1930a; Offerberg, 1967; Geijer and Ödman, 1974; Parak, 1975a; Frietsch, 1979b; Forsell and Godin, 1980; Frietsch, 1980a, b; Vollmer et al., 1984). The volcanic rocks overlie a Lower Proterozoic greenstone basement containing intermediate, mafic and ultramafic volcanic rocks and minor shallow marine sediments, which in turn unconformably overlies Archean granitic gneiss (Offerberg, 1967; Forsell and Godin, 1980). The distribution of the volcanic rocks in the Kiruna region suggests emplacement within grabens or calderas (Geijer, 1968). However, later intrusion of potassic granites, syenites and granites at 1.5–1.6 Ga from Rb–Sr ages (Welin et al., 1971) and tectonism has obscured the evidence of possible large-scale structural controls.

Bergslagen district, central Sweden

The Grängesberg iron oxide deposits are located in the northern Bergslagen district of Sweden. The Bergslagen district is composed of metavolcanic and metasedimentary rocks intruded by synvolcanic granitoids (Åberg et al., 1984). The metavolcanics consist of rhyolitic to dacitic flows, ignimbrites and volcaniclastic sediments with volumetrically minor basalts (Vivallo and Rickards, 1984). Radiometric dating of the granitoids has shown they were formed in the Early Proterozoic between 1.87 and 1.85 Ga (Welin et al., 1980; Åberg et al., 1983, 1984; Åberg and Strömberg, 1984). This Early Proterozoic sequence is thought to be juvenile crust, locally overlying Archean basement (Huhma, 1986; Patchett et al., 1987). Rickard (1988) suggested that the volcanic rocks of the Bergslagen district formed in an extensional environment.

St. Francois terrane, southeastern Missouri

In southeastern Missouri, iron oxide deposits occur in a mid-Proterozoic ($\sim$1.5 Ga) continental setting, the St. Francois terrane, which consists primarily of alkalic rhyolite and cogenetic granite, with lesser basalt, trachyte, and trachyandesite (Kisvarsanyi, 1981). The terrane is thought to consist of numerous ring complexes and resurgent calderas with volcanic rocks preserved mostly within the calderas (Sides et al., 1981; Kisvarsanyi, 1980, 1981). Emplacement of the igneous rocks appears to have been controlled by northwest-trending basement structures (Sims et al., 1987). A continental rift or proto-rift tectonic setting during the Middle Proterozoic has been proposed (Kisvarsanyi, 1975; Kisvarsanyi, 1980).

Stuart shelf, South Australia

In South Australia, iron oxide deposits occur within silicic–intermediate igneous rocks of the Middle Proterozoic basement of the Stuart Shelf region. Exploratory drilling into the largely unexposed basement has revealed Archean and Lower Proterozoic gneiss, calcareous and hematitic metasedimentary rocks, and deformed granites, which are intruded and overlain by undeformed Middle Proterozoic granitoids and silicic–intermediate volcanics of the Burgoyne batholith (Paterson, 1986a; Parker, 1987). The older basement may be broadly correlative with the Archean and Lower Proterozoic rocks of the Gawler Craton that bounds the Stuart Shelf to the south and west (Roberts and Hudson, 1983; Paterson, 1986a; Parker, 1987; Mortimer et al., 1988); the deformed granites are syntectonic intrusions attributed to the Kimban Orogeny (1.8–1.6 Ga; Webb et al., 1986). The undeformed Middle Proterozoic grantitoids are coarse-grained, A-type granitoids, emplaced at approximately 1.59–1.6 Ga (Table 2) and are thought to form part of the Hiltaba supersuite of plutonic rocks, 1.59–1.45 Ga (Rutland et al., 1980; Cooper et al., 1985; Webb et al., 1986; Creaser, 1989; Reeve et al., 1990). Associated felsic to intermediate volcanic rocks may be correlative with anorogenic silicic volcanic rocks (Giles, 1988) that are abundant in the Gawler Range Province (Gawler Range Volcanics, 1.59–1.49 Ga), and may be genetically related to the Hiltaba plutonic rocks. The Olympic Dam deposit is hosted by the undeformed, Middle Proterozoic Roxby Downs granite. Clasts of felsic to intermediate ash-flow tuffs and flows are present in the Olympic Dam breccia complex (Oreskes and Einaudi, 1990) and other deposits on the Stuart Shelf, such as the Acropolis deposit, southwest of Olympic Dam (Paterson, 1986 a,b). These volcanic rocks are similar in composition and texture to the Gawler Range volcanic rocks. North-northwest-trending photo-lineaments that transect the Olympic Dam area may reflect extensional grabens superimposed on the east-northeast trending igneous terrane at $\sim$1.4 Ga (O'Driscoll, 1982; Roberts and Hudson, 1983). These structures apparently focused mineralization, and subsequently controlled the distribution of Middle–Late Proterozoic continental red beds (Pandurra Formation) and younger dolerite dike swarms and intrusions (Parker, 1987).

TABLE 2

Age of host rocks and mineralization

Region/District	Age of host rocks	Age of mineralization
South Australia Stuart Shelf	Middle to late Proterozoic, 1.85–1.6 Ga from U–Pb zircon ages in the Roxby Downs granite (Creaser, 1989; Reeve et al., 1990). Igneous rocks in the adjacent Gawler Range which are petrologically similar yield 1.45–1.59 Ga ages (Rutland et al., 1980; Cooper et al., 1985; Webb et al., 1986).	Rb–Sr dating of sericite from the altered Roxby Downs granite yields 1.31 Ga date (Gustafson and Compston, 1979). The maximum U–Pb date on pitcheblende from Olympic Dam is 1.4 Ga (Trueman, 1986). A minimum age of 676 ± 200 is derived from Rb–Wr dating of the Woomera Shale overlying the Olympic Dam deposit (Cooper et al., 1985).
Northern Territory Australia Redbank District	1.6–1.5 Ga from Rb–Sr on volcanic rocks (Knutson et al., 1979)	Not known with certainty; may be close to host rock age.
Yukon, Canada Wernecke Mtns.	Galena from Gillespie Lake Group (uppermost Wernecke Supergroup) yields model Pb age of 1.28 Ga (Morin, 1979).	Monozoite from NOR breccia yields 1.27 Ga U–Pb date (Parrish and Bell, 1987); phlogopite and biotite from Quartet Mtn. breccias give K–Ar dates of 1.51 and 1.04 Ga, respectively (Archer et al., 1977; Godwin et al., 1982); wide spread of U–Pb dates, the oldest is 1.19 Ga (Archer et al., 1986).
Great Bear Magmatic Zone N.W.T., Canada	U–Pb on zircons from plutonic rocks hosting the mineralization yields 1.86–1.87 Ga (Hildebrand, 1986).	Geologic evidence indicates age of mineralization overlaps age of igneous activity (1.87 Ga).
Inner Mongolia, China Bayan Obo District	Middle Proterozoic sediments, 1.7–1.85 (Drew et al., 1989) or 1.4–1.5 (Qui, 1990).	U–Pb dates on monozite yield 1.4–1.5 Ga (Philpotts et al., 1988); La–Ba dating of monozite yields 1.35 Ga (Nakai et al., 1989); Nd–Sm dating of allanite–huanghoite–parisite yields an age of 1.43 Ga (Nakai et al., 1989). A range of late Proterozoic to Paleozoic ages have been determined by K–Ar, Rb–Sr, Th/Pb (Nakai et al., 1989; Chao, 1990; Qui, 1990) and probably indicate late thermal resetting due to Carboniferous and Permian igneous activity.
Northern Sweden Kiruna District	Whole rock Rb–Sr dates from volcanic rocks yields spread of ages 1.6–1.9 Ga (Welin et al., 1971). The Kiruna sequence is bracketed in age by the Haparanda and Lina intrusive suites which have been dated by U–Pb in zircons as 1.89–1.85 Ga and 1.8–1.77 Ga respectively (Skiöld, 1987; Skiöld and Cliff, 1984). Whole rocks Sm–Nd isotope compositions of footwall and hanging wall rocks lie on a 1.89 Ga isochron (Cliff et al., 1990).	Volcanic dikes cut and are cut by the magnetite mineralization at Kirunavaara and clasts of mineralized rock are found in the overlying Hauki (Vakko) sediments suggesting that mineralization occured during or immediately following deposition of host volcanic rocks. Based on a 1.88 Ga U–Pb zircon date on post-ore dikes, ore formation occurred between 1.88 and 1.90 Ga (Cliff et al., 1990).
Southeast Missouri	U–Pb dating of host volcanic rocks yields 1.4–1.5 Ga ages (Bickford and Mose, 1975).	Post ore dike at Pea Ridge gives a Rb–Sr date of 1.29 Ga (Snyder, 1968). Hematite mineralization at the Pilot Knob (surface) deposit appears to be syngenetic or early diagenetic. These data suggest that mineralization occured roughly synchronously with volcanism (1.4 Ga).

Deposits hosted by sedimentary rocks

Redbank district, Northern Territory, Australia

The Redbank district of the Northern Territory of Australia may be considered transitional between igneous-hosted and sedimentary-hosted deposits in that copper–iron deposits occur in a volcano–sedimentary sequence. Mineralization occurs within breccias that cut the the Middle Proterozoic Tawallah Group (1.6–1.5 Ga), a mixed sequence of volcanic and sedimentary rocks in a continental setting (Knutson et al., 1979). The Redbank district is located along a series of major faults cutting the Wearyan Shelf, a sedimentary platform developed adjacent to the Middle Proterozoic Batten Trough (Orridge and Mason, 1975), which has been interpreted as an intracontinental rift. Voluminous silicic to intermediate volcanics (Fiery Creek Volcanics) occur adjacent to the Batten Trough and southwards to the Mount Isa Trough and are probably related to an extensional event. The Tawallah Group, which hosts the Redbank deposits, consists of a lower volcanic and an upper sedimentary sequence. The base of the Group consists of approximately 150 m of mafic to intermediate volcanic rocks, overlain by a 150 m thick interflow dolostone, and capped by 180 m of mixed basaltic to trachytic flows and a thin upper porphyritic rhyolite (Roberts et al., 1963). The volcanic sequence is overlain by subaerial to lacustrine sandstones and siltstones. Hematized intermediate to silicic dikes, which are believed to be cogenetic with the volcanics, intrude the entire sequence.

Wernecke Mountains, Yukon Territory, Canada

In the Wernecke Mountains and southern Richardson Mountains in eastern and northeastern Yukon Territory, Canada, mineralized breccia bodies cut the Wernecke Supergroup, a thick succession of weakly metamorphosed sedimentary rocks of Middle Proterozoic age. Megascopically similar, but poorly mineralized, breccias are also recognized cutting the Wernecke Supergroup in the Ogilvie Mountains of central Yukon Territory (Lane, 1990). The Wernecke Supergroup has been subdivided into three groups (Bell and Delaney, 1977; Delaney, 1981). The oldest is the Fairchild Lake Group, which consists of an approximately 1.5 km thick section of metamudstone, metasiltstone and quartzite (Archer and Schmidt, 1978; Hitzman, unpubl. data). This Group forms an upward-shallowing sequence: deep-water clastic sedimentary rocks containing rare carbonate horizons, probably derived from an adjacent eastern carbonate shelf, are overlain by shallow marine terrigenous rocks and minor carbonate beds. The disconformably overlying Quartet Group consists of a uniform 3 km thick section of dark grey- to brown-weathering sandstone, siltstone and mudstone with very minor silty dolostone, and reflects an upward-shallowing sequence from starved basinal facies to an open, shallow marine environment (Bell, 1986a, b; Hitzman, unpubl. data). The Gillespie Lake Group conformably overlies the Quartet Group, and consists of over 1 km of stromatolitic dolostone, oolitic dolostone and parallel-laminated to wavy-bedded dolostone indicative of supratidal depositional conditions (Mustard et al., 1990). The only evidence of contemporaneous igneous activity are undated, volumetrically minor intermediate to mafic volcanic rocks, minor gabbro and diorite dikes and sills that cut the Wernecke Supergroup (Roots, 1990).

The age of the Wernecke Supergroup is poorly constrained. A minimum age of 1.5–1.2 Ga (Table 2) is given by the age of the breccia bodies that cut the lower portion of the sequence (Archer et al., 1977, 1986; Godwin et al., 1982; Parrish and Bell, 1987), and a Pb–Pb date of 1.28 Ga from galena in stratiform bodies within the Gillespie Lake Group (Morin, 1979).

Although the base of this mid-Proterozoic succession is not exposed, the thin-skinned style of fold and thrust belt deformation in the Wernecke Mountains suggests that the Wernecke Supergroup overlies an Early Proterozoic basement. The breccia bodies hosting iron ± copper mineralization are located along the Richardson Fault Array, a major north-trending fault zone that controlled block faulting and graben development from the mid-Proterozoic to the Tertiary. Although Archer and Schmidt (1978), Laznicka (1977a,b), Laznicka and Edwards (1979) and Laznicka and Gaboury (1988) view the breccia bodies in the Wernecke Mountains as discrete breccia zones that have undergone later deformation, Bell (1986a, b) considers the breccia bodies as portions of a major megabreccia complex including much of the central Wernecke Mountains which formed by dissolution of evaporites low in the Wernecke Supergroup and subsequent collapse.

Bayan Obo district, Inner Mongolia, China

In the Bayan Obo district of Inner Mongolia, China, Fe–REE mineralization is hosted by the mid-Proterozoic Bayan Obo Group, which overlies Archean and Lower Proterozoic granitic gneiss (Jishun et al., 1987). The Bayan Obo Group consists of a metamorphosed 1.8 km succession. It is comprised of a 750 m thick basal quartzite section containing minor carbonaceous shale and arkose which grades upward into an approximately 400 m thick section of locally stromatolitic dolostone and limestone which is in turn overlain by approximately 350 m of black slate with intercalated carbonate rock. Minor silicic volcanic rocks occur within the Proterozoic section (Li, 1983). The host rocks for the Bayan Obo deposits form roof pendants within a composite batholith of Permian age (Drew and Meng, 1990). The Bayan Obo district is located near the northern margin of the Sino-Korean plate, and the Bayan Obo Group is thought to have been deposited in an intracratonic rift at 1.7–1.85 Ga (Wang, 1985; Drew et al., 1989).

Province dimensions

Einaudi and Oreskes (1990) have noted that iron oxide deposits in the Stuart Shelf and southeastern Missouri provinces form linear arrays approximately 125 km in length and 40 km wide, with known deposits spaced 10–30 km along the trend. In the Wernecke and south Richardson Mountains breccia bodies are found in a linear array 250 km long and approximately 50 km wide. However, breccia bodies along this trend appear to be distributed irregularly (Hitzman, unpubl. data). In the Bergslagen district, the sixteen deposits in the Grängesberg area occur within a 30 × 5 kilometer belt. Similar spatial analyses for other provinces are made difficult by later deformation or lack of sufficient data.

