Chapter 2

Proterozoic Lode Gold and (Iron)-Copper-Gold Deposits: A Comparison of Australian and Global Examples

GREG A. PARTINGTON†

Ross Mining N.L., P.O. Box 1546, Milton, Queensland 4064, Australia

AND PATRICK J. WILLIAMS

Economic Geology Research Unit, James Cook University, Townsville, Queensland 4811, Australia

Abstract

More than 150 Moz of gold has been added in production and resources from Proterozoic deposits in the last ten years, and many Proterozoic basins are now considered high priority exploration targets. The bulk of Proterozoic gold is produced from lode gold and Cu-Au (U-REE-Ba-F) deposits which are found in northern Australia, South Dakota, West Africa, Canada, South Africa, Scandinavia, and Central America.

Proterozoic lode gold deposits are restricted to late collisional stages in the development of Proterozoic orogenic belts. They appear to have a systematic sequence of events in common and occur in linear belts associated with regional ductile structures at, or near, the greenschist facies brittle-ductile transition. Gold occurs in a large variety of rock types and has a close spatial association with regional-scale domes, anticlines, strike-slip shear zones, duplex thrusts, and in some deposits, geochemically distinct granites. Deposit styles can be subdivided into several types, directly related to the host structure and to contrasts in host-rock competency and mineralogy. These deposits have fluids and geochemical associations that overlap those of Archean lode gold deposits.

Proterozoic Cu-Au- (Fe) deposits formed in a broader range of crustal and tectonic environments and display a great variety of structural and host-rock controls and styles. It is evident in all districts where the timing relationships are known that these deposits have spatial and temporal relationships to granites. These deposits display a range of fault and shear zone controls and are commonly associated with regions of geometric complexity, structural intersections, or regionally anomalous structural orientations. There is considerable evidence of variable fluid chemistry in Cu-Au-(Fe) deposits. Districts are commonly characterized by regional metasomatism and alteration at both regional and deposit scale which is commonly intense. Fe oxide-Cu-Au environments tend to produce similar alteration assemblages in all aluminous rock types. The influence of magmas as sources of fluid and ore components appears to have been greater in at least some Cu-Au-(Fe) systems and the associated granitoids are typically oxidized and include both mafic and felsic varieties. Sodic alteration styles are commonly prevalent regionally; the larger ore systems in particular are hosted specifically within substantial bodies of rock that are depleted in Na and enriched in K-Fe-(H).

Introduction

IN THE PAST there was a perception that Proterozoic gold deposits were relatively small and unimportant (Hutchinson, 1987; Woodall, 1990). However, the contribution from Proterozoic gold deposits has increased significantly in the last ten years due to recent exploration success in regions such as Africa, South America, and northern Australia, together with historical production from areas such as America and Australia (Fig. 1; Table 1). Consequently, many Proterozoic rock systems are now considered high priority exploration targets. A number of new operations have been opened (Table 1), adding more than 150 Moz compared to ten years ago (cf. Woodall, 1990).

Many different types of Proterozoic gold-bearing deposits have been described. Davidson and Large (1994) recognized seven categories in Australia alone. However, in terms of world production the most important Proterozoic gold deposits can more simply be classified into lode deposits formed late in orogenic cycles, and those with (Fe)-Cu-Au affinities. Other deposit types, for example, unconformity U-Au ± Pt and Pd deposits (Carville et al., 1990; Hancock et al., 1990) and base metal-rich gold deposits like Broken Hill in Australia (Davidson and Large, 1994), will not be considered further in this paper since they account for less than 7 percent of the total inventory of Proterozoic gold production. Australia has a particularly important place in Proterozoic gold studies given the coincidence of important lode gold deposits and (Fe)-Cu-Au deposits (Fig. 2). With one major exception (Telfer), these Australian deposits formed between 1850 and 1500 Ma, which corresponds to a major global peak of continental metallogeny (Barley and Groves, 1992).

†Corresponding author: e-mail, gapmas@ozemail.com.au

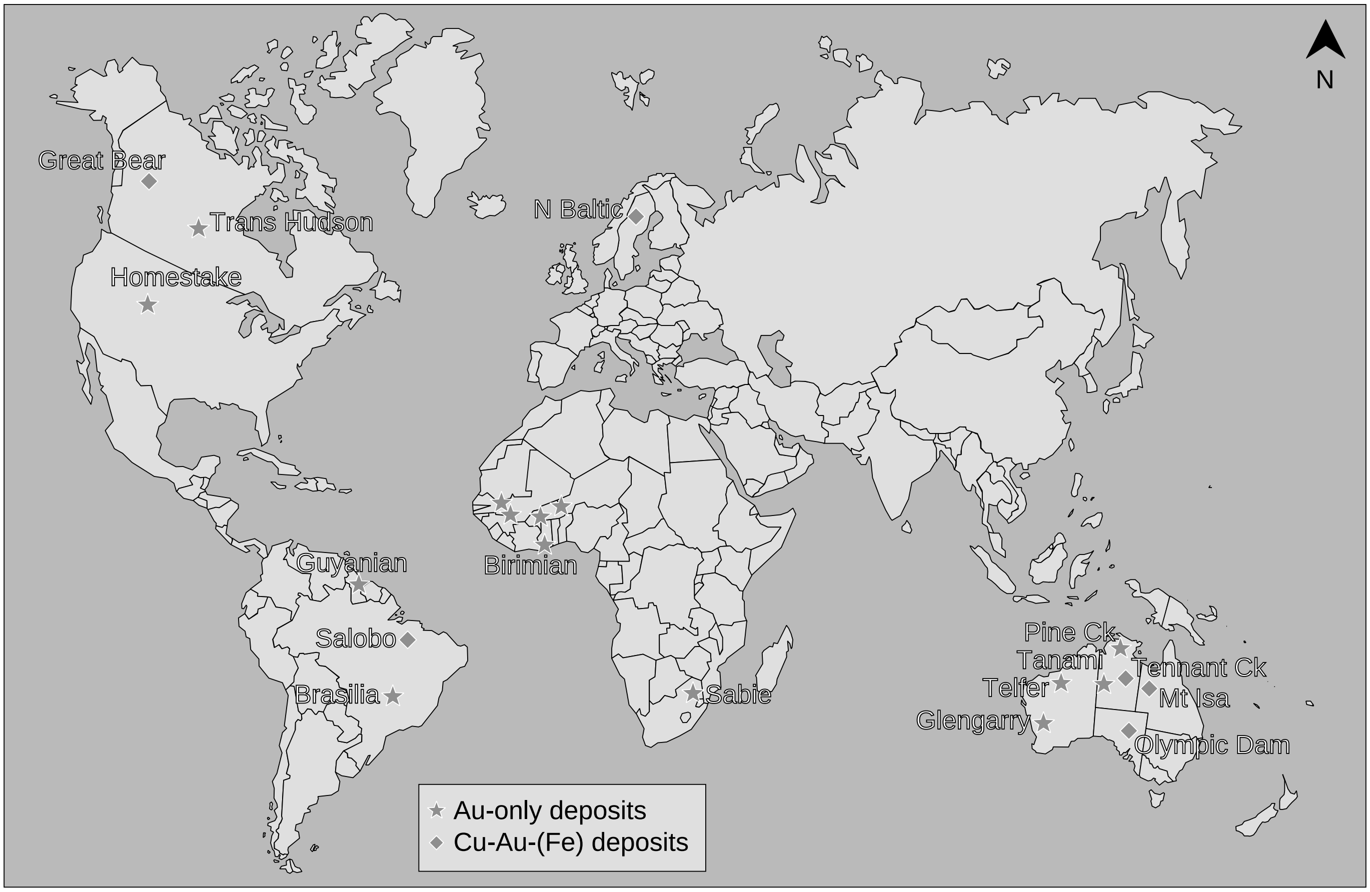

FIG. 1. World location map for Proterozoic gold deposits, showing the general location of both the lode gold and Cu-Au deposits.

Proterozoic lode gold deposits are found in the Pine Creek geosyncline in northern Australia, the Glengarry basin and the Paterson province in Western Australia, the Tanami Desert in central Australia, the Black Hills in South Dakota, the Birimian Supergroup of West Africa, in Guyana and French Guiana, Central America, the Trans-Hudson orogen of northern Saskatchewan and Manitoba, the Sabie-Pilgrim's Rest gold field in South Africa, and the Proterozoic Brasilia fold belt (Fig. 1; Tables 1 and 2).

Proterozoic terranes, especially within Australia, also display a particular endowment of Cu-Au (± U ± REE ± Ba ± F) deposits with a distinctive Fe oxide association (e.g., Hitzman et al., 1992; Davidson and Large, 1994, 1998; Williams, 1998). Additional examples occur in northern Scandinavia, northwestern Canada, southeast Missouri, and potentially in Brazil (Fig. 1). Many of the deposits have only been discovered recently and they occur in a wide range of geologic settings and in detail display very variable characteristics (Table 1). The deposits appear to form part of an as yet unexplained association, which includes Fe oxide-apatite bodies, some of which may have crystallized directly from melts (Hitzman et al., 1992; Nystrˆm and Henriquez, 1994).

Although Proterozoic gold deposits have a variety of geologic characteristics that make each deposit unique, they also have many features in common. Any discussion of Proterozoic gold mineralization would be incomplete without considering the spatial relationship between granites and gold (e.g., Goellnicht et al., 1989; Wall and Taylor, 1990; Caddey et al., 1991; Ansdell and Kyser, 1992; Marcoux and Milesi, 1993; Davidson and Large, 1994). This paper will examine this association using recent data (Matthai et al., 1995a; Klominsky et al., 1996; Oberthur et al., 1996; Partington and McNaughton, 1997; Rowins et al., 1997; Smith et al., 1998; Wyborn, 1998; Wyborn et al., 1998; McLaren et al., 1999; Williams, 1999) and will compare Australian examples with other Proterozoic gold deposits with particular reference to the apparent link between the diverse styles and chemical characteristics of Proterozoic gold mineralization, granite intrusion, and tectonism.

A spectrum of deposits is described in this paper (Table 1) and geochemical data from these are used to constrain chemical controls on mineralization. Fluid inclusion and isotopic data are also used to identify a possible fluid source for gold mineralization. A review of the structural and geochemical controls on mineralization from a macroscopic or

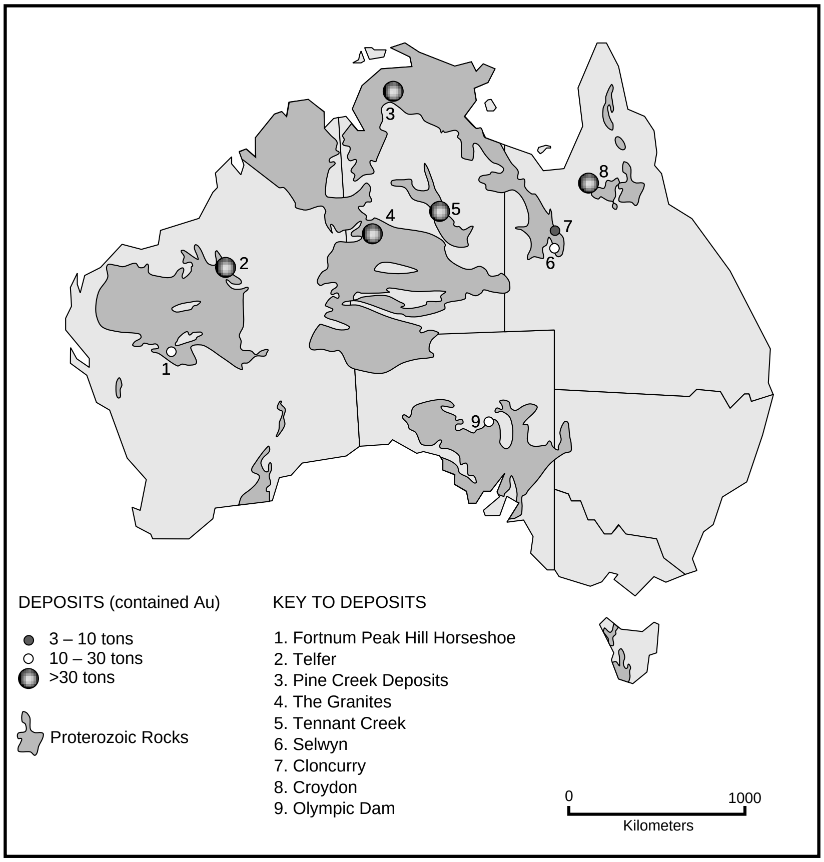

FIG. 2. General location of important Australian Proterozoic gold deposits, subdivided according to lode gold and Cu-Au types with circles showing gold endowment. Production is cumulative including past production and current mineral resources as shown in Tables 1 and 2.

mine scale to a mesoscopic or regional scale is made, and suggested genetic models are discussed with reference to current exploration techniques.

Australian Proterozoic Lode Gold Deposits

Proterozoic lode gold deposits are products of late collisional stages in the development of Proterozoic orogenic belts, as is the case for Late Archean lode gold deposits and Phanerozoic deposits (Kerrich and Cassidy, 1994). They appear to have a systematic sequence of events in common, including syncollisional peak metamorphism and development of ductile fabrics and gold mineralization during postpeak metamorphic brittle-ductile reactivation of earlier structures accompanying postcollisional uplift and cooling. Gold mineralization is younger than its host rocks and postdates regional metamorphism and deformation. Many of the deposits are also spatially associated with granite intrusion, especially in Australia with reduced fractionated I-type granites (Wyborn, 1994; Wyborn et al., 1994, 1998; Partington and McNaughton, 1997). However, some granites are also mineralized, suggesting a complex temporal relationship between the end of orogenesis, granite intrusion, contact metamorphism, and gold mineralization. There are very few areas where there are enough geochronological data to establish the relationship between gold mineralization and granite intrusion clearly.

Pine Creek geosyncline (Australia)

The Pine Creek geosyncline hosts several deposits (Cosmo Howley, Mount Bonnie, and Moline) above 1 Moz that are similar to iron-formation deposits like the Homestake mine in South Dakota and slate belt-style deposits such as those (Enterprise, Chinese Howley, and Goodall) of the Victorian gold field in Australia (Fig. 3; Tables 1 and 2). The Pine Creek geosyncline comprises a supracrustal sequence that consists of fine-grained clastic sedimentary rocks, ironstone, minor evaporite and platform carbonates, acid volcanic rocks, and basic intrusive rocks. Sedimentation and volcanism occurred between 2000 and 1870 Ma in an intracratonic basin formed by crustal extension of predominantly Archean basement. The sediments and basic intrusion were deformed, regionally metamorphosed, and subsequently intruded by the Cullen batholith. This batholith is composed of a series of stocks and flat sheets that coalesce at shallow depths. The presence of numerous roof pendents and the distribution of the thermal aureole around the batholith

TABLE 1. Significant Proterozoic Gold Resources
(metal endowment is cumulative production plus current mineral resource for either an individual deposit or mineral field)

Mine	Location	Age	Au Oz Cu kt	Host Rock	Style	Reference
Cosmo Howley	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	2,000,000	Middle Koolpin Formation iron-rich mudstone	Stratiform quartz vein and replacement lodes associated with thrust faulting along the Howley Anticline	Alexander et al., 1990
Brocks Creek Faded Lily and Alligator	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	872,000	Mt. Bonnie Formation Graywacke siltstone and tuff	Steep southerly dipping quartz veins in shear zone on south plunging anticline	Miller et al., 1998
Toms Gully	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	455,000	Mt. Partridge Group siltstone, shale and tuff	Quartzvein in the plane of an early thrust	Ahmad et al., 1993
Woolwonga	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	363,000	Mt. Bonnie Formation Graywacke and siltstone	Saddle reefs, quartz stockwork and discordant veins in tight southeast plunging anticline with bounding footwall thrust	Ahmad et al., 1993
Goodall	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	228,000	Burrell Creek Formation Shale, graywacke and siltstone	Sub-vertical, north trending sheeted vein sets within shear/fault fracture zone on eastern limb of Howley Anticline	Quick, 1994
Rustlers Roost	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	958,000	Mt. Bonnie Formation Siltstone, tuff and chert	Quartz stockwork and discordant veins in a wide duplex system	Ahmad et al., 1993
Enterprise	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	1,381,000	Mt. Bonnie/Burrell Ck Formations Shale, graywacke and siltstone	Saddle reefs, quartz stockwork and discordant veins in tight south plunging parasitic fold on D_2 anticline	Cannard and Pease, 1990
Union Reefs	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	1,330,000	Burrell Ck Formation Shale, graywacke and siltstone	Saddle reefs, quartz stockwork and discordant veins in Pine Ck shear zone	Ahmad et al., 1993
Mt. Todd	Pine Creek Geosyncline NT Northern Australia	Early Proterozoic	3,400,000	Burrell Creek Formation Shale, graywacke and siltstone	Stacked narrow en echelon quartz veins in axial planar cleavage confined to coarse sandstone beds in turbidites	Ormsby et al., 1998
U-Au Deposits	Pine Creek Geosyncline NT Northern Australia	Mid Proterozoic	1,049,000	South Alligator Formation Shale, tuff, dolerite, sandstone, siltstone	Elongate lenses beneath Middle Proterozoic unconformity	Carville et al., 1990, Hancock et al., 1990, Mernagh et al., 1994
The Granites	Tanami Desert Central Australia	Early Proterozoic	1,300,000	Mt. Charles Beds Metamorphosed sediments, iron-formation, basic volcanics	Intense folding and associated shearing controls mineralization, which is in quartz veins and altered shear zones	Mayer, 1990
Callie	Tanami Desert Central Australia	Early Proterozoic	3,500,000	Mt. Charles Beds Metamorphosed sediments, iron-formation, basic volcanics	Folding and associated shearing controls mineralization, which is in quartz veins in fold axes	Smith et al., 1998
Tanami Corridor	Tanami Desert Central Australia	Early Proterozoic	1,200,000	Mt. Charles Beds Metamorphosed sediments, iron formation, basic volcanics	Folding and associated shearing controls mineralization, which is in quartz veins in fold axes	Tunks and Marsh, 1998
Telfer	Paterson Province Northern Western Australia	Late Proterozoic	10,400,000	Yeneena Group Calcareous and carbonaceous siltstone	Doubly plunging anticlines associated with strike slip faults	Dimo, 1990, Rowins et al., 1997
Glengarry Basin: Fortnum, Peak Hill and Horseshoe	Glengarry Basin Western Australia	Late Proterozoic	500,000	Glengarry Group Graywacke mafic volcanics and pelites	Regional folds associated with transpressional faults	Parker and Brown, 1990, Hanna and Ivey, 1990
Homestake	Black Hills South Dakota USA	Early Proterozoic	57,000,000	Homestake formation Grunerite-siderite iron formation, graphitic shale and basaltic schist	Stratiform quartz vein and replacement lodes associated with thrust faulting; Multiple ore shoots called ledges	Caddey et al., 1991

TABLE 1. (Cont.)