Magmatic affiliation and age of mineralization

With the exception of several deposits in the Great Bear magmatic zone, the Swedish districts, and the southeast Missouri district, a direct genetic relationship between the iron oxide (Cu–U–REE–Au) deposits and a given magma type or magmatic episode remains to be demonstrated. However, many deposits show an indirect or spatial association with igneous rocks. Of these, the Great Bear magmatic zone and Kiruna districts show the greatest evidence of temporal overlap between mineralization and igneous host rocks.

Great Bear magmatic zone

In the southern portion of the Great Bear magmatic zone iron oxide mineralization generally occurs within rhyodacitic and rhyolitic volcanic rocks (Gandhi, 1989; Gandhi and Bell, 1990a). Mineralization is commonly spatially associated with quartz monzonite sills and laccoliths suggesting a genetic relationship (Gandhi and Bell, 1990a; R.T. Bell, pers. commun. 1990). In the Echo Bay–Camsell River region, iron oxide mineralization and associated alteration is spatially related to medium-grained, quartz monzonite–diorite sills and laccoliths, which were emplaced at 2 to 3 km depth beneath andesitic stratovolcanoes. Alteration occurs in the roof of one sill and extends as much as a kilometer from its margin. In another sill, alteration and mineralization occurs both within and above the intrusion. Hildebrand (1986) used the cross-cutting relationships between mineralization and igneous rocks to conclude that magmatism and mineralization were coeval at 1.87–1.86 Ga (Table 2).

Kiruna

At Kiruna and in surrounding regions, iron oxide deposits are hosted by intermediate to felsic volcanic rocks (Fig. 2). In the Kiruna district the volcanic sequence hosting the mineralization is approximately 6 km thick, and iron oxide mineralization has been recognized throughout the sequence. The Kiirunavaara and Luossavaara orebodies straddle the contact zone between a lower volcanic sequence composed of porphyritic, trachytic ash-flow tuffs and flows and related syenitic intrusive rocks, and an upper sequence dominated by rhyolitic ash-flow tuffs and tuffaceous sedimentary rocks (Geijer, 1930a, 1968; Offerberg, 1967; Eriksson and Hallgren, 1975; Frietsch, 1979b). The majority of the volcanic rocks in the district have been altered and the original chemistry of the host rocks in the district, although probably originally alkali-rich, is poorly known.

Various workers have proposed a close genetic association between the ores and the associated porphyries. Evidence pointing to a close temporal link between volcanism and ore deposition includes: (1) the Kiirunavaara orebody is cut by syenitic, rhyolitic and mafic dikes, (2) dikes and sills which cut the massive ore contain clasts of magnetite are themselves veined by magnetite, and (3) sedimentary rocks in the Hauki sequence contain clasts of iron oxide ore (Stutzer, 1907; Geijer, 1960; Geijer and Ödman, 1974; Frietsch, 1979b).

Iron ores in the other districts in the Kiruna area and the Bergslagen district of central Sweden occur in host rocks similar to those at Kiruna. At Grängesberg, the orebodies are cut by dikes which are compositionally similar to the host rocks (Magnusson, 1970; Frietsch, 1974; Lundstrom and Papunen, 1986). Recent work by Cliff et al. (1990) suggests a minimum mineralization age at Kiruna of 1.88 Ga derived from dating a granophyre dike which cuts ore. This age is comparable to the 1.9 Ga age of the host rocks (Skiold and Cliff, 1984). The Kiruna region has been affected by a secondary event at 1.54 Ga responsible for re-setting U–Pb and Rb–Sr systematics; this age corresponds to a later period of granitic intrusion in the region (Welin et al., 1971). Fission track dating of single crystal apatite from the Kiruna and Grängesberg districts by Koark et al. (1978) yielded dates of 486 ± 95 Ma for apatite from Tuolluvaara at Kiruna and 283 ± 73 Ma for apatite from Grängesberg. However, based on the geological evidence cited above that volcanism and mineralization are probably closely related, these Phanerozoic dates probably represent reset ages, related to later metamorphism.

Southeast Missouri

The St. Francois terrane in southeast Missouri is largely composed of silicic–alkalic igneous rocks containing very high SiO_2, K_2O/Na_2O, Fe/Mg and F, and low CaO, MgO and Al_2O_3, and characterized by alkali feldspar phenocrysts and iron-rich mafic minerals (Sims et al., 1987). The volcanic rocks consist of rhyolite ignimbrites and porphyritic flows with subsidiary, and generally later, basalt and porphyritic andesite flows and dikes. Comagmatic intrusions for the rhyolites include fine-grained granophyre and coarse-grained rapakivi granite. Comagmatic intrusions for the more mafic and intermediate rocks are a suite

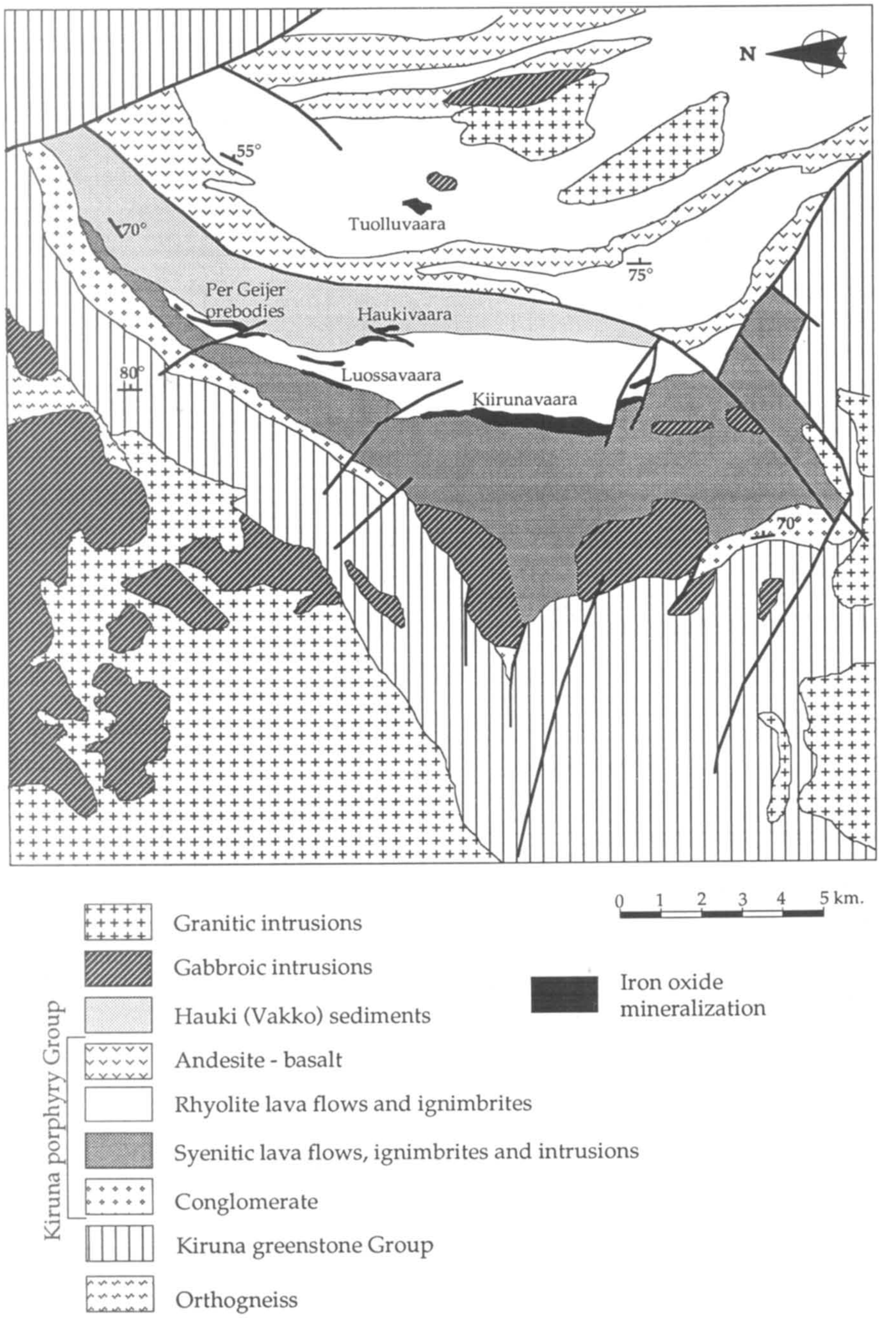

Fig. 2. Geological map of the Kiruna district, Sweden. Modified from Offerberg (1967), Eriksson and Hallgren (1975) and Forsell and Godin (1980).

of subvolcanic ring intrusions of trachyte, syenite and trachyandesite. High-silica, two-mica "tin granites" are the final igneous episode observed in the province. Iron ores are mostly restricted to the rhyolitic volcanic rocks, although at Iron Mountain ore occurs within an andesite flow (Murphy and Ohle, 1968), and at Boss-Bixby mineralization occurs primarily within syenite sills and to a lesser extent within adjacent volcanic rocks (Snyder, 1968;

Kisvarsanyi and Smith, 1988; Hagni and Brandom, 1990). A genetic relationship has been proposed by Kisvarsanyi (1981) between the iron ores and the later trachyte–syenite igneous suite. At Pea Ridge, a post ore dike cutting massive iron oxide yields a Rb–Sr date of 1.29 Ga (Snyder, 1969), which corresponds to a period of late or post-volcanism alteration which resulted in the resetting of Rb–Sr ages throughout southeastern Missouri (Bickford and Mose, 1975). The available evidence suggests that mineralization was roughly synchronous with volcanism.

Stuart shelf

On the Stuart Shelf, Australia, iron oxide deposits are found in a wide variety of granitic and volcanic rocks, including both deformed and undeformed granitoids. The spatial association of mineralization with igneous rocks has led several workers to suppose a genetic relationship between granitic magmatism and the ore-forming hydrothermal activity, e.g. Mortimer et al. (1988); Creaser (1989); Reeve et al. (1990). However, Oreskes and Einaudi (1990) noted that the coarse grain size of the Roxby Downs granite suggests a relatively deep level of emplacement, whereas the fragments of felsic porphyry and of mineralized sedimentary rocks in the host breccias suggest that the mineral deposit formed in a near-surface environment. On this basis, they suggested that the Olympic Dam deposit formed after uplift and unroofing of the batholith. Attempts at direct dating of mineralization at Olympic Dam indicate a middle Proterozoic age (Table 2), and suggest that mineralization may be as much as 200 million years younger than the enclosing granite. For example, Rb–Sr dates from sericite in altered granite wallrock yields an age of ~ 1.3 Ga (Gustafson and Compston, 1979); U–Pb dating of pitchblende in the ore bodies indicates multiple episodes of uranium mobility with a probable maximum age of 1.4 Ga (Trueman, 1986). These data contrast with an estimated age of 1.59 Ga for the host Roxby Downs Granite, and 1.57 Ga for associated Gawler Range Volcanics (Creaser, 1989). Hence, although Reeve et al. (1990) conclude that mineralization occurred some time between the emplacement of the host Roxby Downs granite (1.59 Ga) and the the end of Gawler Range volcanism (1.57 Ga), the available geologic and geochronologic evidence indicates that mineralization at Olympic Dam is significantly younger than the host granite. If there is a direct magmatic affiliation it lies with the minor, undated mafic and felsic dikes within the breccia complex or with unrecognized igneous rocks at depth (Oreskes and Einaudi, 1990).

Redbank

In the Redbank district, mineralized breccia pipes and hematized silicic igneous dikes cut volcanic host rocks, but no direct evidence of a temporal or genetic link between ore formation and magmatic activity has been described (Knutson et al., 1979).

Wernecke Mountains

In the Wernecke Mountains, the iron-oxide-rich breccias cut a sedimentary sequence, but also cut and are cut by mafic dikes and sills of unknown age (Hitzman, unpubl. data). Thus a weak correlation with local igneous activity may be recognized. The breccia bodies and associated mineralization appear to range in age from 1.2–1.5 Ga (Archer et al., 1977, 1986; Godwin et al., 1982; Parrish and Bell, 1987) although U–Pb analysis on uraniferous specimens yields dates up to Devonian (Archer et al., 1986).

Bayan Obo

At Bayan Obo, minor felsic tuffs and lavas (Li, 1983) have been recognized in the Proterozoic succession hosting the ores (Drew and Meng, 1990; Zeng, 1990), and attempts to date the mineralization have produced a wide variety of results. Philpotts et al. (1988, 1989) and Nakai et al. (1989) suggest that the iron oxide–monazite deposits formed at ~1.4 Ga, based on La–Ba dating of monazite and Sm–Nd isotope systematics of allanite–huanghoite–parisite. However, Chao et al. (1989) argue that 1.4 Ga is the age of the host dolomite, not of the mineralization, while Qiu (1990) believes that minor syngenetic mineralization occurred in the Proterozoic, but the major episodes of mineralization were Paleozoic. Chao and co-workers present K–Ar, Th–Pb and Ar–Ar dates on a wide variety of minerals which yield Precambrian to Ordovician ages (Chao et al., 1989; Chao, 1990) and conclude that mineralization began at approximately 800 Ma and culminated during the Caledonian orogeny at approximately 420–440 Ma. Still younger dates have been produced by La–Ba dating of allanite, which has yielded an age of 298 Ma (Nakai et al., 1989).

The large spread of ages, similar to the range observed in the Wernecke Mountains breccia bodies, suggests resetting of isotopic systematics in originally Middle Proterozoic mineralization.

Source regions, carbonatites, and banded iron formations

Meyer (1988a,b) has argued on geochemical grounds that a genetic connection exists between Proterozoic iron oxide (Cu–U–Au–REE) deposits and carbonatites, and/or crustal melts that included banded iron formation in their source regions. This view has been endorsed by Hauck (1990), who states that alkaline magmas are the ultimate source of the iron-rich fluids responsible for these deposits. Hauck (1990) argues that iron-rich fluids are derived from these magmas by liquid immiscibility; mineralization occurs as these iron-rich liquids differentiate by changes in the local chemistry (primarily oxygen fugacity) or physical environment (pressure and/or temperature) in the crust.

The evidence from the Great Bear magmatic zone, Kiruna, and southeast Missouri for emplacement of large volumes of alkali-rich plutonic and/or volcanic rocks of intermediate composition shortly before and during ore formation suggests that alkalic igneous activity was genetically linked with mineralization. However, the absence of significant igneous rocks in the Wernecke Mountains and at Bayan Obo indicates that the volcanic component of igneous activity is not essential to ore formation. If a plutonic link is essential as a source of heat, hydrothermal fluids, and/or metals for these deposits, then ore deposition may occur at structural levels significantly higher than those of the plutons.

Morphology, contact relations and textures of mineralization

The morphology of iron oxide–(Cu–U–REE–Au) ore bodies ranges from steep, pipe- or dike-like morphologies, commonly along faults or intrusive contacts, to tabular-concordant morphologies within stratified volcanic or sedimentary rocks. Morphologies are typically complex.

Discordant bodies and breccias

Structural control of ore bodies is particularly evident in the discordant breccia bodies in the Wernecke Mountains and at Olympic Dam. In the Wernecke Mountains, breccia bodies occur as dike-like or sill-like zones ranging from a few meters to more than 100 meters wide, or as elongate, bulbous bodies from 100 m to over 3 km in diameter. The vast majority of breccia bodies in the Wernecke Mountains appear to have formed along faults and/or anticlinal axes oriented either north-northwest, colinear with the major faults in the Richardson Fault Array, or east-northeast, in a conjugate orientation to the major structures.

The seven kilometer long Olympic Dam deposit consists of a number of coalescing, steeply-dipping breccia bodies, many of which are elongate and dike-like, and strike in NNW direction parallel to regional photo-lineaments. Oreskes and Einaudi (1990) interpret this overall morphology to reflect structural control by steep NNW-striking faults that post-date the emplacement and unroofing of the Olympic Dam granite. In both the Olympic Dam and the Wernecke breccia bodies, as well as at the Shepard Mountain veins in southeastern Missouri (Meyer, 1939), the iron-rich rocks display steep, wavy foliation textures, which may represent relict mylonitic fabrics.

A number of other deposits also display discordant morphologies, but without obvious structural control. In southeast Missouri, the Iron Mountain and Pea Ridge deposits display discordant morphologies with similarities to breccia pipes. At Iron Mountain, ore occurs in a dome-shaped shell with vertical limbs. In cross-section the deposit forms a near-circular, hollow, pipe-like body (Murphy and Ohle, 1968)—a pattern of mineralization similar to that seen in many breccia pipes associated with igneous stocks. At Pea Ridge, magnetite ore occurs as a discordant, vertical, dike-like mass, crescentic in plan view with an 800 m strike length, 100 m thickness, and a vertical extent in excess of 700 m (Snyder, 1968), which is in turn cut by late-stage discordant breccias. Einaudi and Oreskes (1990), noting the discordance between the magnetite ore body and the volcanic rocks as described by Snyder (1968), suggested that if the ore body formed in a vertical attitude, then it must have post-dated significant tilting of the enclosing volcanic rocks.

Concordant bodies

In contrast to the examples discussed above, many of the iron oxide deposits that occur in stratified volcanic and/or sedimentary rocks are tabular and concordant with their host rocks. The most striking example is the famous Kiirunavaara orebody which is 4 km long, extends at least 1.5 km downdip and averages 90 m in thickness. Similar morphologies are displayed by the Per Geijer and Haukivaara orebodies (Parak, 1975a), the Grängesberg deposits in central Sweden (Magnusson, 1970; Frietsch, 1974; Lundstrom and Papunen, 1986), the Pilot Knob surface and the Cedar Hill deposits in southeast Missouri, (Snyder, 1968, 1969), and the Bayan Obo deposit in China (Argall, 1980; Philpotts et al., 1989).

Massive and stockwork bodies

In addition to obviously discordant or concordant morphologies, many deposits form irregular bulbous masses or stockwork zones that have sometimes been referred to as "ore breccias." Mineralization of this style is particularly well developed in northern Sweden. "Ore breccias" occur beneath and lateral to the tabular-concordant Luossavaara ore body and are the dominant style of mineralization at the Tuolluvaara ore body in the Kiruna district (Parak, 1975a). In the Svappavaara district, the ore bodies consist of lenses or irregular masses of massive magnetite surrounded by ore breccia and disseminated magnetite in the host rocks (Eriksson and Hallgren, 1975; Lundberg and Smellie, 1979; Frietsch, 1980a, 1980b). Irregular morphologies are also common in the Great Bear magmatic zone and in the Acropolis deposit on the Stuart Shelf, where irregular bodies of massive magnetite are enclosed by a stockwork of magnetite ± actinolite ± apatite veins (Badham and Morton, 1976; Hildebrand, 1986; Paterson, 1986b; Gandhi, 1988, 1989; Gandhi and Bell, 1989; Gandhi and Lentz, 1990).

Contact relations

A distinctive and perhaps genetically important feature of these deposits is that they typically display gradational contacts with their host rocks over a scale of centimeters to meters. Sharp contacts are largely restricted to structurally-controlled zones. The majority of the massive iron oxide bodies such as those in northern and central Sweden, southeastern Missouri, on the Stuart Shelf and in the Great Bear magmatic zone display some degree of gradation between massive iron oxide and the enclosing wall-rocks. This gradation is expressed in a variety of ways, including increasing density of veins or abundance of iron oxide disseminations, or decreasing abundance and size of wall-rock fragments, toward the center of the deposits, and as noted above, by stockwork zones surrounding massive ores.

Ore textures

Ore textures are highly variable, both between and within deposits, and include massive iron oxide rocks, various types of breccias with an iron oxide matrix, and layered, foliated or stratified iron ore. In many cases original wall-rock textures are locally preserved within the ores, suggesting that a significant portion of the ore may have formed by replacement of the host rocks.

Evidence of replacement

Replacement of host-rocks has been most clearly demonstrated in deposits hosted by volcanic rocks. In a detailed petrographic study of the host rock at Pilot Knob, Missouri, Panno and Hood (1983) described gradational contacts between iron oxide mineralization and host-rock tuff, and observed relict glass shards and pumice fragments in the iron oxide ore. They suggested that the glassy matrix of the tuffs was preferentially replaced by early iron-rich fluids, and that further alteration and mineralization led to complete replacement of the volcanic rock by iron oxide. In the Kiruna district, magnetite textures in the outer portions of several of the massive bodies mimics that of the adjacent host volcanic rocks. Locally the magnetite bodies contain feldspar and quartz phenocrysts which decrease in abundance toward the center of the bodies (Geijer and Ödman, 1974).

In the "ore breccia" style of mineralization common to igneous-hosted deposits, evidence for wall-rock replacement includes relict igneous textures and remnant phenocrysts and clasts within massive magnetite, and the highly irregular, embayed edges of clasts in many of the "ore breccias," such as at Tuolluvaara (Geijer and Ödman, 1974; Parak, 1975a), at Painirova and Mertainen (Lundberg and Smellie, 1979), at Iron Mountain (Spurr, 1927; Singewald and Milton, 1929; Murphy and Ohle, 1968), and at Acropolis (Paterson, 1986b). Replacement textures also occur within the breccias at Olympic Dam. Around the margins of the deposit—outside the breccia bodies—sericitized feldspars within the host granite are dusted or rimmed by hematite, and cut by hematite veinlets. Within the breccia bodies, relict granite textures have been recognized underground, on the scale of 10s of meters, and at a microscopic scale, as relict zoned feldspar pseudomorphed by hematite (Oreskes and Einaudi, 1990).

Replacement textures have also been noted in deposits within sedimentary rocks. As in the "ore breccias" in igneous rocks, the edges of breccia fragments in the Wernecke Mountains are commonly highly embayed by the iron-oxides and alteration minerals that form the breccia matrix (Hitzman, unpubl. data). Iron oxide also extends into some clasts as a selective replacement of individual sedimentary laminae.

Sedimentary textures

Virtually all of these deposits contain rocks that display layered texture or foliation, and several contain hematitic rocks with distinct sedimentary textures. The clearest example of this occurs at Pilot Knob, where a large portion of the surface deposit consists of layered hematite displaying features such as graded bedding, fine laminations, scour and fill structures, and cross bedding (Anderson, 1976). Raindrop imprints and mudcracks are preserved on some bedding surfaces. The fine-grained, hematitic siltstones are interstratified with coarse conglomerates consisting of generally rounded, but locally angular, clasts of volcanic rocks, hematitically altered volcanic rocks, and hematite rocks in a hematite matrix. Rarely, some of the clasts display embayed rims suggestive of replacement by the matrix hematite. Meyer (1939) and Panno and Hood (1983) interpreted the textures of these rocks as evidence of selective replacement of matrix silts and of porous volcanic clasts by iron oxides. However, Anderson (1976) and Nold (1988) argued that these clasts were mineralized before incorporation into the sedimentary breccias, and thus favored syngenetic-exhalative deposition of hematite in a sedimentary environment. Parak (1975a, b, 1984) has utilized similar textural evidence to argue for a syngenetic-exhalative origin at Kiruna.

Almost identical hematitic siltstones are found at the Oak Dam prospect on the Stuart Shelf (Parker, 1987), as blocks within a breccia body at Pea Ridge (N. Oreskes, pers. observ., 1988), and as fragments within the heterolithic breccias at Olympic Dam (Roberts and Hudson, 1983; Oreskes and Einaudi, 1990). Indeed, the presence of hematitic siltstones within the breccias at Olympic Dam led to early interpretations of the entire deposit as a sediment-hosted system. However, underground exposures clearly reveal these rocks to occur as blocks and fragments within heterolithic, unstratified breccias. Invariably, the hematitic siltstones are found in the stratigraphically highest portions of deposits, or as clasts in breccia bodies that appear to have collapsed downward. This strongly suggests that they formed at or near the surface, either by erosion of exposed iron oxide deposits, by direct chemical precipitation from iron-rich fluids, and/or by infiltration of Fe-rich hydrothermal solutions into sediments. Conglomerates with an iron oxide matrix could have formed by selective replacement of matrix sediment, by deposition of clasts into a hematitic sediment, or by the infilling by fine iron oxide of an open matrix breccia.

Depth of ore formation

Mineralization in these deposits appears to have occurred from the paleosurface to depths of 4–6 km. Hematite-rich ores commonly display clear evidence of near-surface environments of formation, whereas evidence for the depth of formation of magnetite-dominated ores is generally more ambiguous. Much of the observed variation in mineralogy, wall-rock alteration, and ore textures may be related to depth of formation.

Kiruna

The relationship of mineralization to depth is best displayed in the Kiruna district where present-day exposures allow examination of a 6 km vertical cross-section (Fig. 3). At the greatest paleodepths, mineralization is dominated by magnetite, with minor amounts of chlorite, actinolite and/or albite, which forms disseminations, vesicle-fillings and hairline veinlets within the footwall syenite (Geijer, 1930a). This footwall mineralization extends for several kilometers laterally from the orebodies, and to the base of the footwall volcanic sequence at a depth of 5 kilometers beneath Kiirunavaara. The Kiirunavaara ore body itself consists primarily of massive magnetite with subsidiary intergrown apatite, actinolite and minor quartz. "Ore breccias" at the tops and sides of the massive ores may grade upwards through a zone of magnetite–apatite ± actinolite ± quartz veins into rhyolite with disseminated and veinlet magnetite–apatite ± actinolite.

The rocks above the rhyolite porphyry at Kiruna, forming what is believed to have been the top of the volcanic sequence at the time of iron

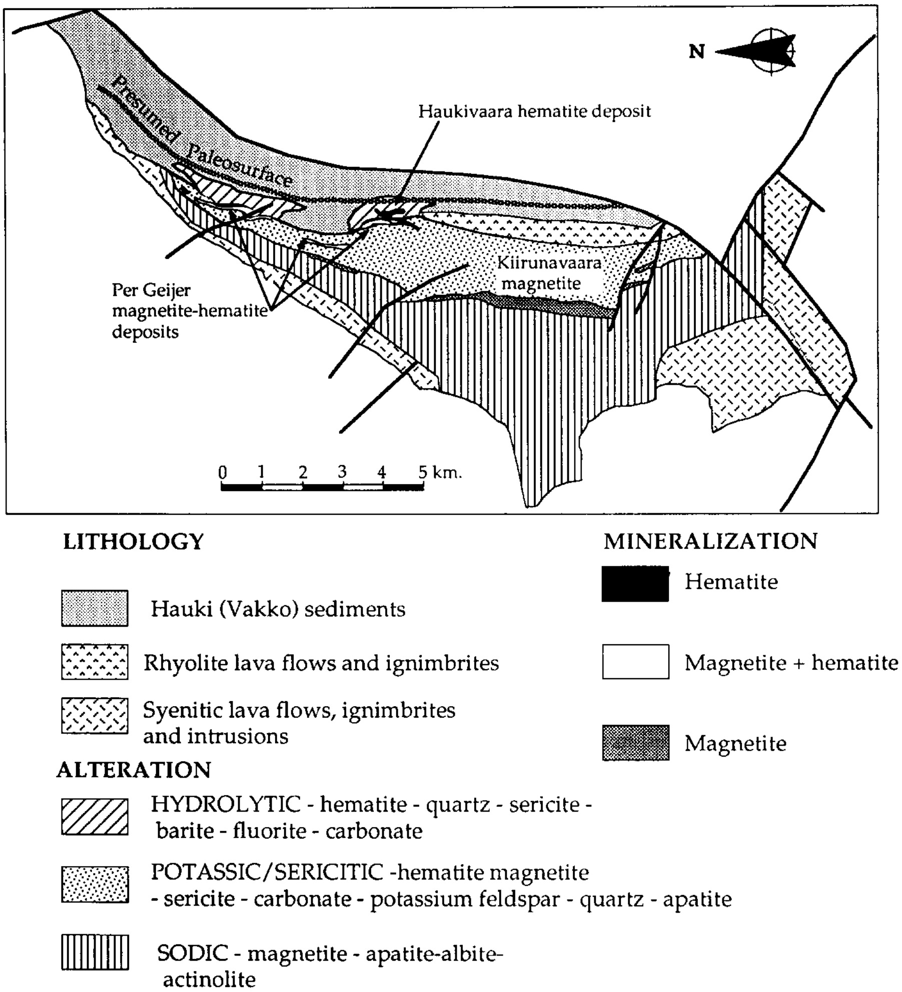

Fig. 3. Generalized cross-sectional reconstruction of the Kiruna district based on present surface geology and illustrating inferred broad alteration relationships. See Fig. 2 for lithologic units.

oxide mineralization, consists of a complex sequence of interlayered rhyolitic ash-flows and flows, volcaniclastic sediments and minor trachytic flows. The top of the volcanic sequence is occupied by the Haukivaara hematite ores (Per Geijer ores), which consist of magnetite and hematite, intergrown with quartz, sericite, apatite and barite. These rocks display distinct layering: layered iron oxide and apatite is common near the base of these ore bodies; quartz-banded iron oxide predominates near the top. Minor phases in the ore included barite, calcite, and fluorite. Both the hematite ores and adjacent altered rocks contain discontinous layers and pods of disseminated to massive barite and calcite. Weak sericitization, silicification, and calcitization continue above the level of the Haukivaara ores into the overlying Vakko sediments. Assuming the Haukivaara ores formed at the same time as the Kiirunavaara ores, these data suggest a minimum vertical extent of mineralization of 5–6 kilometers.

Shallow origin of hematite ores

The upward gradation in the Kiruna district from magnetite-dominated ores at depth to hematite-rich ores at higher levels is consistent with the evidence that deposits formed in a near-surface environment are dominated by hematite. For example, at Olympic Dam, hematite-rich breccias are exposed in drill core over a vertical extent of >1 kilometer (Fig. 4). The breccias contain fragments of hematitic siltstone, laminated barite, and rounded cobbles of volcanic porphyry that indicate incorporation of surficial sediments into the breccia complex. In particular, the laminated barite suggests venting of hydrothermal fluids into the surficial environment, or precipitation in large open spaces in a very shallow hydrothermal environment. This implies that the breccia complex formed at or near the paleosurface, and extended for a vertical extent of one kilometer or more. It is not known whether the hematite ore at Olympic Dam grades into magnetite-dominant assemblages at depth.