Mine	Location	Age	Au Oz Cu kt	Host Rock	Style	Reference
Obuasi	Ghana	Early Proterozoic	20,000,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with shoots in dilational structures	Resource Information Unit, 1998, Oberthur et al., 1998
Obotan	Ghana	Early Proterozoic	1,300,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with shoots in dilational structures	Resource Information Unit, 1998, Oberthur et al., 1998
Konongo	Kumasi, Ghana	Early Proterozoic	1,680,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with shoots in dilational structures	Resource Information Unit, 1998, Oberthur et al., 1998
Bibiani	Sefwi belt South Ghana	Early Proterozoic	1,700,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with shoots in dilational structures	Resource Information Unit, 1998, Oberthur et al., 1998
Prestea	Ashanti Belt, West Ghana	Early Proterozoic	6,000,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with shoots in dilational structures	Resource Information Unit, 1998, Oberthur et al., 1998
Bogosu	Ashanti Belt, West Ghana	Early Proterozoic	1,700,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Sear hosted with shoots in dilational structures	Resource Information Unit, 1998, Oberthur et al., 1998
Damang	Ashanti Belt, West Ghana	Early Proterozoic	4,000,000	Birimian greenstone, including dolerite, conglomerate, argillite and graywacke	Mineralization controlled by the Damang anticline in shears and quartz stockwork al., 1998	Resource Information Unit, 1998, Oberthur et
Yamfo	Sefwi-Yamfo belt Ghana	Early Proterozoic	4,300,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with shoots in dilational structures	Resource Information Unit, 1998, Oberthur et al., 1998
Siguiri	NE Guinea	Early Proterozoic	3,400,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted	Resource Information Unit, 1998, Marcoux and Milesi, 1993, Huot et al., 1987
Poura	Central Burkina Faso	Early Proterozoic	710,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Quartz reefs in shear zones	Resource Information Unit, 1998, Marcoux and Milesi, 1993, Huot et al., 1988
Kodieran	Wassoulou region Mali	Early Proterozoic	1,700,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted	Resource Information Unit, 1998, Marcoux and Milesi, 1993, Huot et al., 1987
Loulo	West Mali	Early Proterozoic	3,600,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted	Resource Information Unit, 1998, Marcoux and Milesi, 1993, Huot et al., 1987
Sadiola	West Mali	Early Proterozoic	4,200,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Fault hosted	Resource Information Unit, 1998, Marcoux and Milesi, 1993, Huot et al., 1987
Syama	South Mali	Early Proterozoic	5,300,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with elongate shoots forming fine quartz vein stockwork related to thrusting Also supergene overprint	Resource Information Unit, 1998, Marcoux and Milesi, 1993, Olson et al., 1992

TABLE 1. (Cont.)

Mine	Location	Age	Au Oz Cu kt	Host Rock	Style	Reference
Tiawa	Niger	Early Proterozoic	1,800,000	Birimian greenstone, including basalt, andesite, argillite and graywacke	Shear hosted with shoots in dilational structures	Resource Information Unit, 1998
Teberebie	Tarkwa Basin, West Ghana	Early Proterozoic	9,000,000	Tarkwaian conglomerate	Paleoplacer	Resource Information Unit, 1998 and Oberthur et al., 1998
Tarkwa	Tarkwa Basin, West Ghana	Early Proterozoic	14,200,000	Tarkwaian conglomerate	Paleoplacer	Resource Information Unit, 1998, Oberthur et al., 1998
Iduapriem	Tarkwa Basin, West Ghana	Early Proterozoic	4,400,000	Tarkwaian conglomerate	Paleoplacer	Resource Information Unit, 1998, Oberthur et al., 1998
Abosso	Tarkwa Basin, West Ghana	Early Proterozoic	2,500,000	Tarkwaian conglomerate	Paleoplacer	Resource Information Unit, 1998, Oberthur et al., 1998
Omai	Guyanan Craton, South America	Early Proterozoic	4,000,000	Felsic and mafic volcanic rocks, quartz porphyry, siltstone, sandstone, shale, graywacke	Shear hosted with shoots in dilational structures	Marcoux and Milesi, 1993, Voicu et al., 1999
Trans-Hudson, Including Nor-Acme, Flin Flon, Star Lake, and Rio	Flin Flon area; Saskatchewan and Manitoba, Canada	Early Proterozoic	1,000,000	La Ronge and Flin Flon greenstone belts; Mafic volcanics, sandstone, siltstone, shale, diorite, graywacke	Shear hosted with shoots in dilational structures; Quartz veins associated with sulfide	Ibrahim and Kyser, 1991, Ansdell and Kyser, 1992
Sabie-Pilgrim's Rest	East Transvaal Basin, South Africa	Early Proterozoic	1,800,000	Malmani Subgroup and Pretoria Group, including dolomite, shale and sandstone	Stratiform reefs controlled by duplex thrusting	Harley and Charlesworth, 1992, Boer et al., 1993
Aitik	Northern Baltic Shield (Sweden)	Early Proterozoic	6,400,000 *3000*	Schist, gneiss	Disseminated, minor veins in arcuate (?retrograde shear-hosted) muscovite-biotite schist zone along granitoid contact	Zweifel, 1976: Wanhainen and Martinsson, 1999
Bidjovagge	Northern Baltic Shield (Norway)	Early Proterozoic	50,000 *54*	Graphite schist, amphibolite	Veins, breccia and stratiform disseminations localized near hinge of tight anticline	Bjørlykke et al., 1987, Ettner et al., 1994
Pahtohavahre	Northern Baltic Shield (Sweden)	Early Proterozoic	50,000 *32*	Graphite schist	stratabound veins on fault intersection with graphitic horizons in antiformal hinge	Lindblom et al., 1996, Martinsson, 1997
Alemao	Carajas district, Southern Amazon Craton, Brazil	Early Proterozoic	4,800,000 *2475*	Metavolcanic rocks, Diorite dikes	Steeply dipping tabular breccia bodies	CFBarreira, verbal communication , 1999
Pojuca	Carajas district, Southern Amazon Craton, Brazil	Early Proterozoic	*531*	Gruneritic iron formation	Stratiform replacement and fracture fill	Winter, 1994
Salobo	Carajas district, Southern Amazon Craton, Brazil	Early Proterozoic	16,000,000 *9950*	Fayalite-magnetite-almandine-biotite ironstones in a metagraywacke and amphibolite sequence	Laterally-extensive, thick massive ironstone lenses	Minorco 1997 Annual Report, Lindenmayer and Teixeira, 1999, Requia and Fontbote, 1999
NICO – Bowl Zone	Great Bear Magmatic Zone, NWT, Canada	Early Proterozoic	1,200,000 *2*	Hornfelsed arkose, graywacke and rhyolite	Au-Co-Bi mineralized biotite-amphibole-magnetite schist zone at fault intersection with basement discontinuity	Goad, 1998

TABLE 1. (Cont.)

Mine	Location	Age	Au Oz *Cu kt*	Host Rock	Style	Reference
Sue-Dianne	Great Bear Magmatic Zone, NWT, Canada	Early Proterozoic	30,000 *105*	Rhyodacitic ignimbrite	Breccia body on fault intersection	Johnson and Hattori, 1994, Goad, 1998
Boss-Bixby	St Francois Terrain, midcontinent USA	Mid Proterozoic	no data	Syenite, rhyolite	Breccia, vein and replacement mineralization at contact of central intrusive complex	Hagni and Brandom, 1989, Kisvarsanyi, 1989
Gecko/K44	Tennant Creek Inlier, northern Australia	Early Proterozoic	140,000 *246*	Graywacke, siltstone, shale, hematitic shale	Metasomatic ironstone in parasitic anticlinal hinge	Main et al., 1990, Huston et al., 1993
Juno	Tennant Creek Inlier, northern Australia	Early Proterozoic	800,000 *1*	Graywacke, shale, hematitic shale	Metasomatic ironstone in parasitic anticline	Large, 1975
Nobles Nob	Tennant Creek Inlier, northern Australia	Early Proterozoic	1,100,000	Sandstone, shale, hematitic shale	Metasomatic ironstone at shear zone intersection with shale unit	Yates and Robinson, 1990
Peko	Tennant Creek Inlier, northern Australia	Early Proterozoic	390,000 *148*	Graywacke, shale	Metasomatic ironstone in faulted parasitic anticline	Wedekind et al., 1989, Rattenbury, 1992
Warrego	Tennant Creek Inlier, northern Australia	Early Proterozoic	1,500,000 *139*	Quartz-muscovite schist, quartzite, slate	Metasomatic ironstone possibly in fold limb	Wedekind and Love, 1990
White Devil	Tennant Creek Inlier, northern Australia	Early Proterozoic	740,000	Graywacke, siltstone, shale	Metasomatic ironstone hosted by shear zone in anticlinal hinge	Huston et al. 1993, Bosel and Caia, 1998
Olympic Dam	Gawler Craton, southern Australia	Mid Proterozoic	38,600,000 *32,000*	Granite	Fault-controlled shallow level breccia complex (? diatreme)	Reeve et al., 1990, Oresekes and Einaudi, 1990
Eloise	Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	140,000 *170*	Meta-arkose, amphibolite	Replacement deposit in steeply-inclined ductile-brittle shear zone	Baker, 1998, Baker and Laing, 1998
Ernest Henry	Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	2,700,000 *1830*	Metamorphosed intermediate to felsic volcanic rocks	Breccia matrix and replacement mineralization controlled by moderately dipping ductile-brittle shear zone network	Twyerould, 1997, Ryan, 1998
Great Australia	Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	? *44*	Amphibolite	Carbonate-rich vein stockwork at fault splay intersection	Cannell and Davidson, 1998
Greenmount	Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	90,000 *54*	Carbonaceous slate	Vein and replacement mineralization on reverse fault subparallel to bedding with juxtaposition of different stratigraphic sequences	Krcmarov and Stewart, 1998
Mount Elliott	Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	190,000 *120*	Schist, phyllite, amphibolite, trachyandesite dykes	Skarn vein and breccia deposit controlled by brittle faults	Fortowski and McCracken, 1998, Wang and Williams, in press
Osborne	Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	470,000 *345*	Meta-arkose, stratiform metasedimentary ironstone, metaperidotite, pegmatite, amphibolite	Silicified masses controlled by ductile-brittle shear zones with variable moderate to shallow dip Some localization around pegmatites and on ironstone contacts	Adshead, 1995, Adshead et al., 1998
Starra	Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	1,100,000 *115*	Ironstone, schist, calc-silicates, amphibolite	Selective mineralization of metasomatic magnetite ironstones controlled by ductile brittle shear zones	Davidson et al., 1989, Rotherham, 1997, Adshead-Bell, 1998
Tick Hill	SW Cloncurry district, Mount Isa Inlier, Queensland	Mid Proterozoic	500,000 *0*	Calc-silicates, schist, quartzite	High grade native gold selectively developed in feldspathic bodies localized by high strain zone in limb of asymmetric synform	Crookes, 1993

TABLE 2. Summary of Proterozoic Lode Gold and Copper-Gold Deposit Districts
(district metal endowment is cumulative from those deposits listed in Table 1)

Deposit	Gold (Moz)	Geologic setting	Structural setting	Style	Ore minerals	Alteration	Element association	Temporal relationships	Fluid inclusion data	Isotope data	Genetic model	Exploration	Main references
Pine Creek	10.99	Host rocks range from iron-formation through felsic tuff and chert to turbidite sequences and graywacke	Gold deposits occur within the contact aureole of the Cullen Batholith associated with regional scale thrusts, ramp anticlines and strike-slip shear zones	Strata bound in iron formation, in thrusts or as slate belt style bedding parallel quartz veins in folds	Gold, quartz, pyrite, arsenopyrite and minor galena, chalcopyrite and sphalerite	Quartz, chlorite, sericite, carbonate, K-feldspar, sulfide and tourmaline	Au, SiO_2, CO_2, Ag, K, Fe, S, Ba, As, Bi, W and Sb.	Gold mineralization postdated granite intrusion and contact metamorphism, but was synchronous with late tectonism at about 1810Ma	Early quartz veins from moderate salinity CO_2 ± CH_4 fluid at 1 kbar and 365° C; Gold introduced by a late lower temperature fluid with higher salinity $CaCl_2$ ± $MgCl_2$ at 195°C; Contributions from magmatic and metamorphic sources	$\delta^{34}S$ range 4 to 10 ‰, $\delta^{13}C$ range +1 to –32 ‰, δD range –27 to 57 ‰ and $\delta^{18}O$ from 8.05 to 16 ‰ Implying a mixed magmatic metamorphic source. CO_2 and S from sediments	Isotopes suggest mineralization is related to metamorphic dewatering during contact metamorphism. Both magmatic and metamorphic fluids mixed to form the gold deposits. There is an association with HHP granites which are important in the genesis of the gold deposits	Exploration confined to the zone of contact metamorphism. Defined by mapping, geophysics and gravity Soil geochemistry along mapped structures, . followed by drilling	Cannard and Pease, 1990; Ormsby et al. 1998; Alexander et al. 1990; Quick, 1994; Ahmad et al, 1993; Nicholson and Eupene, 1990; Stuart-Smith et al, 1993; Partington and McNaughton, 1997
Tanami	6.00	Deposits hosted by sequences of basalt, clastic sediments and iron-formations, which wrap around syn- to post-orogenic granites	All deposits are related to regional scale anticlines and associated shear zones. Mineralization related to strike-slip faults and thrusts	Quartz stockwork veins, mineralised shear zones and stratabound replacement horizons	Gold, quartz, pyrite, arsenopyrite and minor chalcopyrite and sphalerite	Quartz carbonate, chlorite, sericite, sulfide and ankerite	Au, SiO_2, CO_2, Ag, As, and Cu	Mineralization was synchronous to late in deformation sequence with a strong association with regional anticlines. Gold synchronous with granite intrusion.	Low fluid salinities and low CO_2. Fluid interpreted to be magmatic to metamorphic. Deposition temperature 300°C	$\delta^{18}O$ range 9 to 11.4 ‰ and $\delta^{34}S$ around 12 ‰	Gold deposited between 3-6 km at 300°C. Fluids interpreted to be sourced from either the granite or its metamorphic aureole.	Mapping and magnetic interpretation followed by vacuum drilling. Geochemistry used to define drilling	Plumb, 1990; Smith et al., 1998; Tunks and Marsh, 1998; Mayer 1990
Telfer	10.40	Yeneena Group comprising siltstone, sandstone, and quartzite	Mineralization localized in two en echelon asymmetric doubly plunging anticlines that define the north-trending Telfer Dome	Ore contained in a series of vertically stacked stratiform Au-Cu horizons linked by stock work and sheeted veins	Gold, quartz, banded to massive pyrite, ± chalcopyrite, pyrrhotite, galena, and sphalerite	Quartz, chlorite, sericite, albite, carbonate, tourmaline	Au, SiO_2, CO_2, Cu, As, Bi, Co, Ni, Ag, La, Ce, Y, B, Pb, Zn, W and Mo	Gold mineralization was late in the deformation sequence contemporaneous with the intrusion of regional granites.	Moderate to high salinity H_2O–NaCl ± CO_2 ± CH_4 fluid. Pressure 2 kbar with two fluid temperatures at 250°C and 450°C. Igneous, metamorphic and meteoric fluids implicated.	$\delta^{34}S$ range 2 to 10 ‰, $\delta^{13}C$ range –3 to 3 ‰ and $\delta^{18}O$ range 13 to 18 ‰. These implicate sediments as source of metals with input from granite and metamorphic fluid.	Gold related to granite intrusion as heat source that set up hydrothermal systems to leach metals and S from sediments. Fluid then channeled to form replacement reefs	Rock-chip sampling followed by surface mapping underground mapping and deep drilling	Goellnicht et al., 1989; Dimo, 1990; Rowins et al., 1998
Homestake	57.00	Hosted within quartz-veined, sulfide-rich segments of carbonate facies iron-formation in sequence of calcareous, pelitic rocks	Host rocks complexly deformed by a series of tight isoclinal sheath folds, and synchronous, extensive ductile and brittle-ductile shearing	Strata-bound replacement of iron-formation	Gold, quartz, pyrrhotite, arsenopyrite and minor pyrite	Quartz, siderite, chlorite, carbonate and sulfide	Au, Ag and As.	Intrusion of granite northeast of mine at ca 1720Ma post-dated regional metamorphism and was contemporaneous with late brittle deformation and gold mineralization	H_2O and CO_2 ± CH_4-rich fluid	$\delta^{34}S$ range 5.6 to 9.8 ‰, δD –112 to –56 ‰, and $\delta^{18}O$ of 12.3 to 13.8 ‰	Timing of gold mineralization in relation to granite intrusion and isotope data suggest that mineralization is related to metamorphic dewatering during contact metamorphism	Detailed mapping of the structure and underground diamond drilling	Caddey et al., 1991; Rye and Rye, 1974; Rye et al., 1974

TABLE 2. (Cont.)