The hematitic siltstone clasts at Olympic Dam are visually indistinguishable from the hematitic siltstones exposed in the upper ore body at Pilot Knob, which have been interpreted as surficial deposits. (Anderson, 1976; Nold, 1988). However, Meyer (1939) argued, on the basis of ore textures and geological relations, for formation of these siltstones by replacement of pre-existing sedimentary rocks, albeit at depths no greater than 1000 meters or so. In either case, the evidence is consistent for formation of these hematite-dominated ores in very shallow environments, that is, within one kilometer of a paleosurface. By analogy, one might expect that other hematite-dominated systems formed in relatively shallow environments, but direct evidence is lacking.

Wall-rock alteration and mineralization

Iron oxide deposits typically display extensive wall-rock alteration. The type and intensity of this alteration, and associated mineral paragenesis, varies considerably both within and between districts, but appears to depend on the original wall-rock composition and the depth of formation of the deposit (Table 3).

Kiruna

The deposits in the Kiruna district exhibit a wide range of wall-rock alteration assemblages that are typical of the relations observed in the northern and central Swedish iron oxide districts, and appear to show a strong relationship with depth. Details of alteration and mineral paragenesis are derived from Geijer (1915, 1930a, 1960, 1968), Geijer and Magnusson (1952), Ekström (1973), Magnusson (1970), Geijer and Ödman (1974), Eriksson and Hallgren (1975), Parak (1975a, b, 1984), Frietsch (1978, 1979a, b, 1980a), and Forsell and Godin (1980, pers. commun., 1987), as well as from observations by one of the authors (MWH). The alteration at Kiruna shows a general transition from sodic alteration at deeper levels (albite-rich) to potassic alteration at intermediate levels (potassium feldspar +sericite), to sericitic and silicic alteration in the uppermost portion of the system (sericite + quartz). The description that follows progresses from deep to shallow.

In the footwall volcanics, the dominant alteration assemblage is magnetite–albite–actinolite–chlorite (Fig. 3). Plagioclase in the footwall volcanics throughout the district has largely been converted to albite. The density of magnetite $\pm$ albite $\pm$ actinolite veins increases abruptly immediately below the Kiirunavaara and Luossavaara orebodies. The contact of the Kiirunavaara ore body with the footwall volcanic rocks is marked along part of its length by a centimeter to half a meter thick amphibole layer that contains minor titanite (Geijer and Ödman, 1974); given the paucity of titanium in the rest of the deposit, this titanite may represent indigenous titanium mobilized during hydrothermal alteration of the host volcanic rocks. The orebody itself consists primarily of magnetite with intergrown apatite, actinolite, and minor quartz. Remnants of

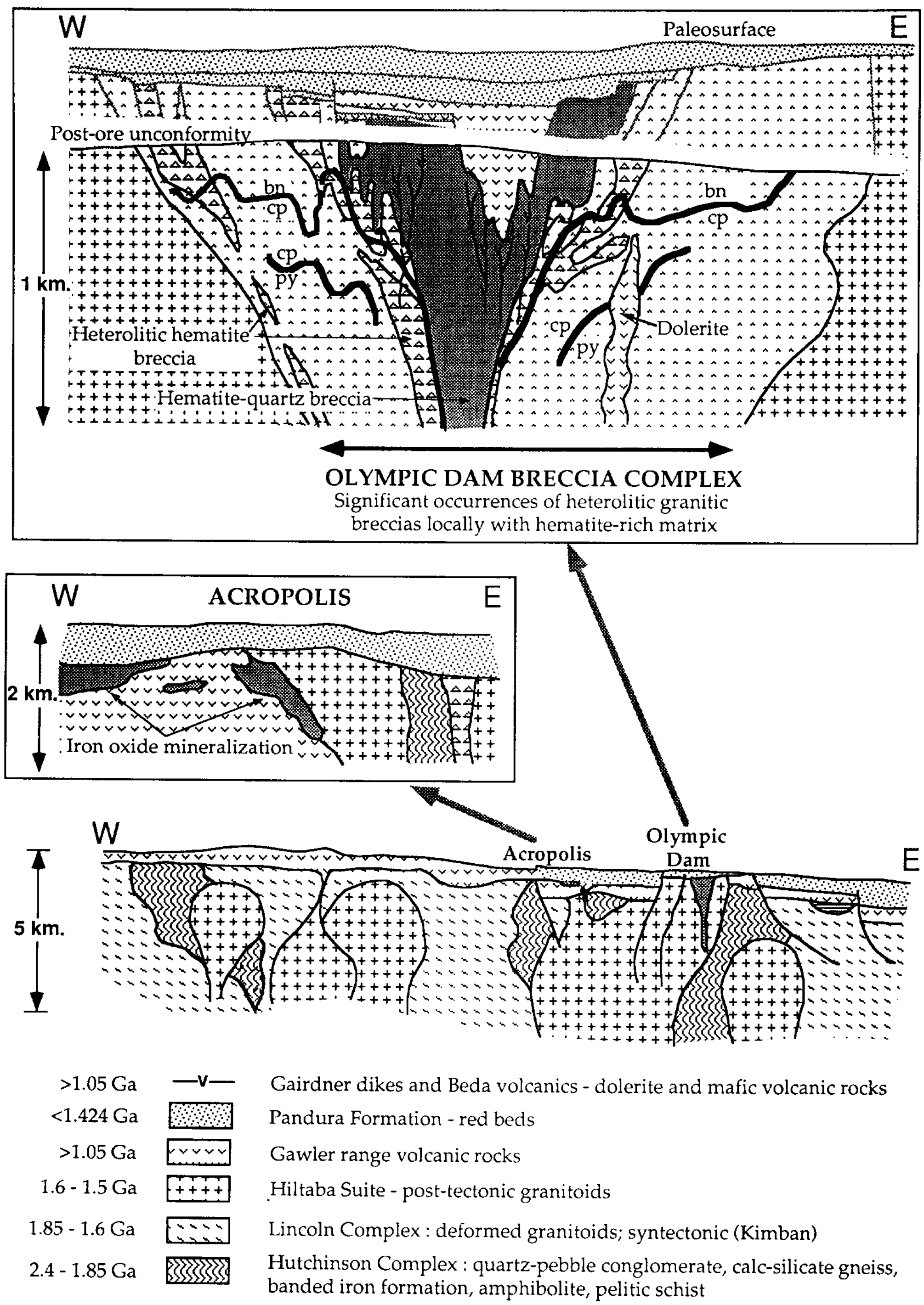

Fig. 4. Schematic summary of the geology of the Olympic Dam and Acropolis deposits and their regional setting on the Stuart Shelf. Based on Parker (1987), Oreskes and Einaudi (1990) and Reeve et al. (1990).

TABLE 3

Depth of mineralization and alteration assemblages

Region/District	Host Rocks	Depth from paleosurface	Alteration assemblage	Alteration type
South Australia Stuart Shelf	Granitic and felsic volcanic rocks	0–2 km Based on the presence of sedimentary and volcanic rocks as clasts in the breccia complex and the flaring out upward geometry of the breccia complex at Olympic Dam.	Hematite–sericite–barite–fluorite	Sericitic
Yukon, Canada Wernecke Mountains	Sandstone, siltstone, shale and minor carbonate rocks	Quartet Group breccias 1.5–2.5 km	Quartet breccias: Carbonate–chlorite–magnetite–hematite–sericite–barite–fluorite	Potassic/Sericitic
		Fairchild Lake Group breccias > 2.5 km Depth are based on stratigraphic reconstructions assuming mineralization occurred during or soon after Gillespie Lake Group deposition.	Fairchild breccias: Albite–paragonite–sericite–carbonate–magnetite	Sodic
Great Bear Magmatic Zone *N.W.T., Canada*	Silicic–intermediate volcanic and plutonic rocks	Echo Bay–Camsell River region – 2.5 to 3 km below paleosurface from stratigraphic reconstructions. Southern Great Bear – depth unknown, probably < 2.5 km	Echo Bay–Camsell River region: – Albite–magnetite–apatite–actinolite–chlorite Southern Great Bear: magnetite–hematite–epidote–chlorite–potassium enrichment	Sodic Potassic (?)
Inner Mongolia, China Bayan Obo District	Limestone, dolostone, sandstone, and shale	Depth from paleosurface unknown but regional geology suggests less than 3 km.	Above stratiform orebodies: potassium fieldspar–biotite–magnetite–fluorite–REE minerals–hematite Below stratiform orebodies: albite–magnetite–sodium amphibole–pyroxene	Potassic Sodic/Calcic "Skarn"
Northern Sweden Kiruna District	Intermediate and felsic volcanic rocks.	Depths based on stratigraphic reconstruction: Haukivaara ores: 0–250 m	Hematite–quartz–sericite–barite–fluorite–carbonate	Hydrolytic/Silicic
		Per Geijer ores: 250 m–1.5 km	Hematite–magnetite–sericite–carbonate–potassium feldspar–quartz–apatite	Potassic/Sericitic
		Kirunavaara ores: 2–6 km	Magnetite–apatite–actinolite–albite	Sodic
Southeastern Missouri	Felsic intermediate volcanic rocks.	From Paleosurface (Pilot Knob – surface deposit) to 1–2 km (?) depth.	Pilot Knob (surface): Hematite–quartz–carbonate Other deposits: magnetite–hematite–potassium feldspar–sericite–actinolite	Hydrolytic/Silicic Potassic/Sodic

trachyte and rhyolite enclosed within the ore are commonly converted to albitite. Large wall-rock fragments may display albitite rims on less altered interiors (Geijer and Ödman, 1974).

Post-ore alteration is also recognized at Kiirunavaara. The orebody is cut by porphyritic dikes that display potassic alteration with secondary potassium feldspar growth, sericitization of plagioclase, and disseminated hematite. Hematite $\pm$ quartz, barite, and/or fluorite veins locally cut the ore, and late high-angle faults may be filled with quartz and pyrite.

Towards the top of the hanging-wall rhyolite, beneath the Per Geijer ores, igneous potassium feldspar crystals are partially converted to quartz + biotite, locally associated with disseminated pyrite. The upper and lower contacts of the Per Geijer ore bodies are locally occupied by breccias consisting of silicified and/or sericitized volcanic rocks in a hematite, magnetite, and/or apatite matrix. Locally the footwall may contain abundant apatite veins, up to 2 meters wide, that narrow and neck downwards. Rhyolite porphyry adjacent to these veins may contain abundant disseminated apatite and zircon, as well as numerous quartz–hematite veins radiating from the apatite veins.

The rocks that host the Per Geijer ores are commonly intensely altered so that original rock textures may be difficult to distinguish. Red-orange, secondary potassium feldspar replaces original potassium feldspar and forms irregular pods and veins; sericite appears to replace groundmass and plagioclase phenocrysts. Sericitization and silicification increase upwards towards the top of the sequence, which is occupied by the Haukivaara ores. Hematite in these orebodies is intergrown with quartz, sericite, apatite, and barite. Both the hematite ores and the adjacent altered rocks contain discontinuous layers and pods of disseminated to massive barite and calcite. Weak sericitization, silicification, and calcitization continue above the Haukivaara ores into the overlying sedimentary rocks. These relationships suggest that hematite-rich alteration, associated with sericite and silica, is both a high-level and a late-stage alteration assemblage.

Southeast Missouri

The style of wall-rock alteration in the southeast Missouri deposits is a function of wall-rock composition, timing, and apparent paleodepth of exposures. Wall-rock alteration associated with magnetite bodies in felsic igneous rocks, such as a Pilot Knob (Meyer, 1939) and Pea Ridge (Emery, 1968) is dominated by potassic assemblages that include potassium feldspar, quartz, and sericite. In mafic/intermediate volcanic rocks, such as at Iron Mountain (Murphy and Ohle, 1968), magnetite–hematite bodies and veins contain more calcic assemblages that include andradite and actinolite, and wall-rock alterations minerals consist of epidote and chlorite. The actinolite–magnetite association of Pea Ridge is thought to represent alteration of mafic dike fragments (Emery, 1968).

In those deposits where a temporal evolution of alteration–mineralization has been documented, hematization, silicification, and sericitization are late events. For example, at Pea Ridge, massive actinolite and magnetite bodies in brecciated rhyolite are cut by breccias containing hydrothermal potassium feldspar and biotite (Nuelle et al., 1989). These are cut in turn by late breccias accompanied by silicification and sericitization of the host rhyolite, chloritization of actinolite, oxidation of magnetite to hematite, and deposition of quartz, barite, calcite, and minor Cu, REE, and Au mineralization (Day et al., 1989; Marikos et al., 1989; Seeger et al., 1989).

At the Pilot Knob surface deposit, which probably represents the shallowest paleodepths recognized in the southeast Missouri district, hematite is the dominant iron oxide mineral, hydrothermal feldspar appears to be rare or absent, and wall-rock alteration consists of silicification and sericitization. Thus, considering all of the deposits in the district, there is the suggestion of a transition in space and time from early and deep magnetite accompanied by potassium feldspar or actinolite, to late and/or shallow hematite accompanied by quartz, sericite, and lesser calcite and barite, a pattern similar to that observed in the Kiruna district.

Great Bear magmatic zone

In the Echo Bay–Camsell River region, altered areas are centered on a number of quartz monzonite to monzonite plutons. These altered areas are zoned and consist of an inner (deeper) zone of intensely albitized rock, an intermediate zone of magnetite-apatite–actinolite veins and disseminations, and an outer (shallower) pyritic zone (Hildebrand, 1986; Reardon, 1989, 1990). Alteration associated with iron oxide mineralization may extend over 1 kilometer from the plutons. Toward the roof of the pluton, plagioclase is commonly replaced by albite and quartz, grading into a zone of intense albitization in the uppermost portions of the pluton and in adjacent host andesitic lavas and sedimentary rocks. In the zone of intense albitization, original textures are largely destroyed and in situ brecciation is common. Outward from the pluton, albitized host rock contains disseminated magnetite, and is cut by veins, stockwork zones, and massive pods of magnetite–apatite–actinolite that are similar to the "ore breccias" of the Kiruna district. Most common are small (1–2 cm wide) veins of pink apatite with disseminated octahedra of magnetite or martite and fibrous green amphibole selvages. Elsewhere, veins contain only amphibole or amphibole–apatite and lack magnetite. Rarely, massive replacement of the sedimentary country rock, particularly of limestone, has resulted in the formation of layered magnetite–apatite–actinolite. The outer sulfide zone displays weak albitization of the host rocks with abundant disseminated to semimassive pyrite.

Wall-rock alteration in the Echo Bay–Camsell River region is predominantly sodic and probably formed at paleodepths of 2–3 km. It is quite similar to that in volcanic rocks below the Kiirunavaara orebody, formed at paleodepths on the order of 5–6 km. The distal actinolite-rich alteration in the Great Bear magmatic zone is comparable to alteration assemblages at intermediate depths in the Kiruna district and in most of the magnetite-dominated deposits of southeast Missouri.