Deposit	Gold (Moz)	Geologic setting	Structural setting	Style	Ore minerals	Alteration	Element association	Temporal relationships	Fluid inclusion data	Isotope data	Genetic model	Exploration	Main references
Birimian	91.49	Host rocks are Birimian greenstones consisting of ultramafic rocks, basalts, chert and shale, felsic volcanic rocks, graywacke, and conglomerate	Mineralization related to regional scale faults, including strike-slip and thrust faults and regional scale folds	Gold deposits are structurally controlled as veins, vein stockwork or alteration in shear zones	Gold, pyrite, arsenopyrite, pyrrhotite, galena and sphalerite	Quartz, chlorite, carbonate, tourmaline, sulfide, sericite and albite	Au, SiO_2, CO_2, Ag, As, B, Ba, Mo, Sb and W	Mineralization late in deformation sequence and post-dated calc-alkaline granite intrusion at the end of orogenesis at 2090 Ma. Gold mineralization is interpreted to be just younger than granite intrusion	H_2O-CO_2 fluids with 3 wt % NaCl salinity and temperatures . estimated between 255–310°C	$\delta^{13}C$ range –9.5 to 15.7 ‰, $\delta^{18}O$ range 12.8 to 15.6 ‰, δD range –37 to 53 ‰ and $\delta^{34}S$ range –5.3 to 10.2 ‰. Mixed meta-morphic and magmatic source suggested. CO_2 and S from sediments	Strong structural control on mineralization during late metamorphism possibly related to granite intrusion. Genesis for these deposits is similar to that proposed for Archaean mesothermal gold deposits	Structural mapping, including geophysical interpretation followed by geochemical sampling, generally soil/laterite. Anomalies are followed up by scout and detailed drilling	Huot et al., 1987; Eisenlohr, 1992; Eisenlohr and Hirdes, 1992; Milesi et al., 1992; Olson et al., 1992; Klemd et al., 1993; Marcoux and Milesi, 1993, Oberthur et al., 1996; 1998; Hirdes et al., 1996; Bourges et al., 1998;Robb et al., 1999
Tarkwaian	?	Host rocks are Tarkwaian conglomerate	Gold along bedding planes, cross-beds and in fractures	Paeoplaers	Gold and heavy minerals	NA	NA	Mineralization derived from and synchronous with Birimian lode gold deposits	Similar inclusions to Birimian deposits	NA	Paleoplacer models similar to Witwatersrand.	Mapping, including geophysical interpretation followed by geochemical sampling, generally soil/laterite. Anomalies are followed up by scout and detailed drilling	Huot et al., 1987; Klemd et al., 1993
Trans Hudson	1.00	Host rocks include graywacke, basalt, felsic volcanic, gabbro, conglomerate, diorite, and granite	Jogs or splays in regional scale brittle shear zones are the main structural controls on mineralization	Veins or altered shear zones	Gold, quartz, pyrite, arsenopyrite, galena and chalcopyrite	Quartz, carbonate, chlorite, albite, sericite, tourmaline and sulfide	Au, SiO_2, CO_2, Ag, As, Mo, Bi and Sb	Gold late in the tectonic sequence with alteration overprinting regional metamorphism, deformation and contact metamorphism at about 1780 Ma	Fluids dominated by low salinity H_2O-CO_2-NaCl with mineralization at 360°C and 2 kb	Mixed metamorphic magmatic source. $\delta^{18}O$ range 9.9 to ‰, $\delta^{34}S$ range 2.8 to 5.5 ‰ and $\delta^{13}C$ range –3.8 to 7.3‰	Mineralization . postdates all documented igneous events. Devolatilization reactions during metamorphism similar to other mesothermal gold deposits suggested genesis of other meothermal gold deposits	Historic prospecting. Modern exploration followed up old discoveries with diamond drilling	Ibrahim and Kyser, 1991; Ansdell and Kyser, 1992; Field et al. 1998
Pilgrim's Rest	1.80	Hosted in the Transvaal Sequence comprising . carbonaceous shale, dolomite and sandstone	Mineralization is stratiform controlled by bedding parallel shearing	Shallow dipping bedding parallel mineralized shear zones. Bedding discordant veins and stockwork zones link the ore zones	Gold, quartz, pyrite, arsenopyrite ± bismuth, chalcopyrite, galena and sphalerite	Quartz, pyrite, sericite and carbonate.	Au, SiO_2, CO_2, Cu, As, Bi, Co, Ni, Ag, La, Ce, Y, B, Pb, Zn, W and Mo.	Gold mineralization was synchronous with late duplex thrusting believed. to be regionally linked to the intrusion of the Bushveld complex	Fluid inclusion data suggest major input from a magmatic fluid	Stable isotope data implicates a magmatic fluid	Gold related to granite intrusion. Fluid then channeled by duplex thrusts to form replacement reefs.	Historic prospecting. Modern exploration followed up old discoveries with diamond drilling.	Harley and Charlesworth, 1992; Boer et al., 1993

TABLE 2. (Cont.)

Deposit	Gold (Moz)	Geologic setting	Structural setting	Style	Ore minerals	Alteration	Element association	Temporal relationships	Fluid inclusion data	Isotope data	Genetic model	Exploration	Main references
Guyanian Craton	4.15	Barama-Mazaruni Supergroup consisting of Quartz feldspar porphyry, rhyolite dykes, andesite and basalt volcanics	Mineralization related to regional scale faults, including strike-slip and thrust faults and regional scale folds	Brittle ductile structures with laminated, crack-seal, fracture fill and breccia veins	Gold, pyrite, arsenopyrite, pyrrhotite, galena and sphalerite	Quartz, chlorite, carbonate, tourmaline, sulfide, sericite and albite	Au, SiO_2, CO_2, Ag, As, B, Ba, Mo, Sb and W	Late orogenic, related to Trans-Amazonian orogeny with close spatial relationship to felsic intrusives. Gold deposited in early reactivated structures	Fluid inclusion data suggest shallow level of deposition and temperatures estimated between 127–266°C	δD and $\delta^{18}O$ both plot outside metamorphic and magmatic box consistent with seawater influx	Strong structural control on mineralization during late metamorphism possibly related to granite intrusion. Genesis for these deposits is similar to that proposed for high-level Archaean mesothermal deposits	Structural mapping, including geophysical interpretation followed by geochemical sampling, generally soil/laterite. Anomalies are followed up by scout and detailed drilling	Marcoux and Milesi, 1993; Voicu et al., 1999
Brasilia fold belt gold deposits	NA	Late Proterozoic mobile belt consisting of metamorphosed siltstone, sandstone, chert and shale	Regional northeast-trending gently dipping reverse shear zones	Boudinaged quartz veins in imbricate reverse shear zones	Gold, pyrite, chalcopyrite, galena and sphalerite ± arsenopyrite	Quartz, sericite, sulfide and carbonate	Au, SiO_2, CO_2, K_2O, As, and Pb	Gold mineralization was synchronous with late duplex thrusting	Low salinity, < 7 wt % NaCl, moderately dense H_2O-CO_2 ± CH_4 fluids at 300–375°C at 1.5 to 3 kb	No published data	Ore deposition occurred during thrusting, prograde metamorphism generated gold-bearing fluids, which were focussed along thrusts higher in the sequence	Historic prospecting. Modern exploration in 1980s but data not released.	Hagemann et al. 1992
Carajás, Brazil	> 20	Cu-Au and Cu-Zn deposits hosted by Archean metagraywacke, amphibolite and iron formation	Little data	Lenticular to dike-like metasomatic ironstones and breccia bodies; selective replacements of Fe-rich metasediments	Chalcopyrite, bornite, chalcocite, gold, molybdenite, sphalerite, pyrrhotite, uraninite, magnetite, fluorite, fayalite, almandine, grunerite	Quartz, albite, biotite, K feldspar, chlorite, pyrosmalite, stilpnomelane, calcite, epidote and sericite	Cu,Au, Zn, Co, Mo, U, LREE, F, and CO_2	Limited constraints on tectonic setting, structural controls, absolute age, and granite relationships	Moderate- to high-salinity (up to ca 50 wt % NaCl equiv), 280-500°C, Na-Ca-K-Fe brine and CO_2 + CH_4 inclusions. 3.7–1.4 kb	No data	Hydrothermal replacement with l inks to Proterozoic Fe oxide Cu-Au deposits proposed	Airborne magnetic and radiometric prospecting. Drilling through laterite	Winter, 1994; Lindenmayer and Teixeira, 1999; Requia and Fontboté, 1999
Tennant Creek	4.71	Au-Cu deposits hosted by early Proterozoic siliciclastic sedimentary rocks	Parasitic fold hinges and shear zones. Some stratigraphic localization	Mineralization selectively affected some of about 650 structurally-controlled ironstones	Chalcopyrite, gold, pyrite, pyrrhotite, magnetite, hematite, and bismuthinite	Chlorite, talc, muscovite, dolomite, siderite, quartz, magnetite, hematite, and pyrite	Cu, Au, Bi, Pb, Zn Co, Se, U, and CO_2	Syn- to late orogenic mineralization broadly synchronous with emplacement of ca. 1850 Ma Tennant Creek granites	Ironstone stage 150–250°C, 10–30wt % NaCl equiv. Au-Cu stage: 300–450°C, 10-50 wt % NaCl equiv. CO_2, CH_4 and N_2 present	$\delta^{34}S$ –6 to +6 % in ores	Two stage systems involving basinal and magmatic fluids. Cu-Au deposition influenced by reaction with ironstones	Historic prospecting. Ironstones are magnetic targets	Wedekind et al., 1989; Rattenbury, 1992; Huston et al., 1993; Khin Zaw et al., 1994; Compston and McDougall, 1994; Davidson and Large, 1998

TABLE 2. (Cont.)

Deposit	Gold (Moz)	Geologic setting	Structural setting	Style	Ore minerals	Alteration	Element association	Temporal relationships	Fluid inclusion data	Isotope data	Genetic model	Exploration	Main references
Cloncurry (Eastern Mount Isa Block)	5.19	Cu-Au deposits hosted by early Proterozoic greenschist to amphibolite grade supracrustal rocks	Ductile-brittle shear and fault -deposits. Some stratabound in carbonaceous host rocks	Replacements, vein stockwork and breccia bodies	Chalcopyrite, gold, magnetite, hematite, pyrite, pyrrhotite, quartz, calcite	Albite, K (Ba) feldspar, quartz, biotite, garnet, magnetite, diopside, hornblende, muscovite, and chlorite	Cu, Au, Ag, Fe, F, P, Co, Ni, As, Se, Ba, LREE, U, and CO_2	Syn- to late-orogenic mineralization broadly synchronous with emplacement of 1550–1500 Ma Williams-Naraku batholiths	Complex and variable high-salinity brine inclusions, 200–500°C coexist with carbonic (CO_2, ± CH_4) inclusions trapped at > 1.5 kb. Dominantly magmatic origin preferred	Fluid $\delta^{18}O$ of 6 to 11‰, δD −20 to −90 ‰	Various physical and chemical ore deposition mechanisms from high salinity aqueocarbonic , magmatic or mixed magmatic-metamorphic fluids	Drainage, soil and rock chip geochemistry. Airborne and ground magnetic and EM prospecting. Drilling concealed basement	Davidson et al., 1989; Rotherham, 1997; Twyerould, 1997; Adshead et al., 1998; Baker, 1998; Laing, 1998; Rotherham et al., 1998; Williams, 1998; 1999; Williams et al., 1999
Olympic Dam	38.6	Cu-Au-U deposit hosted by early Proterozoic granite	Fault controlled breccia complex formed at shallow crustal levels	Breccia, minor veins	Bornite, chalcopyrite, chalcocite, native copper, haematite, magnetite, pyrite, barite, fluorite, pitchblende, coffinite, uraninite, bastnaesite, florencite	Quartz, hematite, sericite, chlorite, and carbonate	Cu, Au,U, Ag, Co, LREE, F, and Ba	Mineralization synchronous with brecciation and ultramafic, mafic and felsic dyke emplacement shortly after consolidation of ca. 1590 Ma granite host	Early very high salinity (up to 70 wt % salts) and high temperature (up to 580°C) brine and carbonic inclusions. 0.5–1 kb. Main phase mineralization at varying temperature (<300°C) and salinity	Magnetite-quartz oxygen isotope pairs suggest early stage temperatures of 400–500°C and fluid $\delta^{18}O$ of 7–10‰. Fluid during main phase mineralization had low $\delta^{18}O$ ranging from −2 to +6 ‰	Mineralization during mixing of - deep-seated brine with surficial fluid.	Drilling coincident gravity and magnetic anomalies on regional l ineament through deep cover	Roberts and Hudson, 1983; Oreskes and Einaudi, 1990; 1992; Reeve et al., 1990 Haynes et al., 1995

suggests a shallow depth of intrusion (Stuart-Smith et al., 1993). The batholith can be subdivided into three separate suites, with the oldest dated at 1835 to 1825 Ma, a transitional suite at 1825 to 1818 Ma, and a younger suite dated at 1800 Ma.

At a regional scale, gold in the Pine Creek geosyncline occurs in linear belts associated with regional ductile structures at or near the greenschist facies brittle-ductile transition. Gold occurs in all rock types except granite, and the high-grade deposits have an association with carbonaceous or iron-rich sediments. Mineralization also has a close spatial association with ca. 1800 Ma reduced leucogranites of the Cullen batholith and their contact metamorphic aureoles (Ahmad et al., 1993; Klominsky et al., 1996; Partington and McNaughton, 1997). Deposit styles can be subdivided into several types (Fig. 3) related to the host structure and to the contrast in host-rock competency and mineralogy. For example, more competent lithologies in turbidite sequences form vein-stockwork deposits (Fig. 3d–f; e.g., Enterprise and Mount Todd, Table 1), whereas those with both contrasting competency and geochemistry form strata-bound vein and replacement deposits (Fig. 3aBd; e.g., Cosmo Howley, Table 1).

The Pine Creek lode gold deposits are spatially related to regional anticlines that formed early above thrust-ramp and duplex structures (Klominsky et al., 1996; Partington and McNaughton, 1997; Fig. 4). Suitable trap sites within these structures appear to have been required, hence the strata-bound nature of some of the gold deposits beneath thick dolerite sills or graywacke units on anticlinal crests (e.g., Fig. 4). The thrusts appear to have acted as channelways for hydrothermal fluids from deep larger structures into anticlines and subsequent trap sites (Klominsky et al., 1996; Partington and McNaughton, 1997). Consequently, the style, and to some extent, the potential size, of the gold deposits depends on the size of the hosting structure and on competency contrasts of particular rock packages. These commonly depend on the stratigraphic position of mineralization and the presence of preexisting structural heterogeneities or alteration such as silicification or production of hornfels due to granite intrusion (Partington and McNaughton, 1997).

The mafic rocks show a marked depletion in elements such as Ca and Mg, and a strong concentration of Au, Si, CO_2, K, Fe, S, Ba, Au, As, Bi, B, Mo, W, and Sb (Table 2). Where gold occurs in sedimentary rocks, the altered host

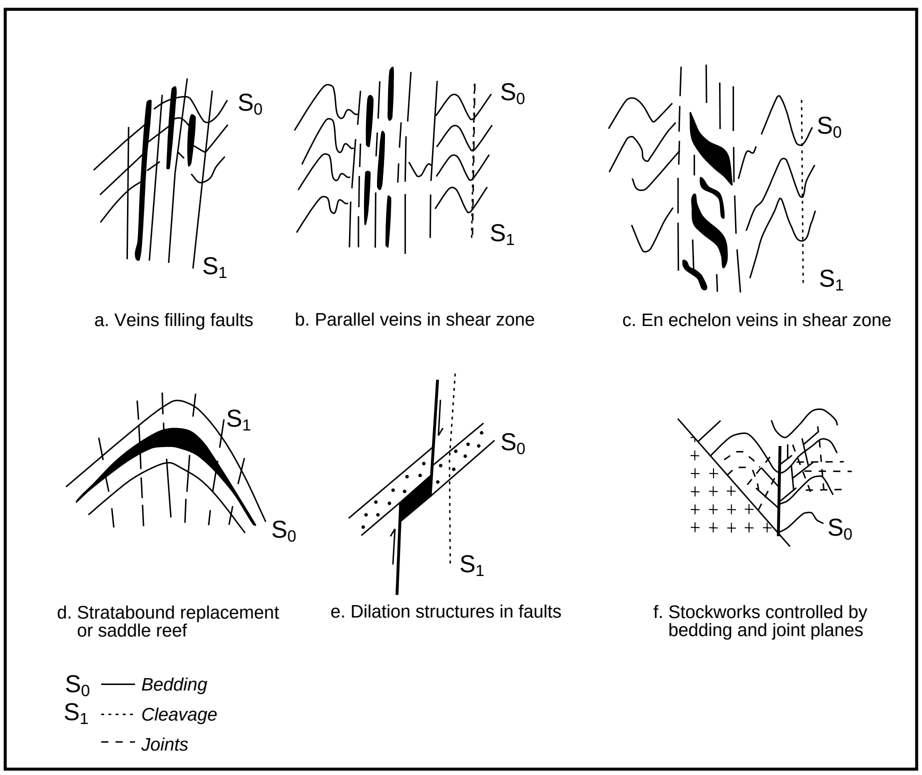

FIG. 3. Style of mineralization in Proterozoic lode gold deposits as listed in Table 1 (based on Stuart-Smith et al., 1993). a. Veins filling brittle faults and fractures, similar to those described from the Tanami region and the cross lodes at Telfer and Sabies-Pilgrim's Rest. b. Quartz veins in ductile shear zones with alteration as described from the Birimian and Trans-Hudson deposits. c. En echelon quartz veins in shear zones as described from the Tanami, Birimian, and Trans-Hudson deposits. d. One of the most common types with either bedding-parallel veins or replacement of reactive beds forming saddle reefs. Examples of this style come from Telfer, Cosmo Howley, Homestake, Woolwonga, and Sabies-Pilgrim's Rest deposits. e and f. Less common types which have been described from the Pine Creek geosyncline and the Tanami region.

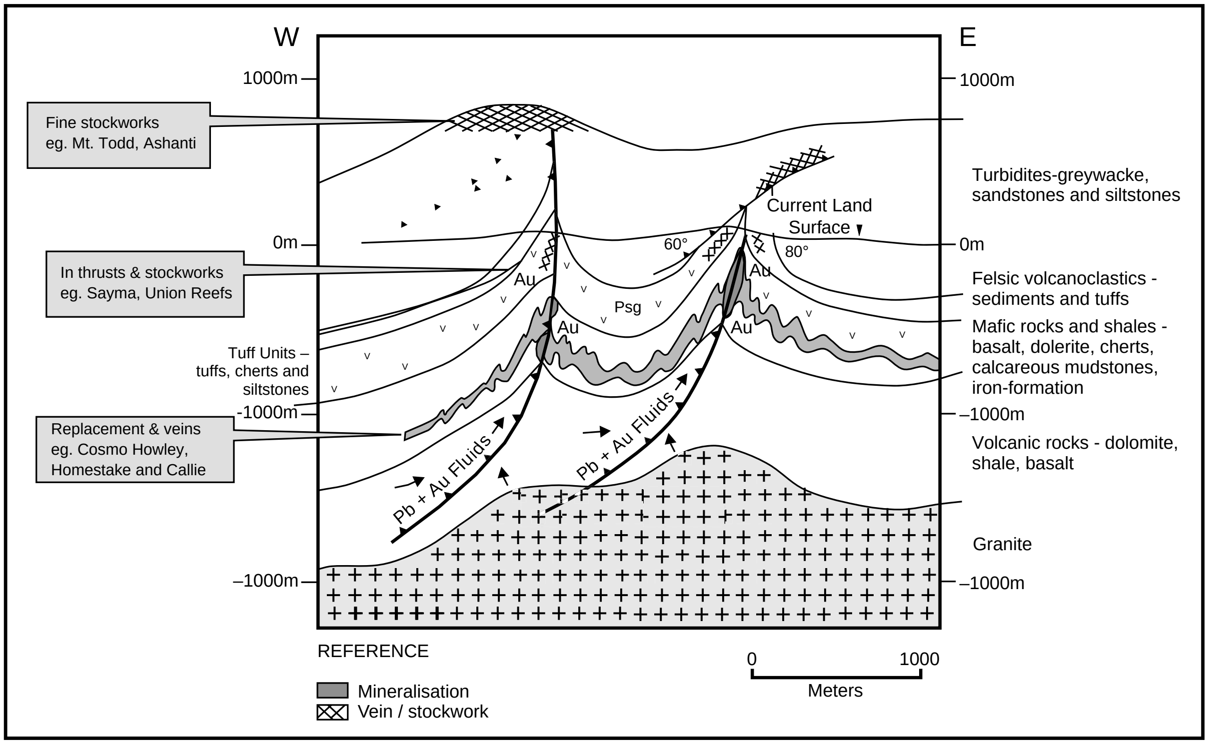

FIG. 4. Summary compilation of all the elements present in Proterozoic lode gold deposits, showing the possible relationship between mineralization in reactivated duplex thrust systems and overlying anticlines. The diagram also shows the possible relationship between contact metamorphic fluids and those derived from granites, with the possible source of the metals coming from either contact metamorphism or regional metamorphism; psg = felsic or mafic volcanic rocks (modified from Partington and McNaughton, 1997).

rock is depleted in K, Rb, and Ba and enriched in Au, Si, CO_2, Fe, Ca, Na, S, As, Sb, W, Bi, Pb, and Zn (Table 2; e.g., Ahmad et al., 1993; Partington and McNaughton, 1997).

Fluid inclusion studies reveal that early quartz veins were deposited from a moderate-salinity, $CO_2 \pm CH_4$-rich fluid at moderate to high temperatures and around 1 kbar in the Pine Creek deposits (Table 2; Ahmad et al., 1993; Partington and McNaughton, 1997). Late fractures, with visible gold, record the passage of a second, higher salinity fluid in which CaCl2 ± $MgCl_2$ was dominant over NaCl, with slightly lower deposition temperatures. It is interpreted that the early higher temperature fluid deposited quartz veins and some sulfides, including arsenopyrite (e.g., Matthai et al., 1995a, b), and that a lower temperature fluid was involved in the introduction of gold and crosscutting vein sulfides. The fluid inclusions suggest fluid contributions from both magmatic and metamorphic sources (Wygralak and Ahmad, 1990; Ahmad et al., 1993).

Wygralak and Ahmad (1990) found that in the Pine Creek geosyncline $\delta^{34}S$ values in sulfides range from 4 to 10 per mil. The δD values in fluid inclusion water range from –57 to +27 per mil, and the calculated ranges of fluid $\delta^{18}O$ values were 5.5 to 10.3 per mil (Table 2). These ranges imply a mixed magmatic-metamorphic source and overlap with other Proterozoic and Archean lode gold deposits (Fig. 5).

Lead isotope studies in the Pine Creek geosyncline by Matthai et al. (1995a), Klominsky et al. (1996), and Partington and McNaughton (1997) show that ore-related sulfides collected from a variety of hydrothermal deposits and the initial ratios from temporally related granites fall on a linear trend. It is clear from these data that many of the gold deposits have similar initial lead, whereas the spatially related granites have a range of initial leads that is different from that of the deposits. This indicates that if the granites contributed lead to the ore fluids, then the contribution was minor. In addition, the relatively homogeneous lead isotope composition for a significant number of deposits and prospects implies coeval mineralization and an unusually homogeneous lead source on a scale of 100 km.

Tanami Desert (Australia)

Significant modern discoveries have recently been made in the Tanami region of central Australia (Fig. 2; Table 1). These include the Granites, the Tanami Corridor mines, and the rich Callie deposit (Mayer, 1990; Plumb, 1990; Smith et al., 1998; Tunks and Marsh, 1998). The gold mineralization in the Tanami region has many similarities to the deposits of the Pine Creek geosyncline since it is structurally controlled, related to reactivated regional folds and thrusts, and associated with iron-rich contact-metamorphosed metasediments (Fig. 3a, b, c, and e). Mineralization occurred late in the

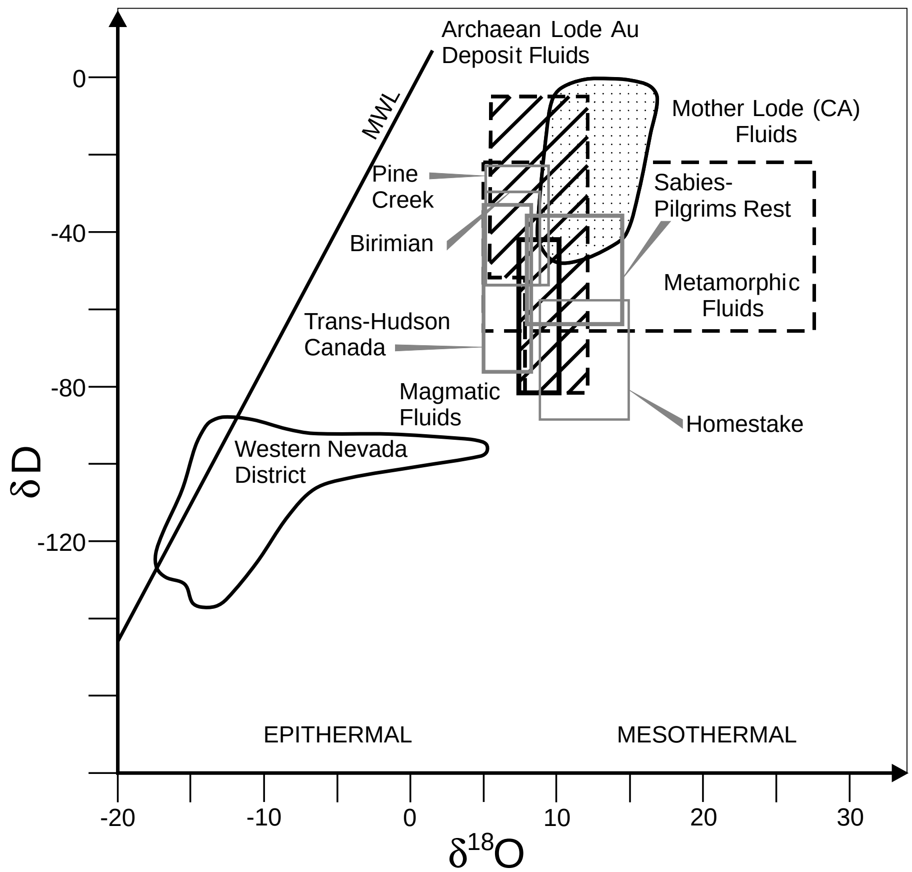

FIG. 5. Summary of oxygen and hydrogen isotope data from Proterozoic lode gold deposits (modified from Boer et al., 1993). Boxes for the range in values for the Pine Creek (Wygralak and Ahmad, 1990; Ahmad et al., 1993; Partington and McNaughton, 1997), Birimian (Oberthur et al., 1996; Robb et al., 1999), Trans-Hudson (Ibrahim and Kyser, 1991; Ansdell and Kyser, 1992), Homestake (Rye and Rye, 1974), and Sabies-Pilgrim's Rest (Boer et al., 1993) deposits have been plotted in relation to Archean deposits, which they overlap. Metamorphic and magmatic fluid boxes are also plotted and the subdivision between epithermal and mesothermal deposits is also shown.

tectonic cycle and is spatially related to syntectonic granites. Most gold occurs in quartz chlorite veins with pyrite and arsenopyrite in shale or basalt, and unlike the Pine Creek geosyncline, some gold is hosted by granite.

Structural controls on mineralization at the Tanami deposits have been described by Smith et al. (1998) and Tunks and Marsh (1998). Mineralization is associated with regional-scale anticlines, similar to that in the Pine Creek geosyncline and at Homestake (Fig. 4; Caddey et al., 1991; Partington and McNaughton, 1997). As in other Proterozoic deposits, late-stage reactivation of early structures controlled mineralization (e.g., at Callie; Smith et al., 1998). Mineralization is related to fractures and veins that progressively increase in density to the center of shear zones (cf. Fig. 3a and b). Vein textures indicate periodic reactivation of the structures.

There are little published fluid inclusion and isotope data from the Tanami gold deposits. However, a study of auriferous quartz veins and breccias from the Tanami gold mine by Tunks and Marsh (1998) suggests that the veins and breccia were deposited from approximately 300°C, low-salinity (5 wt %), low-CO_2 fluids (Table 2). The $\delta^{34}S$ value is around 12 per mil and calculated $\delta^{18}O_{water}$ values are between 9 and 11.4 per mil (Table 2). These values are consistent with a hybrid magmatic and metamorphic fluid which may have been generated during granite emplacement and associated contact metamorphic devolatilization (Tunks and Cooke, 1998).

Telfer (Australia)

The orebodies at Telfer are in many respects similar to those described at the Homestake and Cosmo Howley mines. Gold mineralization was discovered in a stratified gossan in a doubly plunging anticline called the Main Dome (Dimo, 1990). The mine commenced on a reserve of 1 Moz that was subsequently expanded with more than 10 Moz defined to date (Fig. 2; Table 1). Gold occurs in strata-bound horizons hosted by the Telfer Formation (Dimo, 1990), which comprises weakly regionally metamorphosed siltstone, sandstone, and quartzite. Granites spatially associated with Telfer have been dated at 600 Ma. The plutons are postorogenic discordant bodies that intrude regional fold structures as laccolithic thin sheets with subhorizontal tops. Mineral-

ization is generally conformable and strata bound, following the folding in the sedimentary host rocks (cf. Fig. 3d). The host rocks are hornfelsic, forming biotite, cordierite, and andalusite. The regional and thermal metamorphic assemblages were overprinted by the mineralizing event. Gold is associated with pyrite, quartz, minor chalcopyrite, bornite, and chalcocite (Table 2). The sulfides are present as disseminated blebs and euhedral crystals of pyrite that replaced the host rocks.

Mineralization at Telfer is controlled by regional noncoaxial folding that produced the Telfer dome (Fig. 4; Goellnicht et al., 1989; Rowins et al., 1997). This is a variation of the controls described for the Pine Creek geosyncline and is similar to that described for the Sabie-Pilgrim's Rest deposits described by Harley and Charlesworth (1992). The orebodies at Telfer are a series of vertically stacked stratiform to strata-bound lenses controlled by anticlinal hinges that extend over 1,500 m in vertical depth. The lenses are linked by lower grade stockwork vein arrays and sheeted vein sets. The strata-bound nature of the reefs appears to be controlled by bedding-plane slip and replacement of favorable sedimentary horizons (Rowins et al., 1998). Mineralization occurred late in the deformation sequence, and reactivation of early structures was important in localizing gold mineralization (Goellnicht et al., 1989).

Telfer is unusual in that it has a greater concentration of Cu and Co compared to the other Proterozoic lode gold deposits (Rowins et al., 1997, 1998). Alteration associated with gold mineralization at Telfer is distinguished by quartz, sericite, calcite, tourmaline, albite, pyrite, ± chalcopyrite, pyrrhotite, galena, and sphalerite (Table 2; Rowins et al., 1997, 1998).

Detailed fluid inclusion and isotopic studies at Telfer support an epigenetic origin for the fluids at Telfer (Goellnicht et al., 1989; Rowins et al., 1997). The fluids contained H_2O, CO_2, and CH_4 and were moderate to high salinity with between 15 to 54 wt percent NaCl equivalent (Table 2). This is interpreted to reflect their derivation as heated formational brines with some minor input from evolved granite fluids (Rowins et al., 1997, 1998). The $\delta^{34}S$ values from hypogene vein pyrite (–7 to +10 per mil) overlap those of syngenetic sulfides in the carbonaceous host rocks (–23.8 to +11.2 per mil). S/Se ratios are greater than 100,000 in vein pyrite, implying a sedimentary source of S and Se in the ore fluids. Pb isotope compositions of ore-associated sulfides are similar to those of Pb in the host rocks, implying that the Pb contribution from magmatic sources was minor. The $\delta^{13}C$ values in mineralized veins range from –3 to +3 per mil and $\delta^{18}O$ values from 13 to 18 per mil (Table 2). These data are consistent with C derived from dissolution of primary marine carbonate and O comprising a mixture of formational contact metamorphic and subordinate magmatic fluids (Rowins et al., 1998).

Other Proterozoic Lode Gold Deposits

Homestake (U.S.A.)

The Homestake gold mine is the largest example of a Proterozoic lode gold deposit (Fig. 1; Table 1) and has the largest production of any single gold deposit outside South Africa (Caddey et al., 1991). It has operated continuously since its discovery in 1876 and production including reserves total more than 57 Moz. The deposit is hosted within quartz-veined, sulfide-rich segments of a carbonate facies iron-formation, in a sequence of originally calcareous, pelitic to semipelitic and quartzose rocks of the Homestake Formation. These rocks are believed to have been deposited at approximately 2500 Ma (Caddey et al., 1991). The host rocks are complexly deformed by a series of tight to isoclinal sheath folds with synchronous, extensive ductile and brittle-ductile shearing and were subjected to upper greenschist to lower amphibolite facies metamorphism at about 1800 Ma. Intrusion of granite northeast of the mine at approximately 1720 Ma postdated regional metamorphism and appears to have been contemporaneous with later brittle deformation and gold mineralization.

Nine ore ledges or plunging fold structures have produced gold in the mine (cf. Fig. 3a to d). The ore ledges are synclinal folds composed of a series of subordinate anticlines and synclines. The orebodies are relatively undeformed, tabular to pipe-shaped, and developed in dilated segments of late-stage ductile-brittle shears. Gold occurs with quartz, siderite, chlorite, pyrrhotite, arsenopyrite, and minor pyrite (Table 2), largely associated with alteration along shear planes in what has been defined as shear replacement ore (Caddey et al., 1991).

Competency contrasts in the iron-formation are believed to have been the main structural control on mineralization at Homestake (Caddey et al., 1991). These rocks controlled the style of folding and later thrusting. The continuous nature of the iron-formation and increased permeability produced by shearing provided an efficient fluid path along steeply plunging sheath folds and at the steeply dipping interface between the iron-formation and their host rocks. The early structures appear to have been reactivated during the emplacement of the Crook Mountain Granite, and it is at this time that gold mineralization occurred during reverse movement on early shear zones (Caddey et al., 1991).