The iron oxide (Cu–U) deposits in the southern portion of the Great Bear magmatic zone occur primarily in rhyodacite ignimbrites. Preliminary analyses suggest minor enrichment in potassium over sodium adjacent to mineralized zones (Gandhi, pers. commun., 1991). Clasts within ore are commonly altered largely to epidote and minor epidote and chlorite are also present adjacent to some areas of mineralization. However, wall-rock alteration appears to be less intense in the southern portion than in the northern portion of the Great Bear magmatic zone. These differences may reflect contrasting depths of formation of the two deposit types with the plutonic-associated magnetite-apatite deposits being formed at a greater depth than the volcanic hosted magnetite–hematite (Cu–U) deposits.

Olympic Dam

The alteration at Olympic Dam is dominated by hematite- and sericite-rich assemblages, with locally intense silicification. Within one kilometer of the mineralized breccia complex at Olympic Dam, the host Roxby Downs granite is pervasively fractured, weakly sericitized, and locally weakly hematized. With increasing proximity to the breccias, sericitic alteration becomes more intense: plagioclase is completely converted to sericite and potassium feldspar is rimmed by sericite. Mafic clots are converted to chlorite and epidote ± hematite. Close to the deposit, hematization becomes more abundant and appears to overprint early sericitization: adjacent to hematite breccia bodies, sericitized feldspars are dusted with, or rimmed by, hematite; within hematite breccia bodies, feldspar and sericite are totally replaced by hematite. Hematite in the massive zones at the center of the deposit may pseudomorph plagioclase, carbonate, and magnetite. No hydrothermal feldspar has been recognized to date in any of the breccias. Intense hematization in the center of the deposit may be accompanied or overprinted by locally intense silicification.

Oreskes and Einaudi (1990) have elucidated a complex set of veining events which largely overlap the period of brecciation. Based on their work and that of Reeve et al. (1990), a paragenetic sequence of mineralization and alteration has been constructed for the deposit. The earliest stage recognized in the deposit consists of magnetite–pyrite–siderite veins preserved on the eastern edge of the deposit, associated with chlorite–carbonate alteration and weak brecciation of the host granite. The magnetite-bearing assemblage is followed by intense brecciation, sericitization, and hematization of granite adjacent to the breccias, and precipitation of hematite, uraninite, fluorite, and REE minerals. Sericitic alteration is pervasive, and granitic feldspars are entirely converted to sericite. Copper–iron sulfides, together with abundant sericite, formed late relative to intense brecciation and were accompanied by continued precipitation of hematite, fluorite, and uraninite. Continued hematite formation in the central portion of the deposit produced nearly massive hematite zones. Late silicification locally overprinted the hematite breccias. Late barite veins cut the hematite breccias in the core of the deposit. Thus, although a spatial relation of alteration to depth is not evident at Olympic Dam, a temporal relation is: early magnetite is associated with chlorite–carbonate alteration, whereas later hematite is associated with sericitization, and locally, with silicification.

Wernecke Mountains

Vertical exposures of individual breccia bodies in the Wernecke Mountains are on the order of several hundred meters. However, breccia bodies are found throughout approximately 4.5 km of section, which permits examination of different styles of alteration and mineralization through a large vertical interval. As in the Kiruna district, alteration types and zoning associated with the breccia bodies appear to be a function of depth: sodic alteration associated with magnetite is more abundant at depth; potassic alteration associated with hematite is more abundant at higher levels and in distal fringes. Sericite and carbonate alteration are mostly restricted to high stratigraphic levels (Hitzman, unpubl. data).

Breccia bodies cutting the lower Fairchild Lake Group rocks, such as those at Dolores Creek (Fig. 5), commonly display zones of intense sodic alteration characterized by albitization, and wall-rock bleaching which extends for several kilometers from the breccia contact (Laznicka and Edwards, 1979). Bleaching results from destruction of disseminated carbonaceous material and iron oxides in the sedimentary host rock, and its distribution and intensity is controlled by permeability along structures and stratigraphic horizons. In the overlying Quartet Group, breccia bodies such as Igor (Fig. 5) generally lack intense albitization, but contain zones of sericitic and carbonate alteration and iron metasomatism. Bleaching of the wall-rock is less extensive than in the lower units, extending for tens to several hundreds of meters from the breccia contact.

Sodic alteration is best developed in breccias found in the lower Fairchild Lake Group. These breccia bodies generally contain albite as both a major matrix component and as replacement of sedimentary rock fragments. Intense alteration formed massive albitite zones as veins within the breccia parallel to the long axis of the breccia body. Iron oxide occurs as both magnetite and hematite, although much

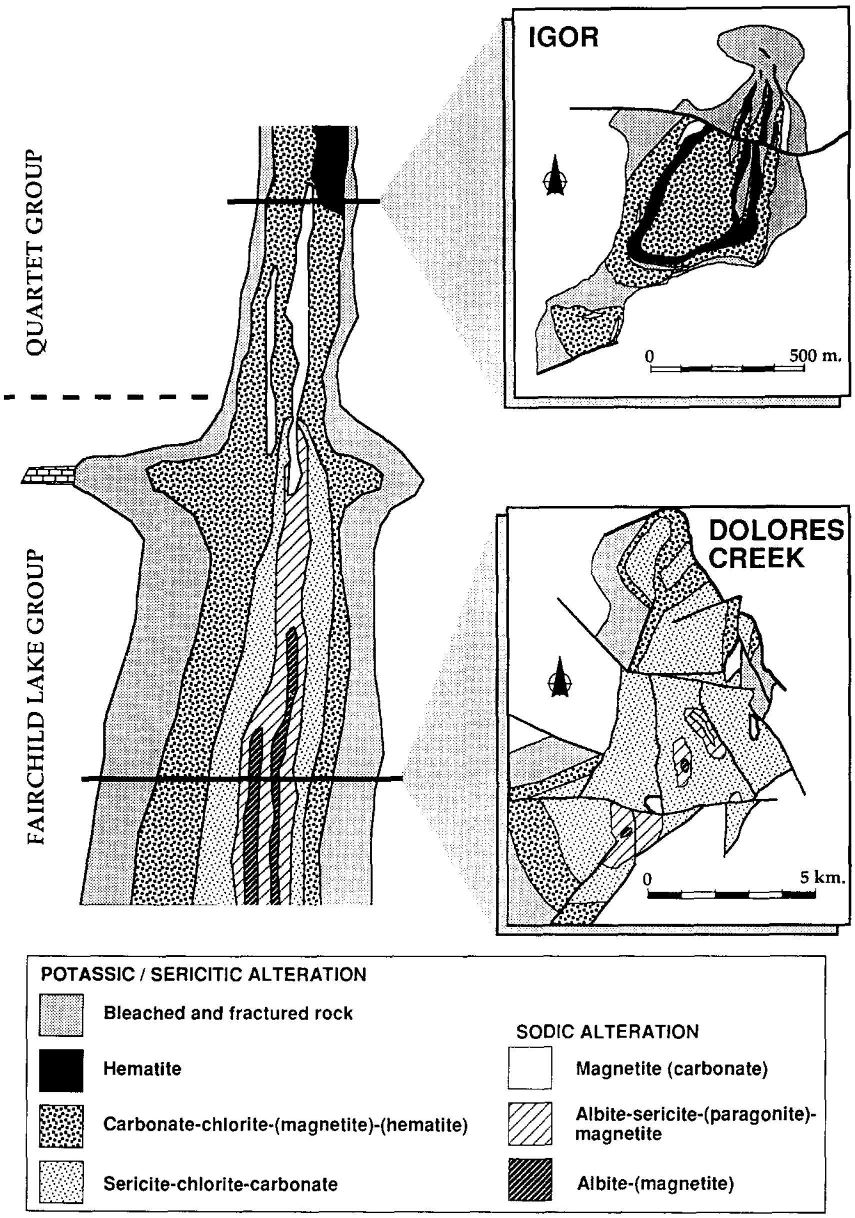

Fig. 5. Schematic summary of the geology of the Wernecke breccias showing relative stratigraphic position of the Igor and Dolores Creek prospects. Geologic maps of Igor from Hitzman (unpubl. data) and Dolores Creek highly modified from Bell (1986a, b), Laznika (1977a, b) and Laznika and Edwards (1979).

of the hematite may be an alteration product of magnetite. Albitite zones are surrounded by less intense sodic alteration consisting of albite–paragonite–sericite–martite, which occurs within and peripheral to breccia bodies and commonly contains between 5–8 wt% Na_2O (Laznicka and Edwards, 1979). Albite is best developed in quartzite and metasiltstone, whereas paragonite is dominant in argillaceous metasiltstone and metamudstone. Limestone and dolostone lenses adjacent to the breccia bodies contain scapolite, chlorite and actinolite. At the fringe of the sodic alteration zone, there is a poorly developed potassic alteration zone characterized by abundant sericite and lesser biotite, which contains 5–9 wt% K_2O, compared to average K_2O values of 1–5 wt% K_2O in unaltered sedimentary host rocks (Laznicka and Edwards, 1979; Hitzman, unpubl. data). This potassic alteration zone is enveloped by a very broad region, up to 2 km wide, displaying weak sericitization of feldspar, chloritization of mafic minerals, and disseminated carbonate and hematite.

The occurrence of albite veins that cut rocks with albite–paragonite–sericite alteration assemblages, and of quartz–albite veins that cut potassic alteration assemblages, suggests that sodic and potassic alteration were part of a continuous hydrothermal event that expanded outward from the center of the breccias. This alteration was roughly synchronous with, or slightly post-dated, brecciation. Prior to this alteration event the majority of the breccia had been altered to quartz–sericite–chlorite assemblages, in both matrix and clasts. However, the presence of albitite clasts in a quartz–sericite–chlorite matrix breccia (Bell, 1986b) demonstrates multiple brecciation and alteration events.

In breccia bodies cutting the stratigraphically higher Quartet Group hematite is the dominant matrix iron oxide (Hitzman, unpubl. data). Alteration affects breccia matrix and some clasts; significant alteration rarely extends into the surrounding host rocks. Early brecciation produced a quartz–sericite matrix with weak quartz–sericite alteration of clasts. Later brecciation and alteration resulted in conversion of the matrix to a chlorite-carbonate (dolomite, calcite)–sericite–quartz ± magnetite assemblage. With increasing alteration, clasts were partially or wholly converted to a similar assemblage. Chlorite–carbonate alteration was followed by massive carbonate–magnetite ± albite ± quartz, which commonly resulted in the obliteration of original breccia textures. Minor barite, fluorite, and apatite are common accessories; pseudomorphs of actinolite by chlorite–quartz are locally abundant. Zones of carbonate–magnetite ± sulfide alteration form elongate vein-like structures parallel to the long axis of the breccia bodies. The carbonate–chlorite and carbonate–magnetite assemblages are cut by hematite-rich breccias. These breccias, which commonly display steep layered or laminate textures parallel to breccia margins, contain a carbonate–quartz–hematite assemblage. Minor constituents include barite, fluorite, chlorite, apatite and sulfides. Where hematite breccias cut or are adjacent to carbonate–magnetite bodies, the magnetite has been martized. Late carbonate–sulfide ± barite veins cut all the breccia units. These relationships indicate that hematite generally formed at later stages than magnetite in the hydrothermal system.

Redbank

In the Redbank district, volcanic and sedimentary rocks on the margins of hematite–K-feldspar–carbonate breccia bodies contain hydrothermal K-feldspar. Over a lateral distance of 30 m beyond the potassically altered wallrocks, veins in the host rocks are zoned outwards from hematite–quartz–chlorite to ankerite–siderite–hematite (Knutson et al., 1979). The age relationships between these various alteration–mineralization assemblages are unknown.

Bayan Obo

The Bayan Obo deposits form a series of stratabound lenses at the contact between a dolostone unit and an overlying slate, and alteration shows a broad pattern dominated by sodic assemblages in the footwall, potassic assemblages in the hanging wall. Below the main ore zones, sodic alteration and weak iron oxide–REE mineralization, consisting of albite–sodic amphibole ± sodic pyroxene veins with minor magnetite and monazite, extends into the gneissic basement and covers an area of approximately 18 × 2 km. The ore bodies appear to be internally zoned: a core of massive magnetite grades laterally into banded magnetite–hematite–monazite–bastnaesite–fluorite. Banded ore is laterally equivalent to dolostone containing disseminated magnetite, monazite, bastnaesite, and fluorite (R. Erikson, pers. commun., 1988). Stratabound ore is overlain by a zone of brecciated hanging-wall slate, in which slate clasts have been largely converted to microcrystalline microcline. The breccia matrix is a mineralogically complex mixture of potassium feldspar, biotite, fluorite, albite, hematite, magnetite, calcite, bastnaesite, monazite, apatite, barite, niobates, pyrochlore, pyrite, and iron carbonate. Hanging-wall slate above the mineralized breccia also shows potassic alteration, and contains secondary potassium feldspar and biotite for an unspecified distance above the orebody (Drew et al, 1989). Chao (1990) and Chao et al. (1989) have proposed a preliminary paragenetic sequence dominated by early monazite and arfedsonite, intermediate magnetite, and late hematite mineralization.

The alteration zonation as currently described is similar to that associated with the Kiirunavaara orebody. The iron oxide orebodies are underlain by rocks showing sodic alteration and are overlain by rocks showing potassic alteration.

Cu–U–Au–REE mineralization

The metal suite Fe–Cu–U–Au–REE is one of the distinguishing characteristics of this class of deposits. Although individual deposits always contain some copper, uranium, gold and/or rare earth elements (particularly LREE) they may not be present in economically significant concentrations. Fluorine, barium, and phosphorus are also enriched. All the deposits contain abundant iron oxide as either magnetite and/or hematite (Table 1). However, unlike many other iron oxide deposits, the level of contained titanium is low, generally below 0.5% TiO_2 and rarely above 2% TiO_2. The low levels of titanium help to differentiate these deposits from otherwise somewhat similar occurrences closely associated with anorthosites, gabbros, and layered mafic intrusions.

Copper–iron sulfides

While iron and copper sulfides are found in both hematite- and magnetite-rich bodies, they appear to be most common in hematitic breccia. Of the hematite-rich deposits, the Olympic Dam deposit is the most significant as a copper deposit; it contains an estimated resource of over 2 billion metric tons with an average grade of 1.6% Cu (Roberts and Hudson, 1983). Significant tonnages of over 1% Cu are also present in hematite breccia bodies in the Wernecke Mountains (Hitzman, unpubl. data) and in the Redbank District (Orridge and Mason, 1975; Knutson et al, 1979). Minor copper sulfide mineralization has been recognized in the immediate footwall of the Kiirunavaara orebody and in hematitic Per Geijer ores in the Rektorn orebody (Hitzman, per. observ., 1987). Chalcopyrite is the dominant copper sulfide mineral, but primary bornite and chalcocite occurs at Olympic Dam (Roberts and Hudson, 1983; Oreskes and Einaudi, 1990) and in the Wernecke breccias (Hitzman, unpubl. data).