Pb isotope studies on galenas from the mine record a Precambrian age for the mineralization at approximately 1600 Ma (Rye et al., 1974). The $\delta^{34}S$ values range between 5.6 and 9.8 per mil in pyrrhotite and arsenopyrite associated with mineralization and from 2.7 to 29.8 per mil in sulfides from the host sediments (Table 2). Rye and Rye (1974) believed these values indicated that the sulfur associated with mineralization was derived from the host sediments. The $\delta^{18}O$ values in quartz and carbonate associated with mineralization ranges between 12.2 and 16.1 per mil, dD values for cummingtonite and chlorite had values of –75.6 per mil and –78 per mil, respectively; $\delta^{13}C$ values range between –1.2 to –11.2 per mil (Table 2; Rye and Rye, 1974). Fluid inclusion studies identified CO_2-CH_4-H_2O in quartz veins associated with mineralization and these gave δD values of –56 to –112 per mil (Table 2). A metamorphic origin for the mineralizing fluids is suggested by the fluid inclusion and isotopic data, which overlap Archean isotopic compositions (Fig. 5).

Birimian deposits (West Africa)

The Birimian and Tarkwaian deposits of West Africa make a major contribution to current production from Proterozoic terranes (Tables 1 and 2). Gold is found in Guinea, Mali, Burkina Faso, Ghana, Niger, Cote d'Ivoire, and Senegal (Fig. 1) with a total gold endowment of approximately 91 Moz (Tables 1 and 2). Mineralization occurs in Early Proterozoic greenstone belts (Marcoux and Milesi, 1993; Oberthur et al., 1998), which were deposited on a basement that includes Archean rocks dated at 3300 Ma. The greenstones belong to the Birimian Supergroup, which is overlain by a series of fluviatile metasediments including sandstone, siltstone, shale, and conglomerate of the Tarkwaian Supergroup (Hirdes at al., 1992). The 2180 to 2150 Ma Birimian Supergroup has been divided into lower and upper sequences. The Lower Birimian comprises an assemblage of tuffaceous shale, siltstone, graywacke, and less common chemical sediments. The Upper Birimian rocks are mostly basalts with rare interflow sediments. There is doubt as to the age relationships of the Upper and Lower Birimian and the Tarkwaian rocks. Workers such as Leube et al. (1990) described the Tarkwaian sediments postdating the formation of the Birimian greenstone belts and granite intrusion. Work by Hirdes at al. (1992) suggests that all groups may be considered coeval. Hirdes et al. (1996) summarized the essential points of the Birimian controversy and presented new age data that indicate that the Birimian province comprises two different-aged greenstone belts separated by 50 m.y.

Two ages of granite intrusion are recognized in the Birimian system. The older Belt granites (2180–2125 Ma) occur as small- and medium-sized Itype dioritic to granitic rocks. The younger Basin granites (2125B2080 Ma) occur as large S-type granodioritic to granitic batholiths (Robb et al., 1999). Mineralization ages have been estimated to be 5 to 30 m.y. younger than those of the Basin granites (Oberthur et al., 1998; Robb et al., 1999).

Regional metamorphism in both the Birimian and Tarkwaian rocks reaches greenschist facies and the absence of biotite indicates that metamorphism did not exceed 420°C. The Tarkwaian Group is metamorphosed to the same extent as the Birimian rocks, deformed by the same folding and thrusting events and intruded by the same granites (Hirdes at al., 1992). An early phase of granite intrusion has been recognized which is believed to be coeval with the volcanic belts at 2180 to 2170 Ma (Hirdes et al., 1992). A late tectonic granite phase has also been recognized, which intrudes the Tarkwaian and Birimian rocks, and has been dated at 2115 to 2070 Ma (Hirdes et al., 1992).

Six types of gold deposits have been distinguished in West Africa by Milesi et al. (1992). These are (1) tourmalinized turbidite-hosted gold deposits (e.g., Loulo district), (2) disseminated Au sulfide deposits hosted by tholeiitic basalts, (3) subvolcanic diorites or rhyodacites (e.g., Yaour-Angovia district, Ivory Coast), (4) Tarkwaian gold-bearing conglomerate (e.g., paleoplacers of the Tarkwa district in Ghana), (5) mesothermal auriferous arsenopyrite and Au-bearing quartz vein mineralization (e.g., Ashanti and Presta deposits in Ghana), and (6) mesothermal Au quartz vein deposits with rare polymetallic sulfides (e.g., Poura mine in Burkina Faso).

The lode gold deposits (types 5 and 6, Table 1) are the most important in terms of production and were formed after the Birimian host rocks were deposited, with mineralization overlapping with the intrusion of the Basin granites, regional metamorphism, and deformation (Marcoux and Milesi, 1993; Oberthur et al., 1998; Robb et al., 1999). Gold deposits in the Tarkwaian Group (type 4) have been described as paleoplacers (e.g., mines such as Tarkwa, Teberebie, and Iduapriem), derived from lode gold in the underlying Birimian Supergroup (Huot and Sattran, 1987; Klemd et al., 1993).

Host rocks to gold mineralization generally consist of greenschist facies metamorphosed basic to intermediate volcanics, volcaniclastic sediments shale, and graywacke. Gold is also found in both the Belt and Basin granites, with more than 20 occurrences recorded (Robb et al., 1999). Transpressional reactivation along reverse faults during a late-stage tectonism appears to have controlled gold deposition (cf. Fig. 4; Eisenlohr, 1992; Hirdes at al., 1992; Bourges et al., 1998) in veins and shear zones associated with intense carbonate-quartz vein stockworks and zones of sheeted quartz-carbonate veinlets and breccias (cf. Fig. 3a to c). Pyrite is the principal gold-bearing mineral and occurs disseminated in the alteration halos of the quartz veins. Many of the deposits are refractory beneath the zone of oxidation. The deposits generally occur as tabular and lens-shaped bodies parallel to regional faults (cf. Fig. 3b) that appear to be controlled by carbonaceous units in turbidite sequences. This geometry produces series of imbricate orebodies within duplex fault systems, not dissimilar to the thrust sequence described in the Pine Creek geosyncline (Olson et al., 1992; cf. Partington and McNaughton, 1997). At the Syama mine the orebodies are believed to formed as en echelon ramp faults between the sole and roof faults in a duplex or stacked thrust fault system. Host-rock competency differences involving graywacke, carbonaceous shale, and basalt are believed to have played a major role in localizing faults and hence gold ore shoots. Gold is associated with chlorite, albite, quartz, pyrite, carbonate, sericite, pyrite, arsenopyrite, pyrrhotite, galena, and sphalerite with alteration mineralogy depending on the host rocks (Olson et al., 1992). Gold distribution is closely related to the intensity of alteration and abundance of sulfide. There is commonly an association between gold and arsenopyrite, but free gold is considered the dominant contributor to mineralization in most deposits.

The sulfide and quartz vein ore types described by Leube et al. (1990) in Ghana were studied by Oberthur et al. (1997) who recorded fineness values of the quartz vein ores of 730 to 954 and greater than 910 for the sulfide ores. Whereas Leube et al. (1990) suggested that the sulfide and quartz vein ores were formed by different processes, work by Milesi et al. (1991) and Oberthur et al. (1997) suggested a genetic link between the two types of deposits.

Structural studies by Eisenlohr (1992) showed that both

the Tarkwaian and Birimian sequences were deformed jointly prior to gold mineralization, suggesting that the quartz vein deposits could not have been the source of the paleoplacer mineralization. Davis et al. (1994) carried out detailed age dating of Tarkwaian conglomerates and Birimian volcanogenic sediments and suggested that mesothermal gold mineralization started between the end of the Birimian and the beginning of Tarkwaian sedimentation and continued until after deformation of the Tarkwaian sediments. The Tarkwaian paleoplacer deposits were probably derived from erosion of the earliest gold-bearing quartz veins.

CO_2 extracted from fluid inclusions in gold-bearing quartz veins from the Ashanti belt in Ghana has $\delta^{13}C$ values ranging from –9.9 to –17.0 per mil (Table 2). This is interpreted to reflect extensive interaction of hydrothermal fluid with reduced carbon in Birimian sediments deeper in the crust (Oberthur et al., 1996). The vein quartz has $\delta^{18}O$ values ranging from 12.8 to 15.6 per mil and δD of –37 to –53 per mil (Fig. 5; Table 2). A mixed metamorphic and magmatic fluid is interpreted from these data (Oberthur et al., 1996). Arsenopyrite and cogenetic pyrite generally have d34S values in the range –5.3 to +10.2 per mil, with the source of the S interpreted to be from the Birimian sediments. As noted by Oberthur et al. (1996) the C, O, H, and S isotope data from the Ashanti belt indicate that the mineralizing fluids interacted extensively with the Birimian rocks at deeper crustal levels. Robb et al. (1999) studied gold in Birimian granite and concluded that the hydrothermal fluids associated with mineralization were similar to the fluids described for the vein and sulfide mineralization. They concluded that the fluids were not magmatic and were most likely metamorphic in origin.

Trans-Hudson deposits (Canada)

The 1900 to 1780 Ma Trans-Hudson orogenic deposits of Canada (Fig. 1) are linked temporally to the Homestake deposit and contain approximately 1 Moz of gold. They are restricted to the La Ronge, Glennie, and Flin Flon tectono-stratigraphic terranes. Gold was initially discovered in 1926 and recent exploration discovered examples such as the Rush Lake, Star Lake, Jolu, and Jasper deposits (Field et al., 1998). These are generally small and have produced less that 250,000 oz each. The Contact Lake deposit is the largest documented deposit, with a resource of more than 350,000 oz.

The host rocks are greenstone belts comprising predominantly mafic to felsic metavolcanic rocks (1900–1880 Ma), including basalt, tuff, andesite, dacite, and rhyolite which were are overlain by turbidite sequences (1850–1840 Ma). The volcanic and sedimentary rocks are intruded by syn- to late tectonic gabbro and granite (1890–1835 Ma). All lithological units have been metamorphosed from prehnite-pumpellyite facies to upper greenschist or lower amphibolite facies. It appears that peak thermal conditions were attained locally during granite intrusion (Ansdell and Kyser, 1992).

Gold is generally contained in quartz veins formed in dilational jogs or zones of competency contrast along brittle-ductile shear zones that cut all rock types (cf. Fig. 3a and e). The shear zones are related to a regional compressional deformation event that was coeval with peak metamorphism. These were reactivated under brittle-ductile conditions during postpeak metamorphic uplift. Gold mineralization appears to have been related to this brittle-ductile event and is assumed to have postdated granite intrusion (Ansdell and Kyser, 1992).

Gold-bearing quartz veins are associated with silica carbonate, albite, pyrite, muscovite, quartz, and chlorite alteration, which overprints the regional and contact metamorphism. The veins may also contain varying proportions of muscovite, tourmaline, biotite, chlorite, carbonate, chalcopyrite, sphalerite, bornite, and galena (Table 2). Dating of tourmaline and muscovite associated with gold mineralization by the $^{87}Rb/^{86}Sr$ gave a possible age of mineralization of 1760 ± 9 Ma. This age is confirmed by $^{40}Ar/^{39}Ar$ plateau ages, including an $^{40}Ar/^{39}Ar$ plateau age of 1791 ± 4 Ma from the Laurel Lake deposit that suggests hydrothermal activity occurred periodically over a period of tens of millions of years (Ansdell and Kyser, 1992).

Fluid inclusion data from the Trans-Hudson deposits indicate that the mineralizing fluids were dominantly CO_2-H_2O-NaCl in composition (0.6–14.7 wt % NaCl equiv). Interpreted conditions are 360° to 420°C at about 2 kbars pressure (Ansdell and Kyser, 1992). The $\delta^{18}O$ values of quartz from the gold-bearing quartz veins range from 9.9 to 13.0 per mil whereas tourmaline and chlorite have values of 7.4 and 2.3 per mil, respectively. Estimated $\delta^{18}O$ fluid values range between 5.5 and 8.6 per mil, with δD values of –34 and –50 per mil. The $\delta^{34}S$ values from pyrite gave a range of 2.8 to 5.5 per mil. The isotopic data are in the range of those of other Archean and Proterozoic lode gold deposits (Fig. 5). The O, H, S, C, and Sr isotope compositions of hydrothermal minerals associated with gold mineralization suggest that the mineralizing fluids interacted extensively with Proterozoic metamorphic and igneous rocks similar in composition to those hosting the deposits (Ibrahim and Kyser, 1991; Ansdell and Kyser, 1992).

Sabie-Pilgrim's Rest deposits (South Africa)

The Sabies-Pilgrim's Rest gold field is a north-trending, 110- by 30-km area, located on the eastern margin of the Early Proterozoic Transvaal basin (Fig. 1). Mining commenced in 1872 and reached a peak in the early 1900s. Nine major mines operated around the villages of Pilgrim's Rest and Sabie, and had cumulative production of approximately 4.3 Moz of gold. Mineralization is hosted by dolomite, sandstone, and shale of the Transvaal Sequence. The lower age of mineralization is 2027 ± 31 Ma from $^{87}Rb/^{86}Sr$ analysis of an interpreted pre- or syn-Bushveld dolerite sill (Harley and Charlesworth, 1992) cut by mineralized quartz veins (Boer et al., 1993).

Gold mineralization formed flat reefs up to several kilometers long, consisting of quartz, pyrite, and carbonate in centimeter- to meter-thick veins. The veins are localized along bedding planes and bedding-parallel shear zones (cf. Fig. 3a and d). Bedding-discordant veins and stockwork

vein systems link some of the reefs. The orientation of dilated riedel and conjugate riedel veins are consistent with simple shear deformation associated with thrusting. Small-scale duplexes and shallowly westward-dipping ramp structures and asymmetric folds are developed in the more carbonaceous-rich units.

The veins are associated with silica, pyrite, and sericite alteration. Early mineralization was dominated by pyrite and arsenopyrite, which is fractured and filled in by a second phase of mineralization that deposited gold, native bismuth, bismuthinite, tetrahedrite, sphalerite, and chalcopyrite (Table 2).

Mineralization within the flat reefs is characterized by high Cu/Au ratios of greater than 500. Three types of fluid inclusions have been identified (Boer et al., 1993) consisting of an early high-salinity type (>20 wt % NaCl equiv) and late-stage, low-salinity aqueous inclusions. Homogenization temperatures lie between 200° and 400°C and the inclusions contain variable amounts of carbon dioxide between 10 and 60 mole percent. The average $\delta^{18}O$ values are 14.2 and 16.1 per mil in the steep and flat reefs, respectively. The ^{18}O content progressively decreases with increasing depth and the $\delta^{18}O$ values in individual reefs systems are remarkably homogeneous. The $\delta^{34}S$ values of reef pyrite range from –2.8 to +3.1 per mil. The δD (–37 to –67‰), $\delta^{18}O$ (+6.5 to +15.4‰), and $\delta^{13}C$ (–11.7 to –1.4‰) values from fluid inclusions coincide predominantly with values of primary magmatic water. Mineralization is interpreted to have occurred at temperatures around 320°C, at pressures in the range 2.2 to 2.5 kbars, and at depths of 7 to 8 km. The hydrogen and oxygen isotope data overlap with those of other Archean and Proterozoic lode gold deposits (Fig. 5) and do not discriminate between magmatic and metamorphic fluid sources (Boer et al., 1993).

South American lode gold deposits

Proterozoic lode gold deposits in South America have been described from the Guyanan craton (Fig. 1; Marcoux and Milesi, 1993; Voicu et al., 1999) and Brazil (Fig. 1; Hagemann et al., 1992). The deposits in the Guyanan craton are located in Proterozoic granite-greenstone belts and are hosted by felsic and mafic volcanic rocks, quartz porphyry, siltstone, sandstone, shale, graywacke, conglomerate, and granite (Marcoux and Milesi, 1993; Voicu et al., 1999). Synvolcanic porphyry stocks at the Omai mine in Guyana have been dated using U-Pb zircon ages of 2120 ± 2 Ma (Voicu et al., 1999). Intrusive quartz monzonite from the same mine gave zircon U-Pb ages of 2094 ± 6 Ma and hydrothermal rutile and titanite from gold-bearing quartz veins gave an age of 2001 ± 4 Ma (Voicu et al., 1999). Five galenas from the Sain-Pierre, Adieu Vat, and Loulouie mines gave an average model age of 2014 Ma (Marcoux and Milesi, 1993).

An early deformation event is distinguished in the granite greenstone belts by isoclinal folding and thrusting, and this was accompanied by regional metamorphism to mid-greenschist facies. Granite intrusion postdated the deformation and regional metamorphism. Gold mineralization appears to be related to a late brittle phase of deformation that reactivated earlier structures. Mineralization typically occurred in vein systems that are brecciated, laminated, and/or have well-developed crack and seal textures.

Gold mineralization is late in the tectonic sequence, associated with quartz, carbonate, muscovite, fuchsite, tourmaline, and chlorite alteration that overprints the regional metamorphism. Gold is mainly contained in pyrite with accessory pyrrhotite, chalcopyrite, galena, stibnite, tellurides, sphalerite, molybdenite, and bismuthinite (Table 2).

At Omai, scheelite in mineralized quartz veins has $\delta^{18}O$ values between 2.8 and 4.3 per mil. Oxygen isotopes measured in vein quartz vary between 13.2 and 14.0 per mil, similar to the $\delta^{18}O$ values of carbonates (13.8‰ for calcite and 14.4‰ for ankerite). The carbon isotopes of carbonates range between 1.7 and 4.7 per mil. The $\delta^{18}O$ values of the mineralizing fluids vary between +5.6 and –2.7 per mil and the δD values between –52 and +18 per mil (Table 2). The isotopic composition of the hydrothermal fluid plots outside both magmatic and metamorphic water boxes (Fig. 5) and is considered to be a Paleoproterozoic equivalent of an epizonal lode gold deposit similar to the Wiluna and Racetrack deposits from the Archean of Western Australia (Voicu et al., 1999).

The Brazilian deposits are located in the Late Proterozoic Brasilia fold belt (700–450 Ma), which comprises a monotonous series of hydrothermally altered phyllites that have been metamorphosed to lower greenschist facies assemblages. The major controls on mineralization are described as northeast-trending and gently northwest-dipping ductile-brittle thrusts (Hagemann et al., 1992). These deposits have only made a minor contribution in terms of production and have no spatial or temporal association with late orogenic tectonism or granite intrusion.

Australian Cu-Au-Fe Deposits

Deposits in this group formed in broad temporal association with phases of granite emplacement. They are hosted by many different rock types, including granites and various supracrustal rocks, and occur in a range of geologic environments from submetamorphic to mid-crustal (upper amphibolite facies) terranes. In some cases (e.g., Olympic Dam; Fig. 2), the host rocks and ore are products of the same tectonothermal event whereas in others there is a large age difference (e.g., >200 m.y. at Ernest Henry, northwest Queensland). Ore formation and associated alteration occurred in a broad temperature range (200°–500°C) and at depths varying from many kilometers in semi-ductile crust (e.g., Cloncurry deposits, northwest Queensland; Williams, 1998) to very shallow (e.g., Olympic Dam, South Australia; Reeve et al., 1990). There is commonly a regional association with sodic (-calcic) alteration (e.g., Williams, 1994; Barton and Johnson, 1996). Cu and Au are the dominant economic components, but the deposits form part of spectrum in which either or both of those elements may occur with significant amounts of Co, Bi, and/or U.

Australian (Fe)-Cu-Au deposits formed in the interval 1850 to 1500 Ma during specific phases of compression

(Figs. 6 and 7). The Tennant Creek deposits formed during the Barramundi orogeny at about 1850 Ma (Compston and McDougall, 1994; Compston, 1995). The main metallogenic event in the Gawler craton was at 1600 to 1580 Ma (Johnson and Cross, 1995) coincident with the Olarian orogeny (e.g., Nutman and Ehlers, 1998), which was responsible for widespread deformation and metamorphism, particularly in the Curnamona craton to the east. The deposits of the Cloncurry district formed in association with the 1550 to 1500 Ma Isan orogeny (Twyerould, 1997; Maccready et al., 1998; Page and Sun, 1998; Perkins and Wyborn, 1998). The Australian cratons are also characterized by rift-related granite intrusion and volcanic activity, including widespread intrusions at about 1740 and 1660 Ma (Figs. 6 and 7). It is notable that there are no known significant Australian Cu-Au deposits of these ages.

Australian (Fe)-Cu-Au deposits have spatial and temporal relationships to granites. These are typically oxidized, magnetite- and/or hematite-bearing metaluminous I-type granites in regional associations that are commonly both alkaline and subalkaline. The granites occur as either bimodal or continuously differentiated series, including primitive rocks such as gabbro, quartz diorite, and quartz monzodiorite together with evolved granites (Stolz and Morrison, 1994; Creaser, 1996; Pollard et al., 1998; Wyborn, 1998). K- and U-rich high heat production granites, commonly with rapakivi textures, are prominent (e.g., Creaser, 1996; Pollard et al., 1998), though sodic granites also occur (Mark et al., 1999; Perring et al., 1999; Mark and Foster, 2000). Radiogenic isotope studies typically implicate a lower crustal source component (Creaser, 1996; Page and Sun, 1998) and the magmas appear to have evolved through both crystal fractionation and hybridization (e.g., Pollard et al., 1998).

The deposits vary considerably in style, reflecting a range of crustal environments, structural settings, and host rocks (e.g., Fig. 7). Strata-bound deposits occur mainly in carbonaceous rocks such as at Greenmount in the Cloncurry district of Australia (Krcmarov and Stewart, 1998). Subeconomic mineralization at the Osborne deposit (Cloncurry) is localized by layered Fe-rich metasediments (Adshead et al., 1998), though the economic lodes are composed of secondary quartz bodies overprinted by a magnetite sulfide association. The small high-grade lodes at Eloise in the Cloncurry district grade to massive sulfide and were formed by selective replacement of rocks, which had been affected by mafic silicate alteration (Baker, 1998). The deposits of the Tennant Creek district are selective partial replacements of small metasomatic ironstones (Wedekind et al., 1989; Huston et al., 1993). Similar processes appear to have occurred in a deeper, more ductile environment at Starra in the Cloncurry district (Rotherham, 1997; Adshead-Bell, 1998).

Vein stockworks and associated breccias dominate in several strata-bound deposits and also in some cases where mineralization was fault controlled such as at Mount Elliott and Great Australia in the Cloncurry district (Cannell and Davidson, 1998; Fortowski and Mccracken, 1998). Other deposits including the large ones at Olympic Dam and Ernest Henry are characterized by a predominance of breccia and minor associated veining in massive host rocks

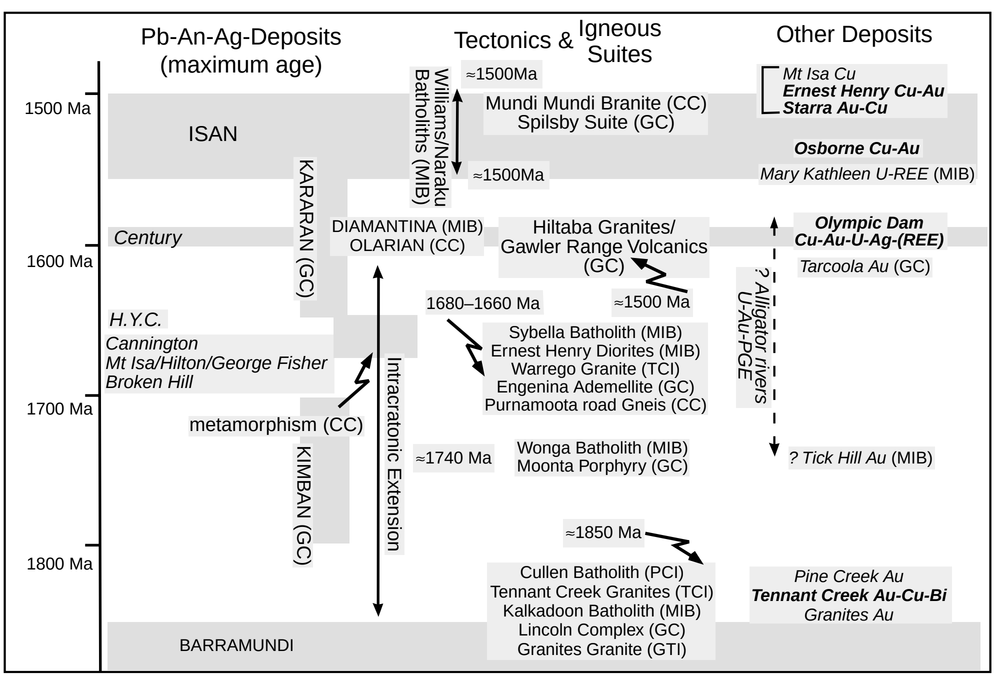

FIG. 6. Summary of major 1900 to 1500 Ma tectonic, magmatic, and Au ± Cu metallogenic events in Australia. Abbreviations: CC = Curnamona craton, GC = Gawler craton, GTI = Granites-Tanami inlier, MIB = Mount Isa block, PCI = Pine Creek inlier, TCI = Tennant Creek inlier.

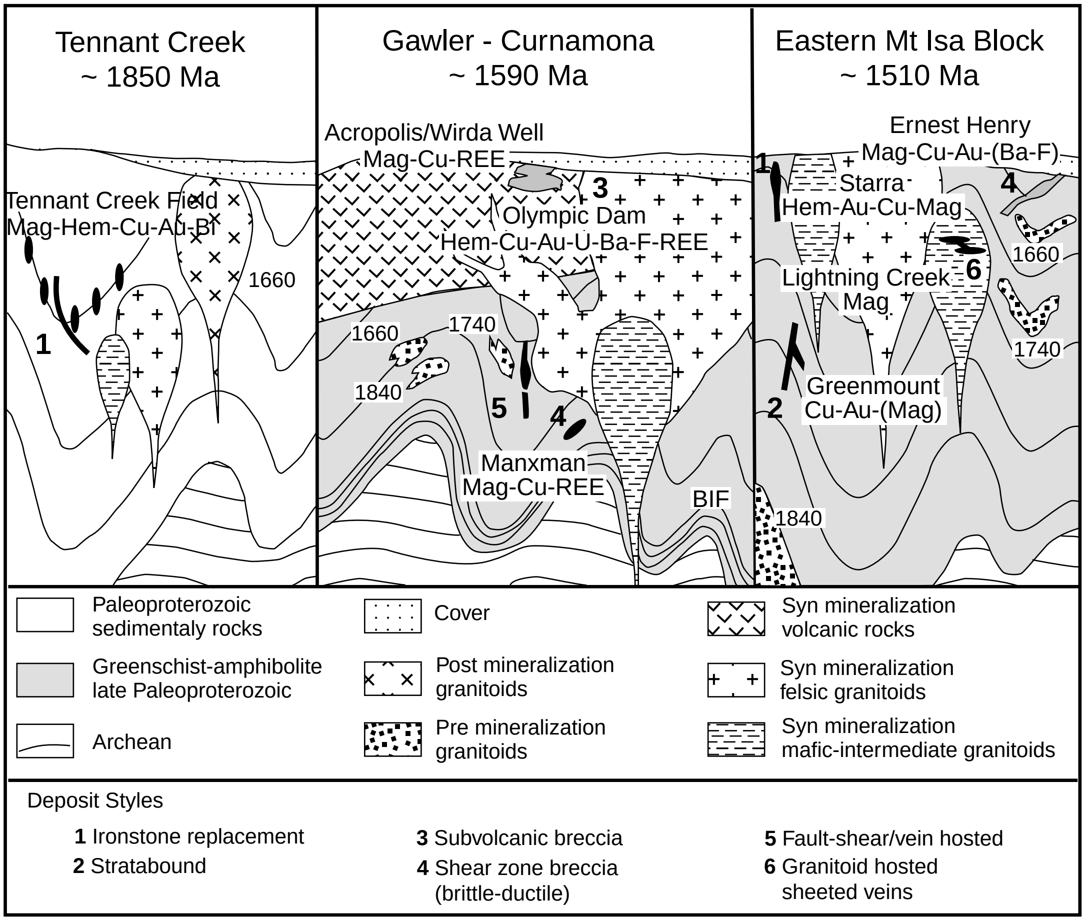

FIG. 7. Schematic representation of geologic settings of Cu-Au-(Fe) deposits and their granitoid associations in the three main Australian Proterozoic districts (adapted from Williams, 1999).

(Reeve et al., 1990; Ryan, 1998). Breccia formation at Ernest Henry appears to have been fault or shear controlled in a brittle-ductile regime and may have had a significant component of matrix formation by replacement processes (Twyerould, 1997). In contrast, partly heterolithic breccias at Olympic Dam, although also localized along a fault system, are interpreted to have formed by multiple phases of explosive activity. This occurred in a shallow structure, possibly representing a diatreme (Oreskes and Einaudi, 1990; Reeve et al., 1990).

The (Fe)-Cu-Au deposits display a range of fault and shear zone controls and are commonly associated with regions of structural complexity, structural intersections, or regionally anomalous structural domains. A relationship to folding is evident in the Tennant Creek district where many of the host ironstones, which also display a degree of stratigraphic control, occur in parasitic hinges and appear to have formed by fluid migration along the dominant cleavage (Wedekind et al., 1989; Rattenbury, 1992). The Cloncurry district is characterized by a complex fault array, with many of the smaller ore deposits concentrated along north-trending regional fault corridors (e.g., Mount Dore fault zone, Levuka trend; Fig. 8). These faults reactivated a dominant, steeply oriented structural grain imposed during the main phase of the Isan orogeny some 30 to 40 m.y. earlier (Adshead-Bell, 1998; Baker and Laing, 1998; Laing, 1998; Perkins and Wyborn, 1998). It is notable, however, that the two largest deposits in the district, Ernest Henry and Osborne, are both located in anomalous structural domains and are also developed in association with a variable dip (Osborne) and strike (Ernest Henry) of bedding and early tectonic fabrics at scales of hundreds of meters to kilometers (Twyerould, 1997; Adshead et al., 1998).

In the higher level deposits the key controlling structures appear to be regional (craton-scale) lineaments as at Olympic Dam (O'Driscoll, 1990), where the Z-shaped plan distribution of brecciation and altered rock (Reeve et al., 1990) is suggestive of a localized dilational-jog geometry.

There is considerable evidence of variable fluid chemistry in (Fe)-Cu-Au deposits from differences in Au/Cu ratios, abundances among the distinctive but inconsistently developed element association of F, Co, As, Mo, Ag, Ba, light REE, W, Bi, and U (Table 1), and variation of alteration assemblages from deposit to deposit. In the case of the Cloncurry

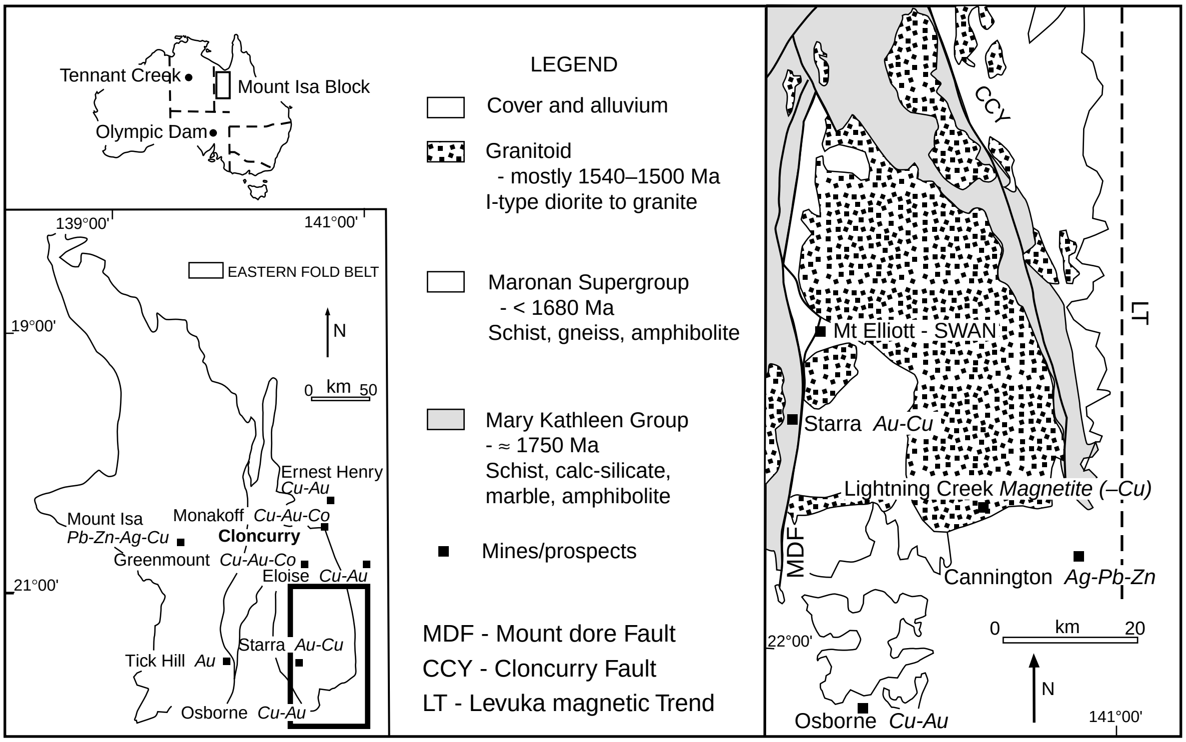

FIG. 8. Outline geology and location of the main ore deposits in the Cloncurry district, northwestern Queensland.

deposits, such variability has been confirmed from direct analysis of fluid inclusions (Williams et al., 1999).

A characteristic feature of (Fe)-Cu-Au districts is regional-scale metasomatism, implying the more localized ore occurrences are products of very large hydrothermal systems (e.g., Williams, 1994; De Jong and Williams, 1995; Mark, 1998). Alteration at both regional and deposit scale was commonly intense (reflecting high fluid to rock ratios and complex saline fluids). Individual ore environments tend to produce similar alteration assemblages in all aluminous rock types, including basic to felsic intrusive and volcanic rocks, pelites, and calc-silicates. Alteration associated specifically with Cu-Au deposits tends to be dominated by Fe-K-(H) phases, particularly in the cases of larger deposits such as Aitik, Salobo, Ernest Henry, and Olympic Dam, which despite their different settings are all characterized by orebodies hosted by even larger volumes of rock in which plagioclase (primary and/or product of earlier Na metasomatism) has been comprehensively destroyed.

Na-(Ca-Fe) alteration is characterized by sodic plagioclase (albite or oligoclase), with K feldspar typically replaced. Other minerals present may include quartz, scapolite, actinolite, diopsidic (acmitic) clinopyroxene, paragonite, titanite, magnetite, hematite, apatite, and carbonates. Assemblages vary from Na-(Fe)-albite ± magnetite; to Na-Ca oligoclase ± actinolite ± clinopyroxene; to Na-Ca-Fe oligoclase ± actinolite ± clinopyroxene ± magnetite.

This type of alteration occurs both in regional-scale alteration systems that may be polyphase and metal depleted (e.g., Cloncurry district; Williams, 1994; De Jong and Williams, 1995; Mark, 1998) and as selvedges to K-Fe alteration in situations where the latter shows a closer paragenetic relationship to Cu-Au mineralization, like several Cloncurry deposits such as Eloise and Starra (Rotherham, 1997; Baker, 1998). It is worth noting that similar types of Na-rich alteration also characterize the wall rocks to Fe oxide-apatite deposits in other parts of the world like Kiruna (Blake, 1992).