Olympic Dam is exceptional in containing large amounts of chalcocite in addition to bornite and chalcopyrite. Although much of this chalcocite is intergrown with bornite and appears to be part of the primary sulfide assemblage, a significant portion occurs with little or no bornite. This very high grade zone occurs beneath a 5 to 250 m thick barren zone below the Late Proterozoic unconformity and transects the zoning of deeper pyrite–chalcopyrite–bornite–chalcocite mineralization. The overall zonation pattern is reminiscent of supergene profiles developed in porphyry copper deposits and suggests that the unusual abundance of chalcocite and very high copper grades at Olympic Dam may in part reflect secondary alteration processes that occurred when the deposit was exposed and weathered at the surface during mid-Late Proterozoic time (Oreskes and Einaudi, 1990).

In many of the deposits investigated to date, magnetite + pyrite is an early assemblage, and copper–iron sulfides either accompany or postdate later hematite. At Olympic Dam the majority of sulfides occur as interstitial grains within the breccia matrix or as microveinlets cutting hematite, suggesting that most of the sulfide precipitation accompanied or postdated hematite formation (Oreskes and Einaudi, 1990). Similarly, at Pea Ridge, early magnetite + pyrite is cut by later chalcopyrite and even later hematite (Nuelle et al., 1989). In the Wernecke Mountains, sulfide mineralization appears to be a late-stage event, and several styles of copper mineralization are observed. In breccias cutting the lower portions of the Wernecke Supergroup, chalcopyrite replaces early magnetite and is intergrown with, or replaces, pyrite, magnetite, martite, and specular hematite. In breccia bodies in the stratigraphically higher Quartet Group, copper mineralization consists of disseminated and veinlet-controlled chalcopyrite, commonly intergrown with pyrite. The sulfides occur either as replacements of magnetite or as interstitial grains within carbonate or specular hematite.

The distribution and occurrence of sulfides in magnetite-rich systems has not been well documented, but several magnetite bodies are known to contain copper sulfides. Iron oxide deposits in the southern portion of the Great Bear magmatic zone contain trace to significant copper sulfides (Gandhi and Bell, 1990a; Reardon, 1990; R.T. Bell, pers. commun., 1990). The Sue-Diane prospect contains geologic reserves of 8 million tons of 0.8% Cu (Gandhi, 1989; Gandhi and Bell, 1990a). The Acropolis prospect on the Stuart Shelf contains copper sulfides in massive magnetite bodies and stockworks; one drill hole intersected 66 m of 0.7% Cu (Paterson, 1986b). In northern and central Sweden, minor copper sulfide mineralization has been recognized in several magnetite bodies (Grip, 1979; Frietsch, 1980d; Parak, 1985; N. Badham, pers. commun., 1989). In southeastern Missouri, Boss-Bixby is the only deposit that contains a substantial tonnage of low grade copper mineralization (Kisvarsanyi and Proctor, 1967; Kisvarsanyi and Smith, 1988). Chalcopyrite is generally the sole primary copper sulfide in these magnetite-rich deposits, although Boss-Bixby contains minor bornite, covellite, and carrollite (Hagni and Brandom, 1990).

Iron sulfide, dominantly pyrite, is also common in many of the iron oxide deposits, especially those that are magnetite-rich. At Kiirunavaara disseminated pyrite and lesser chalcopyrite are found in massive magnetite within several meters of the base of the deposits, and in magnetite and magnetite–actinolite veins in the immediate footwall; sulfides appear to cut magnetite (Hitzman, per. observ., 1987). Pyrite is also found in late quartz veins that cross-cut massive magnetite. In southeast Missouri, pyrite is found in quartz–chlorite–magnetite veins cutting quartz–hematite at Shepard Mountain (Singewald and Milton, 1929; Meyer, 1939) and as locally abundant disseminations in high-grade magnetite ore at Pea Ridge. In the Echo Bay–Camsell River region of the Great Bear magmatic zone, pyrite appears to be disseminated in the magnetite–actinolite mineralization and widely distributed as disseminations and pods within the broader alteration halo to the iron oxide mineralization (Hildebrand, 1986; Reardon, 1989, 1990). Minor pyrite, chalcopyrite, and sphalerite are present in the Bayan Obo deposit (L. Drew, pers. commun., 1990). At Olympic Dam, pyrite is intergrown with magnetite in an early magnetite–siderite assemblage (Oreskes and Einaudi, 1990).

Uranium

Trace to economic grade uranium mineralization has been recognized at a number of the iron oxide deposits. However, there is sparse detailed mineralogical and petrographic data on its occurrence. At Olympic Dam uranium occurs as uraninite inclusions in hematite, as intergrowths of pitchblende with copper–iron sulfides, and as disseminated brannerite (Roberts and Hudson, 1983; Oreskes, 1990a, b; Reeve et al, 1990). Trace to subeconomic concentrations of uranium have been found in a number of other deposits on the Stuart Shelf (Paterson, 1986b; Parker, 1987), but details of mineralogy and paragenesis have not been published. In the Wernecke Mountains, uranium is ubiquitous in the breccias (Morin, 1977; Archer and Schmidt, 1978). The principal uranium minerals are brannerite and pitchblende. Uranium is found in a spectrum of alteration types and occurs with both magnetite- and hematite-rich mineralization. The highest concentrations occur in stratigraphically deep, albitized breccia and in quartz veins peripheral to the breccia bodies (Hitzman, unpubl. data).

In other iron oxide deposits, uranium has been observed in trace amounts. In southeast Missouri, no significant uranium mineralization has been described in the literature. However, radioactive concentrations have been located in hematitic breccia bodies which cut the Pea Ridge magnetite body (M. Marikos, pers. commun., 1988) and in apatite concentrates from the mine (L. Smith, pers. commun., 1988). Soluble uranium has been found in waste rock from the Pilot Knob mine (S.A. Hauck, pers. commun., 1990). There are virtually no chemical data for uranium in the Swedish deposits, although trace uraninite has been noted in the Haukivaara hematite deposit (Parak, 1975a). In the Great Bear magmatic zone, uranium is associated with copper mineralization as disseminations of pitchblende or uraninite in magnetite and as secondary concentrations of pitchblende in fractures (Gandhi, 1988, 1989; Cannuli et al., 1990; Sawiuk et al., 1990).

Rare earth elements

Rare earth elements (REE) are present in anomalous concentrations in all of the iron oxide deposits for which analytical data are available. The most spectacular example is the Bayan Obo deposit, in which the three main ore bodies contain 20 million tons of 35% Fe and 6.19% rare earth oxides (REO), 15 million tons of 33% Fe and 5.71% REO, and approximately 10 million tons of 31% Fe and 1% REO, respectively (Argall, 1980). Mineralization occurs in two habits: lenses of semimassive to massive magnetite (± hematite) with interlayered monazite, fluorite, and bastnaesite that replace dolostone; and cross-cutting veins and breccias containing REE minerals, primarily monazite and bastnaesite (Chao, 1990). Cross-cutting relationships suggest that REE mineralization pre-dated, was synchronous with, and post-dated iron mineralization. The wide variety of rare earth element minerals present in the Bayan Obo deposits, where twenty-eight REE-bearing minerals have been recognized (Qui, 1990; Zeng, 1990), may be due to post-mineralization metamorphism.

At Olympic Dam, as at Bayan Obo, REE mineralization continued throughout the entire alteration and mineralization sequence. Hematite breccias at Olympic Dam contain an average of approximately 5000 ppm REE, with the highest concentrations found in the most hematitic breccias in the core of the deposit (Oreskes and Einaudi, 1990). Four major REE-bearing phases have been identified: bastnaesite, florencite, monazite and xenotime. Bastnaesite and monazite are the most abundant: they occur in early quartz-sericite veins, disseminated in the quartz-sericite matrix of the breccias, and as intergrowths with hematite; bastnaesite also occurs as inclusions in bornite. Elsewhere on the Stuart Shelf, REEs are weakly concentrated in apatite that occurs intergrown with magnetite in the Acropolis deposit (Paterson, 1986a, b).

In the Wernecke Mountains and south Richardson Mountains, REE concentrations in the breccia bodies are greater than twice the normal REE content of the host sedimentary rocks. Progressive REE enrichment is noted throughout the alteration sequence in the breccias, although the highest individual values—over 5% REO in the NOR breccia, south Richardson Mountains—are found in early magnetite-bearing assemblages (Hitzman, unpubl. data). Apatite and monazite are the only REE-bearing phases that have been identified to date. In the Redbank breccias, anomalous REEs are concentrated in apatite and "blue clay," which forms a matrix to the late fluidized breccias (Knutson et al., 1979).

The magnetite ores in the Kiruna district average 0.7% REO (Parak, 1973, 1975a, 1985). The REEs are contained in apatite and less abundant monazite which are not uniformly distributed throughout the ore. The hematitic ores at Kiruna have an average REO content of 0.5%, mostly contained in apatite. Apatite is present in nearly all the northern and central Swedish iron oxide deposits (Frietsch, 1979a), and REO grades are comparable to those at Kiruna (Parak, 1973). The magnetite deposits of the Great Bear magmatic zone also contain high percentages of apatite, and available analyses suggest that REO contents may be similar to the Swedish deposits (Hildebrand, 1986).

The iron oxide deposits in southeast Missouri display a complicated pattern of REE mineralization. In general, these ores contain less apatite than the Swedish ores, and thus the REO content of the iron oxide ores tends to be low, generally below 0.2%. However, local areas of highly anomalous REEs, with up to several percent REO in monazite and xenotime, have been discovered at Pea Ridge in late hematitic breccias cutting the magnetite bodies (Marikos et al., 1989).

Gold mineralization

Many Proterozoic iron oxide deposits contain anomalous gold, but its mineralogy, distribution, and paragenesis are poorly understood. In most cases where information is available, gold appears to occur primarily as native gold formed late in the paragenesis. However, the details of gold occurrences may be complex. At Olympic Dam, anomalous gold values (0.5 gm/tonne) occur throughout the hematitic breccias, and higher grades occur in a silicified fault zone on the eastern side of the central barren core of the deposit. This gold–silica mineralization appears to be a late-stage event that post-dates iron, uranium, and copper (Reeve et al., 1990). Anomalous gold values also occur at the Wirrda Well prospect on the Stuart Shelf (Parker, 1987), at the Damp and Sue-Dianne prospects in the Great Bear magmatic zone (R.T. Bell, pers. commun., 1990), and at Pea Ridge on the outer fringes of the iron oxide bodies and in late breccia bodies (Marikos et al., 1989; Seeger et al., 1989). Gold analyses are lacking for most of the Swedish deposits, but individual samples from both the magnetite and hematite ore bodies at Kiruna contain up to 2 g/tonne Au (Hitzman, unpubl. data). In the Wernecke breccias, Au values of several hundred ppb are common, and there are irregular zones of higher grade that are not easily correlative with observed breccia units or alteration phases. The highest gold values have been found peripheral to the breccia bodies in quartz-brannerite veins (Hitzman, unpubl. data).

Other elements

In addition to copper, uranium, REEs, and gold, a number of the deposits contain anomalous concentrations of other metals. At Olympic Dam, silver is recovered, and the 450 million ton reserve contains an average of 6 ppm Ag (Scott, 1987). The deposit also contains anomalous but subeconomic concentrations of cobalt, nickel, tellurium, and arsenic. The Wernecke breccias contain anomalous molybdenum, generally associated with uranium, and minor cobalt in cobaltite and cobaltian pyrite (Archer and Schmidt, 1978). The Bayan Obo ore bodies are potentially a major source of niobium (Argall, 1980). Late-stage Pb–Zn mineralization also is present at Bayan Obo but may not be related to the earlier iron oxide–REE mineralization. The Grängesberg deposit in central Sweden contains anomalous tin (Arönsson, pers. commun., 1987). The Pilot Knob and Boss- Bixby deposits in southeast Missouri contain trace molybdenum and anomalous cobalt in cobaltian pyrite and carrollite (Hagni and Brandom, 1990).

Summary of alteration–mineralization zoning

The alteration assemblages associated with the iron oxide (Cu–U–Au–REE) deposits are evidence of an important metasomatic component in the ore-forming process, in which alteration mineralogy is controlled by bulk-rock chemistry, composition of the mineralizing fluids, and *P–T* conditions of formation.

Alteration zoning in deposits hosted by igneous rocks

Alteration in iron oxide (Cu–U–Au–REE) systems in igneous rocks appears to show well developed zoning with respect to depth (Fig. 6). In the deeper portions of these systems, sodic alteration, containing albite–magnetite ± actinolite or chlorite, and generally lacking quartz, is dominant. Igneous feldspar is the first mineral to be altered, followed by fine-grained groundmass material and mafic minerals. Locally, especially in rocks of originally intermediate composition, complete conversion to albitite occurs.

Sodic alteration gives way upwards to a zone of potassic or sericitic alteration characterized by a potassium feldspar–sericite assemblage; here, quartz occurs both as veins and intergrown with other alteration products. In felsic rocks the normal assemblage is potassium feldspar–sericite–biotite–quartz. In rocks of intermediate to mafic composition at high structural levels, the assemblage contains potassium feldspar only in veins, and wall-rock alteration consists of the assemblage sericite–chlorite–actinolite ± epidote. Albitization of plagioclase probably occurs through much of this zone.

In some districts the potassic zone is overlain by a zone of sericitic or hydrolytic alteration. In rocks of intermediate composition the assemblage consists of hematite–sericite ± carbonate ± chlorite ± quartz, whereas in felsic rocks the assemblage is sericite–quartz–carbonate. Hematite is the stable iron oxide phase. Such hydrolytic–carbonatic alteration zones appear to have locally extended to the paleosurface, leading to hematite–quartz and barite deposition either as a direct, exhalative precipitant or as a near-surface replacement of sediments.

Sodic, potassic, and sericitic alteration may have formed contemporaneously, with inner and deeper zones encroaching upwards and outwards. There is also evidence from several districts, such as Kiruna and Pea Ridge, that the system may collapse: late quartz–sericite–(hematite) veins cut potassic and sodic alteration zones. Thus initial expansion of alteration zones may be followed by inward incursion of sericite. Breccia bodies have been recognized that cut all of the above alteration assemblages, and these breccias tend to be dominated by a hematite–sericite–(quartz) assemblage typical of hydrolytic alteration. In several systems, there is also evidence of a late period of silicification.