High-temperature K-Fe alteration (>400°C) assemblages are similar to those of potassic alteration zones in porphyry copper (± gold) deposits and characterized by biotite, K feldspar, and magnetite. Other minerals that may be present include amphibole, scapolite, and almandine-spessartine garnet. Biotite is the most indicative mineral of high-temperature potassic alteration since K feldspar can also form at low temperatures (see below). The K feldspar may be rich in Ba as at Ernest Henry (Twyerould, 1997). This type of alteration has an important relationship with magnetite-rich Fe oxide-Cu-Au deposits and is extensively developed at Ernest Henry (Twyerould, 1997). Some small- to medium-sized magnetite Cu-Au ore systems such as Osborne and Starra are characterized by restricted (few meter scale) development of high-temperature potassic assemblages adjacent to siliceous orebodies and hydrothermal ironstones, respectively (Rotherham, 1997; Adshead et al., 1998). In some cases (e.g., Ernest Henry) the high-temperature K-Fe alteration appears to be coeval with the main phase of sulfide deposition, though in others (e.g.,

Osborne, Starra) the sulfides are an overprint, formed at lower temperatures.

Variations in fluid chemistry seem to have produced Ca- and/or Mg-rich variants of this type of alteration, with mineralogical characteristics that are transitional to skarn assemblages. Examples include occurrences of tremolite- and anthophyllite-bearing biotite altered rocks at Osborne (Adshead et al., 1998) and the hornblende-biotite lode rocks at Eloise (Baker, 1998). True skarn assemblages characterized by diopside series clinopyroxenes and andradite-grossularite series garnets occur in several Fe oxide-Cu-Au environments, and in one case at least (Mount Elliott, Cloncurry district), form the main host to ore (Fortowski and Mccracken, 1998). These skarns have low Mn contents, a feature they share with conventional Cu and Cu-Au skarn deposits (cf. Meinert, 1993). Associated minerals include amphiboles, scapolite, magnetite, carbonates, and sulfides, which formed by retrograde alteration of skarn.

Stable isotope studies of magnetite and related minerals suggest that the fluid $\delta^{18}O$ value was typically in the range 6 to 10 per mil and δD in the range –40 to –80 per mil (Gow et al., 1994; Baker, 1996; Twyerould, 1997; Rotherham et al., 1998; Mark et al., 1999; cf. Fig. 9).

Low-temperature K-Fe-H-(CO_2) alteration is typified by K feldspar (adularia-microcline), muscovite-(illite-phengite), chlorite, hematite, siderite, and calcite. Such alteration has a close paragenetic association with sulfides in hematite-bearing Fe oxide-Cu-Au systems such as Olympic Dam (Oreskes and Einaudi, 1990; Reeve et al., 1990). However, the orebodies at Olympic Dam have restricted development within huge masses of rock displaying this type of alteration. In other systems, this alteration is independent of mineralization and barren hematite-(sericite) alteration is common in, and adjacent to, many Australian Proterozoic granites. Similar alteration is associated with late overprinting sulfides in some systems with significant high-temperature alteration, such as Starra and Eloise in the Cloncurry district (Rotherham, 1997; Baker, 1998).

Studies at Olympic Dam indicate that the main phase of alteration involved fluids of varying temperatures (<300°C) and salinities, multicomponent solute geochemistry, and low $\delta^{18}O$ values ranging from –2 to +6 per mil (Oreskes and Einaudi, 1990). Cool, low value $\delta^{18}O$ fluids were also present during hematization and Cu mineralization at the Emmie Bluff prospect, 75 km south-southeast of Olympic Dam (Gow et al., 1994). In contrast, Rotherham et al. (1998) estimated the fluid $\delta^{18}O$ value to have been about 9.5 per mil during hematization, carbonation, and Au-Cu mineralization at 200° to 350°C in the Starra deposit in the Cloncurry district (Fig. 9).

Detailed fluid inclusion studies of Cloncurry ore systems have been undertaken at Eloise, Great Australia, Osborne, and Starra (Adshead, 1995; Baker, 1998; Cannell and Davidson, 1998; Rotherham et al., 1998; Williams et al., 1999). Hypersaline fluid inclusions that were entrapped pre- to syn-mineralization have been observed in each case, and these typically coexist with carbonic vapor inclusions. Homogenization temperatures of the hypersaline inclusions range from 100° to 550°C (mostly >300°C in examples paragenetically related to sulfides). Phase behavior, proton probe analyses, and daughter salt assemblages demonstrate that a number of different fluid types were present including various complex high-salinity (up to 40 wt % Cl) Na-K-Ca-Fe-Mn ± Ba and lower salinity Na-Ca brines. Ore deposition occurred from a spectrum of oxidized (hematite-stable, H_2O-CO_2) and reduced (pyrrhotite-stable, H_2O-CH_4 ± CO_2) fluids. Cl/Br ratios of the high-salinity inclusions are consistent with a magmatic origin and quite different to those of the basinal fluids responsible for the similar-aged giant Cu deposit at Mount Isa (Williams et al., 1999, cf. Heinrich et al., 1993)

Fluid inclusion studies of the Tennant Creek systems suggest that the ironstones formed from low-temperature (250°C) and moderate-salinity fluids. The subsequent Au-Cu mineralization was derived from warmer (350°C) and higher salinity fluids with both oxidized and reduced carbonic components (Huston et al., 1993; Khin Zaw et al., 1994). Tennant Creek and Cloncurry ore fluids therefore have many similarities. The Tennant Creek fluids also contain nitrogen and traces of a number of more complex hydrocarbon species. Khin Zhaw et al. (1994) suggested that the Tennant Creek fluids originated as heated basinal brines, but they did not exclude a magmatic contribution. This accords with the suggestion of Huston et al. (1993) that the compositional variation results from varying proportions of oxidized magmatic and reduced basinal brines.

Oreskes and Einaudi (1990) noted that fluid inclusions associated with older magnetite, pyrite, and siderite at Olympic Dam were entrapped at significantly higher temperatures than those of main phase mineralization. Magnetite-quartz oxygen isotope pairs suggest temperatures of 400° to 500°C and fluid $\delta^{18}O$ values of 7 to 10 per mil (i.e., similar to typical ore fluids in the Cloncurry district; cf. Fig. 9). Conan-Davies (1987) described other high-temperature inclusions with very high salinities (40–70 wt % salts), complex Fe-bearing daughter salt assemblages, homogenization temperatures up to 580°C, and phase behavior suggesting entrapment at pressures of 0.5 to 1 kbars. The latter coexist with CO_2-rich inclusions and therefore the early high-temperature fluids at Olympic Dam were similar to the fluids in various other Cu-Au deposits.

Other (Fe)-Cu-Au Deposits

Carajás, Brazil

There is a growing body of opinion that the base metal Au-(Fe) deposits in the Caraj·s region of Brazil including the very large resource at Salobo (Table 2) should be in the general class described here though some uncertainty remains about the age of mineralization (Winter, 1994; Lindenmayer and Teixeira, 1999; Requia and Fontboté, 1999). Archean host rocks underwent a deformation event at about 2555 Ma (Machado et al., 1991), which is interpreted to have affected some of the ore zone rocks at Salobo (Lindemayer and Teixeira, 1999).

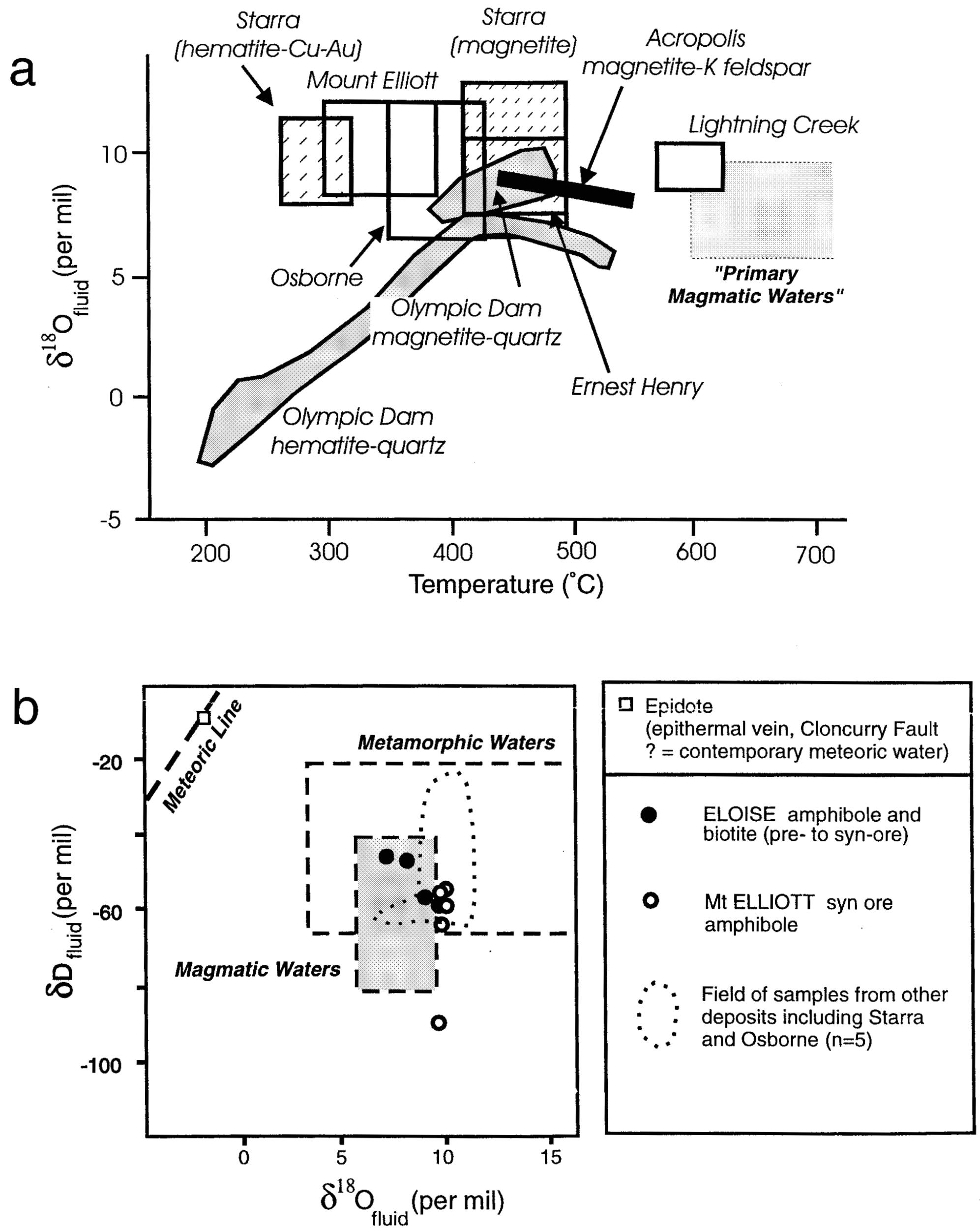

FIG. 9. a. Summary temperature -d18Ofluid diagram for Australian Proterozoic (Fe)-Cu-Au deposits (cf. Oreskes and Einaudi, 1992). Olympic Dam and Acropolis fields based on oxygen isotope equilibration temperatures (Oreskes and Einaudi, 1992). Temperatures for Cloncurry deposits derived from isotope equilibration temperatures, fluid inclusion estimates, and mineral phase relations (Adshead, 1995; Baker, 1996; Twyerould, 1997; Rotherham et al., 1998; Perring et al., 1999; Wang and Williams, submitted). b. Summary d18Ofluid-d18Dfluid diagram for pre- and syn-ore-stage amphiboles and micas in Cloncurry (Fe)-Cu-Au deposits (simplified from Mark et al., 1999).

However, alkali metasomatic events at Salobo are attributed to a Proterozoic metamorphic event associated with the emplacement of 1900 to 1800 Ma anorogenic granites, which on geochemical grounds are a preferred source for ore components including Fe, Cu, light REE, and F (Machado et al., 1991; Lindenmayer et al., 1994a, b). Winter (1994) proposed that Cu-Zn mineralization at the nearby Pojuca deposit occurred after all significant regional metamorphism, implying that it must be no older than these granites.

The Salobo deposit may have formed through wholesale replacement, producing lenticular, cupriferous, Fe-rich rock masses along 2 km of strike extent, with thicknesses exceeding 100 m (Lindenmayer and Teixeira, 1999).

Northern Scandinavia

The Kiruna region of Arctic Sweden is well known for the large Kiirunavaara magnetite-apatite deposit and other iron oxide occurrences (e.g., Hitzman et al., 1992). The area also hosts the Aitik mine (Table 1), some 110 km southeast of Kiruna, which is a major Cu producer. The region is characterized temporally by magmatism, regional alkali metasomatism, and Fe-Cu-Au metallogeny associated with the Svecofennian orogeny (e.g., Witschard, 1984; Cliff et al., 1990; Frietsch et al., 1997).

Aitik ore consists mainly of chalcopyrite-(magnetite-pyrite-pyrrhotite) disseminations and veinlets in biotite-muscovite schist, forming an ore zone 250 to 300 m thick and extending for 2 km along strike (Zweifel, 1976; Wanhainen and Martinsson, 1999). Hornblende-biotite-magnetite-epidote veins are developed peripheral to the main ore zones (Zwiefel, 1976). Near to Kiruna, strata-bound Cu-Au-(magnetite) deposits occur at Pahtohavre where mineralization altered albitized carbonaceous metasediments (Lindblom et al., 1996; Martinsson, 1997); a similar style is developed at Bidjovagge in Norway (Björlykke et al., 1987). Fluid inclusion studies of these strata-bound deposits suggest that the ore fluids were generally similar in character to those of the Cloncurry and Tennant Creek districts in Australia (i.e., saline aqueous carbonic fluids; Ettner et al., 1994; Lindblom et al., 1996).

Great Bear magmatic zone, Northwest Territories,Canada

This region in the Northwest Territories is a fossil continental margin that was active from 1875 to 1855 Ma (Hildebrand, 1986; Gandhi et al., 1998). It is distinguished by magnetite-actinolite-apatite bodies and zones of albitization related to high-level intermediate calc-alkaline plutons (Hildebrand, 1986). The NICO deposit in the southern part of the zone consists of Au-Co-Bi biotite-amphibole-magnetite schists (Goad, 1998). The Sue-Dianne deposit, some 24 km from NICO, is one of several Cu-Ag-mineralized Fe oxide-bearing breccia bodies in the region (Johnson and Hattori, 1994; Goad, 1998). Magnetite-hematite-chalcopyrite-bornite-fluorite occurs as matrix in breccias containing K feldspar-altered felsic volcanic fragments (Goad, 1998) such that the Sue-Dianne breccia is mineralogically intermediate between those of the Ernest Henry and Olympic Dam orebodies in Australia.

St. Francois Mountains, Missouri

The St. Francois terrane in southeast Missouri contains several large iron oxide-(apatite) bodies such as Pilot Knob and Pea Ridge that are mainly hosted by rhyolites (Sims et al., 1987). These belong to an extensive 1480 to 1380 Ma anorogenic granite-volcanic terrane (e.g., Kisvarsanyi and Kisvarsanyi, 1989). The concealed Boss-Bixby deposit is distinct from the other large iron-rich bodies, being essentially an Fe-Cu-Co deposit that lacks apatite. The disseminated breccia matrix and fracture-hosted magnetite-hematite-chalcopyrite-bornite occurs mostly as strongly K feldspar-altered intermediate subvolcanic intrusive rocks (Hagni and Brandom, 1989; Kisvarsanyi, 1989). The main control of mineralization is a ring fracture demarking a central intrusive complex, which itself is located along a regional-scale lineament visible on satellite images (Kisvarsanyi, 1989).

Genetic Models: A Discussion

Proterozoic lode gold deposits

Many Proterozoic lode gold deposits have previously been interpreted to have formed by syngenetic processes, for example, Homestake (Rye and Rye, 1974), Telfer (Turner, 1982), Ashanti (Ntiamoah-Agyakwa, 1979; Leube et al., 1990), and Cosmo Howley (Nicholson and Eupene, 1990). This interpretation was based on the association of gold mineralization with iron-rich sedimentary horizons and the apparent strata-bound nature of the orebodies. However, work in a variety of deposits has shown that syngenetic models are invalid since mineralization postdates not only the deformed volcanic-sedimentary host sequences but also the postpeak regional and contact metamorphism and granite intrusion (e.g., Goellnicht et al., 1989; Wall, 1989; Caddey et al., 1991; Harley and Charlesworth, 1992; Marcoux and Milesi, 1993; Partington et al., 1994; Matthai et al., 1995a, b; Klominsky et al., 1996; Oberthur et al., 1998; Robb et al., 1999). Because of this, there has been a fundamental shift to models similar to those suggested for Late Archean lode gold deposits (e.g., Eisenlohr et al., 1989; Groves and Ho, 1990; Partington, 1990; Groves, 1993; Cassidy et al., 1998; Wit and Vanderhor, 1998; Robb et al., 1999), although in Australia there is more emphasis placed on the importance of the role of highly fractionated reduced I-type granites in the mineralization process (Matthai et al., 1995a; Klominsky et al., 1996; Partington and McNaughton, 1997; Rowins et al., 1997; Wyborn, 1998; Wyborn et al., 1998; McLaren et al., 1999).