Magnetite, in massive bodies or irregular stockworks, is confined to sodic and potassic alteration zones. Hematite predominates in higher level sericitic alteration zones, and in late, cross-cutting breccia bodies. Both magnetite and hematite are found in the potassic zone; however, it appears that much of the he-

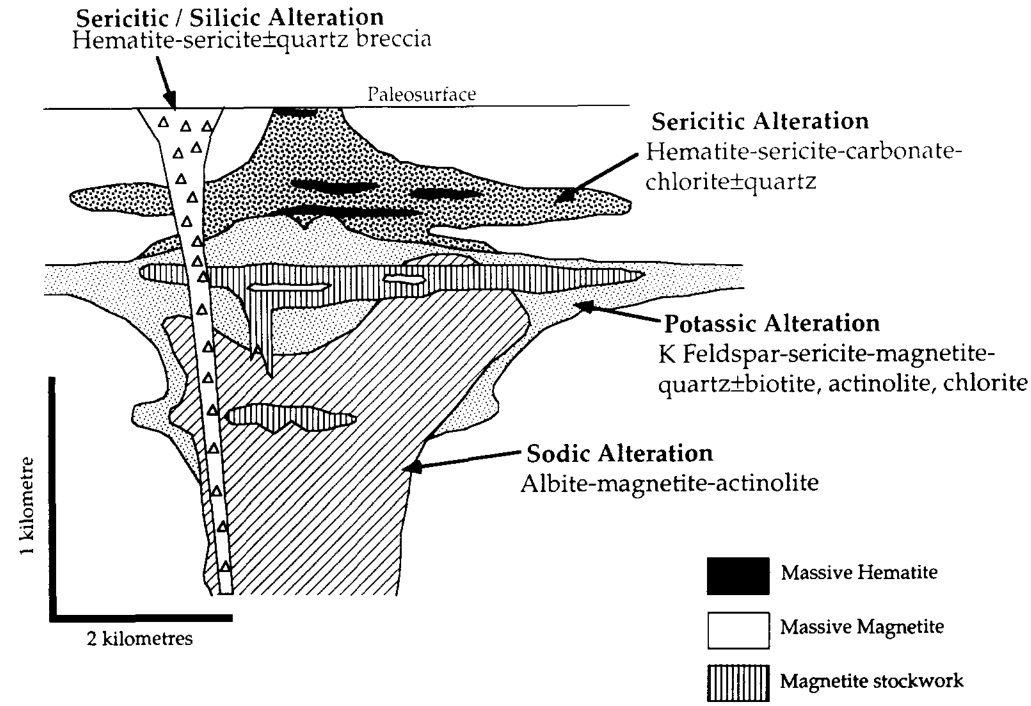

Fig. 6. Schematic cross section illustrating alteration zoning in iron oxide (Cu–U–REE–Au) deposits formed in volcanic and plutonic host rocks such as the Swedish, Missouri, and Stuart Shelf iron oxide deposits. The section extends from the near-surface to several kilometers below the paleosurface.

matite in this zone may be altered magnetite (martite). Cross-cutting breccia bodies are dominated by hematite, although there is evidence that some of the hematite replaces earlier magnetite in some deposits. Copper–uranium mineralization is recognized vertically throughout iron oxide deposits in igneous rocks but generally post-dates much of the alteration and iron mineralization and is more abundant in hematite-dominant systems. The late age of Cu–U mineralization may explain its concentration in the upper alteration zones and in the late, hematite-rich breccia bodies. Rare earth element mineralization is found in all of the stages of alteration, although the highest concentrations of REE minerals are generally associated with potassic and sericitic alteration or in late hematite-rich breccia bodies. The distribution of gold mineralization in these systems is not well defined; it appears to be concentrated in late zones of silicification that may represent the final stages of alteration and cooling. Post-mineralization supergene enrichment may also be important in many deposits.

Alteration zoning in sedimentary-hosted systems

A similar sequence of alteration and mineralization is recognized in systems formed in sedimentary rocks (Fig. 7). However, the mineralogy of individual alteration assemblages is more diverse due to the wider-range of wall-rock composition. Deep alteration in these systems is sodic, characterized by the development of albitite, albite–magnetite, and albite-sodium amphibole assemblages. In pelitic rocks this gives way upward and outwards to a albite–sericite(–paragonite)–magnetite assemblage.

This sodic zone is overlain by, or enclosed

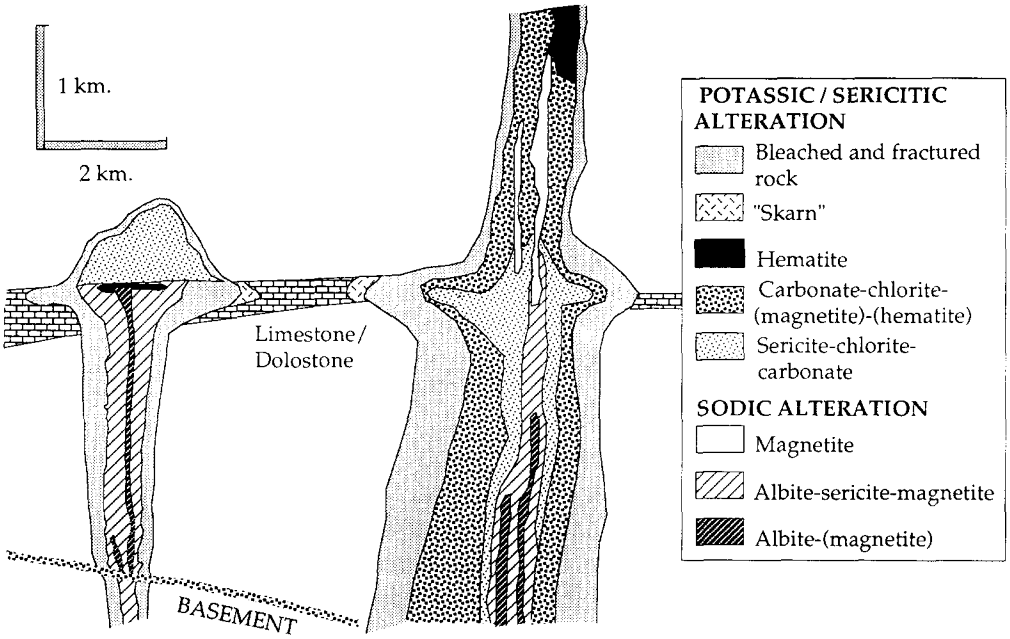

Fig. 7. Schematic cross section illustrating alteration zoning in iron oxide (Cu-U-REE-Au) deposits formed in dominantly sedimentary sequences such as the Wernecke Mountains, Yukon and Bayan Obo, China.

in, a potassic or sericitic alteration assemblage. Depending upon the host rock, this assemblage may consist of either a potassium feldspar-biotite-chlorite-carbonate assemblage in the case of argillaceous sediments, or a carbonate-chlorite-sericite assemblage in the case of siltstone, sandstone or calcareous sandstone.

Bleaching of the host rocks may be a prominent features of the outer fringes of the alteration zones. A well-developed sericitic zone has not been observed in any of the sediment hosted systems examined to date. However, this may be because near-surface deposits are yet to be identified in sedimentary sequences. The behavior of carbonate rocks in these systems is currently not well understood. It appears that carbonate is largely replaced by iron oxide, although calc-silicate development has been noted in the deeper portions of the Wernecke breccias (Hitzman, unpubl. data) and at Bayan Obo.

As in those systems formed in igneous rocks, magnetite is best developed in zones of sodic alteration, whereas both magnetite and hematite are present in the potassic or sericitic zone. Formation of hematite invariably post-dates magnetite. The behavior of copper, uranium, REEs, and gold appears to be the same in the sediment-hosted and igneous-hosted systems: much of the mineralization is late, and post-dates iron emplacement.

Temperatures of mineralization

Temperatures of mineralization are known for only a few of the deposits, but where available, they strongly support a genetic interpretation primarily involving low- to moderate-temperature conditions of formation. Critically, all the available data indicate sub-magmatic temperatures. At Olympic Dam, Oreskes (1990a) used fluid inclusion homogenization temperatures and oxygen isotope geothermometry to obtain maximum estimated formation temperatures near 400° C and apparent salinities of up to 42 wt% equivalent NaCl for early magnetite. Similar data suggests a temperature range of 110–400°C and a salinity range of 7 to 23 wt% equivalent NaCl for hematite mineralization, with a well defined mode of fluid inclusion homogenization temperatures of 180–190° C for hematite–quartz microbreccias from the core of the deposit. Conan-Davies (1987) utilized chlorite geothermometry to obtain consistent estimated formation temperatures of 200–280° C and 360–390° C for Mg- and Fe-rich chlorites, respectively, from hematite breccia.

In the Wernecke Mountains, fluid inclusion studies (Hitzman, unpubl. data) on dolomite, calcite, fluorite, and quartz in magnetite-rich assemblages in the Igor breccia body yield homogenization temperatures ranging from 80 to 300°C and salinities of 5 to 15 wt% equivalent NaCl. Minerals intergrown with hematite at Igor yield homogenization temperatures which cluster around 130° C. These data, and those from Olympic Dam, clearly show that the breccia-style deposits formed at the low-moderate temperatures consistent with near-surface hydrothermal environments.

Unfortunately, little temperature data are available for concordant and "ore breccia" style magnetite ore bodies. These magnetite-dominated systems appear to have formed at greater depth, and therefore perhaps at higher temperatures, than the hematite-rich, breccia-dominated systems. However, the limited data that are available also provides evidence of primarily low- to moderate-temperature, sub-magmatic environments. Preliminary fluid inclusion studies on minerals intergrown with both magnetite- and hematite-rich assemblages from the Pea Ridge deposit indicate very low temperatures of formation—between 100 and 200°C (R.A. Eisenberg, pers. commun., 1990). At the Acropolis deposit in South Australia, very limited oxygen isotope geothermometry suggests magnetite formation temperatures near 400°C (Oreskes and Einaudi, 1992). At Kiruna, oxygen isotope geothermometry suggests temperatures ca. 600°C for magnetite deposits (K. O'Farrelly, in Cliff et al., 1990). Thus the available evidence points to sub-magmatic, hydrothermal temperatures for most of the deposits of this class: early and deeper-level magnetite appears to have formed at somewhat higher temperatures (mostly 150–400°C, locally up to 600°C), and over a broader thermal range, than later and/or higher-level hematite (mainly 100–200°C).

Origin(s) of Proterozoic iron oxide (Cu–U–Au–REE) deposits

The majority of evidence from the iron oxide (Cu–U–Au–REE) deposits suggests they are primarily products of hydrothermal processes formed in an upper crustal environment. As in other hydrothermal systems, the alteration patterns associated with these deposits are coherently zoned and extend significantly beyond areas of massive to semi-massive iron mineralization. Alteration is expressed by replacement and veining of wall-rocks by secondary alteration minerals. Veins may contain iron oxides, phosphatic minerals, silicates, or some combination of these minerals. Iron oxides occurs as disseminations in wall-rocks, as well as in massive bodies and stockworks. Replacement textures of igneous and sedimentary rocks have been demonstrated in several deposits; these alteration and textural relationships are clear evidence of a hydrothermal origin for at least a significant portion of the mineralization.

Relation to iron-rich melts?

Although many previous workers have considered members of this class of deposit as formed by direct consolidation from a iron-rich melt, several lines of evidence suggest that this is improbable, or at least likely to be very rare, both on geological and theoretical grounds. Geologically, many of the deposits show little direct evidence of magmatic process and several have no associated igneous rocks at all. Where igneous rocks are present, as wall-rocks or in the stratigraphic section immediately below the deposits, they have a granitic through syenitic composition: there is no direct evidence of volumetrically significant anorthosites, gabbros or other intrusive rocks that would be a compositionally suitable source for an iron-rich immiscible melt. To this may be added the recently obtained evidence—cited above—of low- to moderate-temperatures of formation, which are inconsistent with a direct magmatic genesis at the site of ore deposition.

Arguments for an origin involving immiscible iron oxide melts have generally focussed on the magnetite–apatite association in many of these deposits, and the similarity of this assemblage to nelsonites associated with anorthosites. In such models, phosphorus acts as a flux which lowers the melting point of iron oxide melts. However, although many of these deposits do contain significant amounts of phosphatic minerals, commonly apatite, there are also deposits which contain virtually no phosphatic minerals. Thus, Philpotts' (1967) comparison of these magnetite + apatite ores with nelsonite deposits appears to be untenable. Furthermore, as pointed out above, the deposits under discussion are characteristically extremely Ti-poor, whereas nelsonite deposits are Ti-rich; indeed they are known as Ti ores. These distinctions clearly separate the two classes of deposits on both economic and geological grounds and undermines genetic arguments based on a close analogy between them.

A genetic model invoking direct consolidation of an iron phosphate melt is theoretically awkward as well. A phosphatic iron oxide melt would be expected to sink in a silicic or intermediate melt, owing to the density contrast between them (Daly, 1915; Philpotts, 1967). Furthermore, the high temperatures required to maintain such melts would favor crystallization and solidification within or near the parent magma, as is indeed the case in Fe–Ti deposits associated with anorthosites. Thus, such melts would be expected to be emplaced at deep crustal levels, in close proximity to their parent magmas, and probably under anhydrous, low f_{O_2} conditions. These constraints are consistent with the anhydrous mineral assemblages and abundance of magnetite and ilmenite observed in nelsonite deposits, but are inconsistent with the hydrous mineral assemblages, abundant hematite, and evidence of formation at high crustal levels seen in most of the iron oxide (Cu–U–Au–REE) deposits.

An often cited example of evidence that iron phosphate melts *do* reach the earth's surface, in spite of the thermal and density constraints that might impede them, is the Tertiary-aged El Laco deposit in Chile (Park, 1961; Frutos and Oyarzun, 1975; Henriquez and Martin, 1978). The composition of El Laco material—low titanium magnetite with apatite—is indeed similar to many of the magnetite bodies considered in this study. However, a comprehensive study of the El Laco occurrence has yet to be completed; until such studies are undertaken, the origin of El Laco must be considered uncertain. However, if a magnetite melt did reach the surface, it would likely be as a supercritical fluid and would display cooling textures similar to that of a basaltic or other non-viscous magma. Yet descriptions of the El Laco deposit suggest textures indicative of viscous flow, such as movement of the material only several hundred meters from the vent. This contradiction indicates a difficulty with a simple magmatic model and suggests that the El Laco iron oxides may have formed either by hydrothermal replacement of viscous flows, such as rhyolites, or by the escape to the surface of a hydrothermal fluid enriched in iron and phosphorus, as suggested by Hildebrand (1986) and Hauck (1990).

Relation to carbonatites?

The metal content of this class of deposit has similarities to suites observed in carbonatites and related ore deposits. However, none of the iron oxide (Cu–U–Au–REE) deposits is known to be associated with carbonatitic volcanic or intrusive rocks. Available carbon isotope data are inconclusive in this regard. Oreskes (1990a) reports an average $\delta^{13}C$ value of $-2.6‰$ from siderite at Olympic Dam; this value is consistent with derivation from either basement carbonates, carbonatites, or exsolved magmatic carbon dioxide dissolved in hydrothermal fluids. Preliminary stable isotopic studies on carbonate from the Kiruna magnetite and hematite bodies (D. Beaty, pers. commun., 1989) indicates a range of $\delta^{13}C$ values from -3 to $-5‰$, similar to those from Olympic Dam, although perhaps more suggestive of magmatic input. Carbonate minerals from several of the Wernecke breccia bodies (D. Beaty, pers. commun., 1989) and Bayan Obo iron oxide mineralization (Y. Qui, pers. commun., 1990) yield similar $\delta^{13}C$ values of -4 to $+2‰$. This broad spread of values probably represents a mixture of carbon from carbonate in the enclosing wall-rock as well as hydrothermal, or hydrothermally exchanged, carbon. Thus, carbon isotopic data are permisive of an ultimate carbonatite source for the fluids, but there is no direct supporting evidence for such an association. Furthermore, the evidence from Olympic Dam (Oreskes and Einaudi, 1990), the Wernecke breccias, Kiruna (Hitzman, unpubl. data), and Bayan Obo (Chao et al., 1990) that REEs have been introduced by hydrothermal fluids obviates the need to invoke alkalic or carbonatitic input of REEs at the site of deposition.

Source(s) of hydrothermal fluids

The relative importance of magmatic- and non-magmatic–hydrothermal fluids in the formation of these deposits remains to be determined. Oxygen isotope ratios from different alteration suites at Olympic Dam indicate the involvement of meteoric water and suggest a possible role for magmatic waters or deeply circulating meteoric waters that have inter-

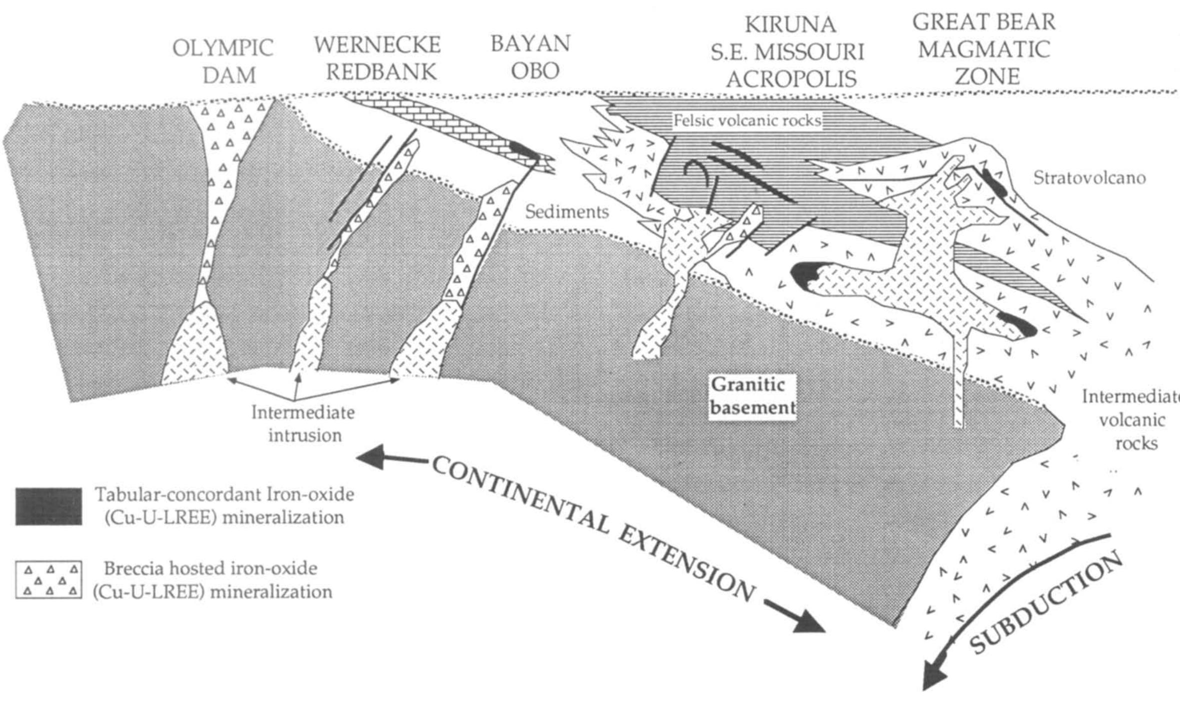

Fig. 8. Schematic representation of the tectonic setting and host rock sequence for iron oxide (Cu–U–REE–Au) deposits. All occur in continental regimes in areas of extension or rifting. The deposits form in a variety of host rocks characteristic of this setting.

acted with basement rocks (Oreskes, 1990a; Oreskes and Einaudi, 1992).

One characteristic of these deposits clearly relevant to the interpretation of their origin is their enormous size: many containing over 100 million tons of mineralized rock. This suggests that large volumes of hydrothermal fluid at an elevated temperature were necessarily involved in their formation. Such large volumes of fluid could have been generated from exsolution of a volatile-rich phase in very large and/or regionally extensive magma bodies, by deep circulation of large amounts of meteoric water, or by expulsion of large amounts of water from a sedimentary prism. The latter case is unlikely because thick clastic wedges are not present in many of the districts. Deep circulation of meteoric water would require a magmatic heat source to generate the necessary heat. The location of these deposits in continental areas undergoing extension (Fig. 8) provides a mechanism for the release of large volumes of magmatic–hydrothermal fluid from large, highly differentiated plutons that may be underplating the crust. Fracturing associated with extension may provide channelways to permit such fluids to reach shallow crustal levels. Alternatively, normal faults generated by crustal extension may facilitate deep circulation and heating of large amounts of meteoric water. The relatively low temperature of the mineralizing fluids suggests that these deposits formed from deeply circulating meteoric or metamorphic fluids, possibly combined with fluids of direct magmatic origin.

Interpretation of hydrothermal processes

A recurrent theme in the descriptions of alteration presented in the preceeding pages has been the relation between relative depth of ore formation, style of alteration, and textures and mineralogy of ore zones (Table 3, Figs. 6 and 7). These relations are comparable to those seen in many magmatic–hydrothermal systems and point to the essential commonality of this class of deposits with other types of ore deposits—rather than to some highly unusual magmatic process.

The depth of ore formation and its relation to patterns of wall-rock alteration, ore zones, structures, and breccias, allows an assessment of the generalized *P–T* trajectory of hydrothermal fluids within any given area of exposure (e.g., Norton, 1982; Sleep, 1983). It can be predicted, for example, that (1) prograde thermal paths would dominate in the high-level fringe recharge zones and at depth (below ore zones) during the inward cycling of fluids towards heat sources; (2) near-adiabatic cooling paths would locally dominate at and above the heat source, especially in the vicinity of major fluid conduits such as faults and breccia columns; and (3) retrograde thermal paths (cooling with declining pressure along an elevated geothermal gradient) would dominate well above the heat source and extend to the surface. The processes of wall-rock alteration and of metal acquisition, transport, or deposition are closely linked to these different *P–T* paths.

The progress of wall-rock alteration as affected by the temperature difference in the equilibrium constant of the K-feldspar–albite exchange reaction has been discussed by Hemley et al. (1971, 1980) and Giggenbach (1984). These authors point out that in granitic rocks, Na-metasomatism should be a common process in zones of descending or heating hydrothermal fluids (prograde path), whereas K-metasomatism should be dominant in major fluid upflow zones (retrograde path). These predictions are consistent with observations described in this paper: sodic alteration at deep levels, where hydrothermal fluids may be interacting with heat sources; potassic alteration at high levels, where hydrothermal fluids are cooling by interaction with country rock or meteoric water. Hemley et al. (1986) have shown that leaching of iron (and other metals) from wall-rocks can occur along a variety of hydrothermal *P–T* paths, including both the prograde thermal path and the near-adiabatic cooling path. The prograde cycle could result in metal leaching associated with sodic alteration; adiabatic cooling could also result in metal leaching, but this portion of the fluid path would be characterized by mild potassic alteration and/or sodic alteration. In contrast, the retrograde path could result in metal deposition associated with potassic alteration and lower temperature sericitic and silicic alteration.

The distribution of sodic vs. potassic alteration and its relation to mineral deposits at both the regional and local scale has been documented in numerous localities in addition to those summarized in this paper. These relations are consistent with the physical and chemical constraints outlined above, and point to the commonality of processes, spanning different ore deposit types, and related to hydrothermal fluid flow in the earth's crust. As an example of regional scale alteration patterns, felsic volcanic rocks of the Bergslagen district of south-central Sweden display extensive high-level potassic alteration associated with mineral deposits and deep-level sodic alteration associated with metal leaching (Lagerblad and Gorbatschev, 1985). Although complex in detail, the K–Na boundary at Bergslagen was a relatively horizontal surface, conformable with the volcanic pile. Rickard (1988) places the K–Na boundary at 130°C and 3 km below the sea floor in an extensional basin with high heat flow; an alternative interpretation, that involves convective fluid flow, is that peak temperatures of 300–350°C were attained on the prograde thermal path dominated by Na-metasomatism, and that the retrograde path produced K-metasomatism (P.A. de Groot, pers. commun., 1990).

Similar patterns have been documented on a more local scale in the Yerington district of Nevada. There, centrally located potassic alteration and bornite-magnetite ore are overlain and cut by quartz–sericite–pyrite–hematite alteration, formed in fluid upflow zones over granite cupolas, whereas peripheral zones of sodic alteration and metal leaching formed in fluid inflow zones at depth (Carten, 1986; Dilles, 1984). In contrast with Bergslagen, the K–Na boundary at Yerington represents a thermal peak of about 400°C for inwardly convecting fluids on the margins of plutons at depths of 2–5 km below the paleosurface.

In spite of variations in scale, temperatures of formation, and origins of hydrothermal fluids, the examples cited above have clear application to Proterozoic iron oxide deposits. The sodic-potassic alteration patterns that dominate the early and deep portions of the Proterozic districts are consistent with mass transfer in intermediate- to high-temperature hydrothermal fluids (approximately 400°C) of high chlorinity and intermediate pH. Deeply circulating meteoric or metamorphic fluids in a prograding thermal environment, possibly combined with hydrothermal fluids of direct magmatic origin, may have scavenged metals and other components that were later deposited as ores in the upper regions of the crust. Evidence of metal-leaching on a regional scale, such as that seen in the Bergslagen district, should be sought in the broad regions of sodic alteration underlying many of the Proterozoic iron ore districts.

The upper portions of the Yerington district, and of porphyry copper deposits in general (cf. Gustafson and Hunt, 1975), serve as analogues to high-level exposures of Proterozoic iron deposits in terms of the evolutionary styles of alteration; in both types of deposit, there is evidence of late, inward incursion of low temperature fluids that caused sericitic and silicic alteration in areas previously dominated by alkali metasomatism (e.g. Kiruna and Pea Ridge). Hydrothermal breccias are common in both deposit types, and these late breccias are dominated by quartz-sericite alteration accompanied by hematite and/or pyrite. In the porphyries, influx of oxygenated meteoric water has been shown on the basis of isotopic studies (Taylor, 1974) to be an important factor in the development of sericitic alteration.

Whether this is also true for the Proterozoic deposits remains to be determined.

In spite of the similarities outlined above between the Proterozoic iron deposits and Phanerozoic systems, there are, of course, striking differences that must be explained and that may relate to differences in concentration of sulfur in the hydrothermal fluids. Porphyry-type deposits are major sources of Cu and Mo and locally important sources of Pb and Zn, and sulfide minerals are far more abundant that iron oxides. That the hydrothermal fluids contain an excess of sulfur over that required to precipitate metals is suggested by the widespread sulfidation of mafic minerals to pyrite and by the abundance of sulfate fixed as anhydrite; in some deposits, there is more sulfur as anhydrite than as sulfides, and in general these deposits represent more of a sulfur anomaly than a copper anomaly. Furthermore, hydrothermal fluids rich in sulfur can generate abundant hydrogen ion through reactions of the type:

$$H_2S + 2O_2 = HSO_4^- + H^+$$

Such reactions are reflected in extensive acid attack of wall-rocks, which are commonly converted to kaolinite and/or pyrophyllite–alunite.

In contrast, the Proterozoic iron deposits are dominated by iron oxide, and sulfides are generally minor or absent. Copper is present in some deposits in economic proportions, but Pb and Zn are notably absent. The very low sulfide/iron oxide ratios in these deposits, combined with the low sulfidation state of assemblages such as magnetite–chalcopyrite, suggest low concentrations of sulfur in the hydrothermal fluids. Thus, the absence of major base-metal sulfide concentrations in most of the Proterozoic deposits may be explained by lack of sulfur rather than by lack of metals in solution. In turn, low sulfur concentrations in hydrothermal fluids yield less hydrogen ion when oxidized at low temperatures, leading to less pronounced acid attack of wall-rocks.

Are these deposits necessarily Proterozoic?

The majority of large deposits of this type are restricted to Early- to mid-Proterozoic (1.1–1.9 Ga) host rocks, and they share a similar regional tectonic setting. They are found in continental areas that contain evidence of rifting or aborted rifting during or prior to the onset of mineralization. Districts and individual deposits are commonly located along major zones of crustal weakness such as regional faults or shear zones. In several cases, mineralization may directly follow the emplacement of the host rocks to the mineralization: a number of the deposits are found in or immediately below regionally extensive felsic volcanic sequences generally characterized by ash-flows (Stuart Shelf, Kiruna, southeast Missouri, Great Bear magmatic zone). Others are found in sedimentary prisms believed to have been deposited in aulacogens or along rifted continental margins (Wernecke Mountains, Bayan Obo, Redbank district). This suggests that the mineralization may be broadly related to crustal extension following the initiation of crustal stabilization in the Early Proterozoic (West, 1980), and in particular, to a global extensional event that occurred at 1.4 Ga, namely, the break-up of a Proterozoic super-continent (Herz, 1969; Burke, 1977; Piper, 1983; Bond et al., 1984; Lindsay et al., 1987; Hoffman, 1989).

However, although the deposits listed in Table 1 are all found in Early- to mid-Proterozoic host rocks, similar deposits also occur in younger rocks. Paleozoic examples include stratiform and diatreme-shaped iron deposits in the Bafq-Seghand district in central Iran (Förster and Knittel, 1979; Taghizadeh, 1977; Förster, 1990) and stratiform deposits in the Avnik district of southeastern Turkey (Helvaci, 1984). Mesozoic examples include deposits in the Cortez Mountains and the Buena Vistas Hills of Nevada (Jones, 1913; Shawe et al., 1962) and the La Serena-Copiapo iron district of Chile (Bookstrom, 1977). Tertiary examples may include the iron deposits of the Durango district of Mexico (Young et al., 1969; Lyons, 1975, 1988; Swanson et al., 1978), the La Perla district of Mexico (Van Allen, 1978), the Abouyan district of Armenia (Zitzmann, 1978), and the El Laco region of Chile (Park, 1961; Frutos and Oyarzun, 1975; Henriquez and Martin, 1978). However, whereas the Proterozoic examples all occur in cratonic environments, the Phanerozoic deposits occur primarily in continental arc environments as well as in areas of extension behind a continental magmatic arc. Thus, we may expect to find similar examples in other areas of Phanerozoic rifting or back-arc extension. However, it is noteworthy that these younger deposits are typically much smaller than their enormous Proterozoic counterparts. This seems to support the interpretation that the Proterozoic deposits are associated with a unique event in Earth history—the break-up of a Proterozoic supercontinent. This event occurred at a time in Earth history when both the continental masses were mostly or entirely amalgamated, and the temperature profile of the Earth elevated compared to later Phanerzoic history (Hoffman, 1989). The combination of these two factors—thermal insulation of the upper mantle by a large region of continental crust, coupled with a high geothermal gradient—may have resulted in large-scale magmatic underplating of the crust, and consequent generation of enormous hydrothermal systems. Secular cooling of the Earth since the Proterozoic has insured that subsequent events might be similar in kind but smaller in magnitude.

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