Many of the major Proterozoic lode gold deposits described above have evidence that mineralization was controlled by reactivated structures (Table 2). The preexisting structural sequence in Proterozoic terranes can be regarded as a fault-fracture mesh that provided conduits for large-scale hydrothermal fluid flow as described by Sibson (1996). Compressional tectonic structures appear to be the main regional controls of these deposits, which may be because thin-skinned thrust tectonics tended to dominate in Proterozoic terranes (Table 2). In these tectonic regimes, high stress areas are concentrated in the overturned steep limbs of buckle folds, which occur above regularly spaced regional duplex structures, as suggested in experiments by Liu and Dixon (1995). This structural geometry results in fluid flow from the high stress areas on the overturned limbs of folds to lower stress areas in hinge zones of anticlines and the hanging wall limbs of the folds (Table 2; cf. Goellnicht et al., 1989; Caddey et al., 1991; Hagemann et al., 1992; Harley and Charlesworth, 1992; Olson et al., 1992; Partington and McNaughton, 1997; Tunks and Cooke, 1998). This type of model not only explains the early regional structural sequence described in many of the deposits but also the possible fluid flow and structural control of Proterozoic lode gold deposits. The presence of orebodies in duplex thrusts

also suggests that gold-bearing fluids were channeled by these structures from a lower structural sequence (e.g., Fig. 4), which is probably dominated by ramp anticlines and thrusts, to an upper sequence dominated by buckle folds (e.g., Fig. 4). Numerical modeling has established that high fluid pressures, significantly above lithostatic pressure, can occur beneath ramping thrust sheets (Smith and Wiltschko, 1996). This fluid pressure is highly dependent on the permeability of the rock package being deformed. The presence of low permeability layers as traps is also important. This can explain, in part, the localization of gold deposits in relatively permeable units (e.g., below dolerite sills, carbonaceous shales, graywacke units, and thick basalt sequences).

Some Proterozoic lode gold deposits are found in close spatial association with granites; some of these in Australia have been classified as high-temperature contact aureole deposits (e.g., Wall, 1989, Wall and Taylor, 1990; Wyborn et al., 1994; Matthai et al., 1995a). For example, in the Pine Creek geosyncline, a model has been described whereby gold mineralization was deposited from high-temperature magmatic fluids during granite intrusion (Matthai et al., 1995a). In addition, the association with carbonaceous metasedimentary rocks is considered to be important in the process of depositing the gold (Wyborn et al., 1994; Matthai et al., 1995a, b). However, there is evidence in the Pine Creek, Telfer, Birimian, and the Trans-Hudson deposits that the relationship is not that simple. Many gold deposits formed significantly after granite intrusion; isotopic data suggest that metals in the ores are not derived by fluids exsolved from granites during crystallization but from metamorphic fluids formed by dewatering of the host stratigraphy during contact metamorphism or regional metamorphism temporally associated with granite intrusion (e.g., Pine Creek geosyncline, Partington and McNaughton, 1997; Telfer, Rowins et al., 1997; Homestake, Rye et al., 1974; Birimian, Robb et al., 1999). It is clear from work in the Pine Creek geosyncline and at Telfer that although granites may have an important role to play, they are not the sources of the metals forming mineralization. This may be important when considering genetic models for deposits such as Sabie-Pilgrim's Rest where although a magmatic fluid was present, it may not have been a source of metals. In fact, the similarities between the Telfer and Sabie-Pilgrim's Rest deposits argue that a model similar to that proposed for Telfer and the Pine Creek geosyncline may be more appropriate.

The timing of mineralization in all the Proterozoic lode gold deposits is late in the tectonic or orogenic sequence, usually after the main phase of deformation, regional metamorphism, and granite intrusion. This timing and other evidence such as fluid inclusion and isotopic data have led many authors to suggest models for mineralization similar to those that have been proposed for Archean lode gold deposits. Oxygen and hydrogen isotope data where measured for all the Proterozoic lode gold deposits overlap with the magmatic and metamorphic fields (Fig. 5; Table 2). The differences between the isotopic compositions appear to be related to the mixture of metamorphic and magmatic fluid and equilibrium of hydrothermal fluids with the immediate host rocks to mineralization. In the cases where data are available, the isotopic compositions overlap with the field for Archean lode gold deposits (Fig. 5). However, any genetic model for deposits such as those in the Pine Creek geosyncline, Telfer, Sabies-Pilgrim's Rest, or the Tanami region must explain the spatial association of chemically distinct granites with mineralization.

A model has been proposed for some Proterozoic lode gold deposits in northern Australia whereby gold mineralization could be formed above and adjacent to Proterozoic granites by long-lasting hydrothermal systems driven by heat produced by radioactive decay of U, K, and Th in those granites (e.g., Klominsky et al., 1996; Partington and McNaughton, 1997; McLaren et al., 1999). Although initially the amount of heat produced is small, it is generated over a long period and therefore can maintain the temperature of an anomalously hot granite above its surroundings for hundreds of millions of years. This is up to 100 times the cooling period of a normal intrusion (e.g., Webb et al., 1985; Solomon and Heinrich, 1992; McNaughton et al., 1993; Klominsky et al., 1996; Partington and McNaughton, 1997; McLaren et al., 1999). The combination of granite intrusion and long-lived radiogenic heat may allow the development of successively superimposed pulses of hydrothermal activity, and consequently, a variety of hydrothermal mineralization over a considerable period of time (e.g., Klominsky et al., 1996).

The contact metamorphic and the metamorphic dewatering models are variants on the same theme and a combination of either or both of these processes can explain the genesis of Proterozoic lode gold deposits. The feature common to both models is that hydrothermal systems formed from overpressured fluids during continuing movement along structures developed early in the tectonic sequence. These systems developed especially within duplex zones beneath ramp anticlines, strike-slip shear zones, and structures associated with emplacement of late granites (cf. Pine Creek, Tanami, Homestake, and Telfer; Fig. 4). These structures acted as channelways for any hydrothermal fluid in the region, focusing fluids along decreasing pressure gradients into structurally and chemically favorable sites for deposition of gold. Fluids derived from contact metamorphism ± cooling granites, or dewatering of the volcano-sedimentary pile due to regional metamorphism ± cooling granites, increasingly in equilibrium with the host rocks to mineralization, would be focused along reactivated early structures, commonly thrust faults, through relatively oxidized sedimentary rocks comprising the base of the sedimentary sequences. This fluid would be trapped by impermeable rocks in suitable structural sites in some deposits at various levels along regional anticlines and at the margins of domes, thus allowing maximum interaction of the fluid with chemically reactive wall rocks (e.g., Cosmo Howley, Telfer, and Homestake).

There has been little study on the transport and depositional mechanisms of gold. It has generally been assumed that because of the similarities with Late Archean gold

deposits, these processes were the same (Partington, 1990; Caddey et al., 1991; Olson et al., 1992; Partington et al., 1994; Matthai et al., 1995a, b; Oberthur et al., 1996; Partington and McNaughton, 1997).

Cu-Au-(Fe) deposits

Genetic interpretations of these deposits have also seen a recent trend away from syngenetic-chemical sedimentary models, which had been applied in some cases (e.g., Davidson et al., 1989, cf. Rotherham, 1997; Lindenmayer, 1990, cf. Lindenmayer and Teixeira, 1999; Requia and Fontboté, 1999). Prior to underground exploration, the Olympic Dam deposit was interpreted to be hosted by sedimentary breccias, which had been affected by syn- to postdepositional hydrothermal activity (Roberts and Hudson, 1983).

Hitzman et al. (1992) suggested that these deposits were formed in association with extensional tectonics and rifting. However, as outlined above, it seems that they may have been generated in a range of settings including continental margin arcs (Great Bear magmatic zone) and collisional orogenic belts (Australian and Scandinavian examples).

In proposing the existence of these deposits as a distinct class, related to Kiruna-type iron oxide-apatite deposits, Hitzman et al. (1992) recognized their hydrothermal origin and proposed that they formed primarily in shallow (less than 4–6 km) crustal environments related to deeper seated, volatile-rich igneous-related hydrothermal systems. However, results of subsequent studies in the Cloncurry district and northern Scandinavia, including fluid pressure estimates from the densities of carbonic inclusions, recognition of structural styles with strong ductile components, and hornblende geobarometry of temporally associated intrusions, suggest that mineralization in these cases occurred at greater depths (6–>10 km; Ettner et al., 1994; Adshead, 1995; Lindblom et al., 1996; Adshead-Bell, 1998; Baker and Laing, 1998; Pollard et al., 1998; Rotherham et al., 1998; Perring et al., 1999). This implies that deposits of this group may form in high-temperature environments with limited influence from surface-derived fluids.

The origin of sodic-(calcic) alteration in Cu-Au-(Fe) districts remains enigmatic. Studies of regional-scale and ore-related assemblages generally imply formation at elevated temperatures (400°–600°C) and involvement of saline NaCl-$CaCl_2$-rich fluids (Blake, 1992; Ettner et al., 1994; De Jong and Williams, 1995; Lindblom et al., 1996; Mark et al., 1999; Perring et al., 1999). Although Barton and Johnson (1996) suggested that the fluids derived their salinity by interaction with evaporites, others (e.g., Lindblom et al., 1996; Mark et al., 1999; Perring et al., 1999) have interpreted such assemblages as products of alteration by Na-rich magmatic fluids, a conclusion which in these cases is supported by stable isotope data and field relationships.

Another issue relating to regional associations is the nature of the relationship between (Fe)-Cu-Au deposits and the iron oxide-apatite deposits that commonly occur in the same terranes. Progress on this issue is currently hampered by an inconclusive and classic debate over the origin of the magnetite-apatite bodies themselves. Some (e.g., Nyström and Henríquez, 1994, 1995) have argued that many such rocks are derived by melt crystallization whereas others suggest that they are hydrothermal replacements (e.g., Hitzman et al., 1992; Bookstrom, 1995) and possibly formed from fluids that had had little or no interaction with magmas (Barton and Johnson, 1996; Rhodes and Oreskes, 1999). The possibility that certain specialized magma systems might produce both Fe-rP-ich melts and S-poor, high-salinity fluids remains essentially untested. However, this seems worthy of future consideration given the evidence for transitional magmatic hydrothermal conditions in some iron oxide systems in Proterozoic Cu-Au districts (e.g., Blake, 1992; Perring et al., 1999).

The source(s) of ore metals and fluids in (Fe)-Cu-Au deposits is uncertain. Barton and Johnson (1996) suggested that several of the distinctive features of the ore fluids may reflect involvement of evaporite-derived ligands, resulting in high fluid salinities and Cl/S ratios. This provides possible explanations for the concentration of only the least soluble chalcophile elements (Cu, Co) along with other elements that form oxides or oxysalts (Fe, REE, U), and for the extensive sodic alteration. The model allows for the waters to be magmatic, basinal, and/or surficial, whereas ore components may be sourced from magmas and/or by leaching of host rocks during fluid circulation (cf. Williams, 1994). However, it needs to be applied with caution, because there are no known evaporite sequences in several of the Proterozoic districts (e.g., Mid-Continent, U.S.A.; Olympic Dam region, Tennant Creek, Australia).

As described above, geochronology has established the broad temporal relationship of granite emplacement to mineralization; stable isotope studies near Cloncurry suggest that the deep hydrothermal systems there could have involved a large component of magmatic water. Gleason et al. (2000) have shown that the Nd isotope compositions of Fe oxide-P-REE deposits in southeast Missouri (including the Cu-bearing Boss-Bixby deposit) imply that the REE are derived from the coeval igneous rocks. These observations point to the involvement of magmas and/or igneous rocks in the hydrothermal systems but do not clearly resolve the ultimate source(s) of water, chlorine, or ore metals.

The granite-hosted Lightning Creek prospect in the Cloncurry district has so far failed to produce significant results for Cu and Au but has geologic features that may have major metallogenic significance (Perring et al., 1999). The prospect is associated with a 9,000-nT magnetic dipole anomaly occupying some 6 km^2, which when modeled is estimated to contain 1,000 Mt of magnetite. Host rocks consist largely of quartz monzodiorite and monzogranite, both of which contain diorite enclaves. These rocks are cut by a set of texturally complex quartz-plagioclase-rich sills and associated magnetite veins that display a range of magmatic to hydrothermal textures (Perring et al., 1999). Petrochemically, the igneous rocks form a coherent group with the sills apparently reflecting the most fractionated components. The hosts are in part altered to a plagioclase-diopside assemblage that retains a magmatic oxygen isotope signature, and the whole package is believed to reflect magmatic

differentiation and fluid evolution processes. Coexisting hypersaline brine and CO_2 inclusions have been found in high-temperature vein quartz, and proton microprobe analyses show that the former are characterized by very high concentrations of Fe (ca. 10 wt %) and Cu (ca. 1 wt %) (Perring et al., 1999; Williams et al., 1999). Lightning Creek therefore appears to represent a magmatic-dominated system in which a late-stage Fe- and Cu-rich, but implicitly S-poor, fluid was evolved leading to deposition of a large mass of hydrothermal magnetite. Large quantities of Cu may have been released into the surroundings to become incorporated into regional fluid systems.

Support for a direct role of magmas in the generation of components of the ore fluids comes from the association of mineralization with a single age of granites in each of the three main Australian districts, consistent with the possibility that specific magma types may be important. Near Cloncurry it is notable that the Lightning Creek granites (known to have produced Fe- and Cu-rich fluids from extreme differentiates) and intra-ore dikes at Mount Elliott both belong to the shoshonitic Eureka supersuite (Pollard et al., 1998). Further evidence for a link to basic to intermediate magmas comes from the "mafic" Mg-Ca-Fe-(± K)-Ni-Co element associations observed in ore systems like Mount Elliott and Eloise (Baker, 1998; Wang and Williams, in press). The common occurrence of carbonate gangue and CO_2 inclusions point to what may have been a critical role for this gas phase in ore genesis. CO_2-rich magmas are susceptible to fluid phase separation at significantly greater depths than is typical of more orthodox magmatic hydrothermal ore systems (e.g., Cu-Mo-Au porphyries) and may help explain the deep-seated character of the deposits near Cloncurry and in northern Scandinavia.

The giant Olympic Dam deposit is clearly a product of a very different geologic environment, though as previously pointed out, it seems that early-stage mineralization involved fluids that were remarkably similar to those of higher temperature, magnetite-dominated systems. Fluid inclusion and stable isotope evidence from the main phase mineralization (Oreskes and Einaudi, 1990) and other features of the deposit such as chalcopyrite-bornite-chalcocite zoning are consistent with mixing of a cool surficial fluid with a warmer, more saline, deep-seated fluid whose origin is a matter of conjecture (Haynes et al., 1995). Johnson and Mcculloch (1995) showed that Nd isotope data imply a mantle source of REE, and therefore, possibly for other ore components at Olympic Dam. They suggested a genetic link with the abundant intra-ore alkaline mafic and ultramafic dikes.

Fluid inclusion determinations and calculated stable isotope equilibrium temperatures suggest that these deposits formed in different temperature regimes (e.g., 400°–500° C at Ernest Henry, Twyerould, 1997; 200°–350° C at Starra and for main phase mineralization at Olympic Dam, Rotherham et al., 1998; Oreskes and Einaudi, 1990). A large range and combinations of different depositional mechanisms have been proposed including cooling, various types of wall-rock reactions, fluid mixing, and fluid unmixing.

Twyerould (1997) suggested that magnetite and sulfide deposition at Ernest Henry was facilitated by a reaction that promoted the partial consumption of preexisting K feldspar to produce fluorphlogopitic mica. This assumes that coherent depletions of immobile elements in the ore reflect their complete dissolution from the host rock rather than a volume increase and dilution by hydrothermal precipitates. There is good evidence that other wall-rock reaction mechanisms such as fluid reduction by carbonaceous matter (Ettner et al., 1994), magnetite (Huston et al., 1993; Rotherham, 1997), and sulfidation by Fe silicates (Baker, 1998) have contributed to ore deposition in specific cases. CO2-brine phase separation could also have been a powerful sulfide-depositional mechanism and has been suggested to have played a role in several of the Cloncurry deposits (e.g., Adshead, 1995). Mixing of disparate fluids has been invoked as the dominant mechanism of ore formation at Olympic Dam (Haynes et al., 1995) and also proposed to have been important in the Tennant Creek field (Skirrow, 1999). In this context it is interesting to note that two of the common distinctive gangue minerals of Cu-Au-(Fe) deposits, namely barite and fluorite, are common products of mixing systems since the solubilities of their components can be much higher than those of the minerals themselves.

Conclusions and Implications for Future Exploration

Proterozoic lode gold mineralization is structurally controlled in late orogenic brittle-ductile structures. Structural models similar to those proposed for other mesothermal styles of gold mineralization can therefore be applied (e.g., Late Archean lode gold deposits, Groves and Ho, 1990; Groves, 1993; Wit and Vanderhor, 1998; Victorian slate belt, Cox et al., 1995). Some deposits have an association with I-type granites; they are generally highly fractionated and reduced, and in all cases, there is a possible fluid input from both magmatic and metamorphic sources. However, none of the lode gold deposits have evidence for the metals having been sourced from the spatially associated granites.

Some deposits, such as Cosmo Howley, Telfer, and Homestake, can be classified as part of the mesothermal stratiform to strata-bound type of gold deposits in iron-formation or iron-rich metasedimentary rocks (e.g., Murchison province, Western Australia; Northeastern Goldfields province, Western Australia). Deposits in West Africa are more similar to the shear zone style of deposit described from Archean terranes in Western Australia and Canada. Other deposits in the Pine Creek geosyncline and the Tanami region have more in common with slate belt styles of mineralization.

The genesis and structural controls on gold mineralization have some important implications for exploration and may explain why certain structurally defined districts are intensely mineralized. Those areas that have developed duplex thrust fold systems appear to be significantly more mineralized than areas with buckle folding. The presence of shear systems linking anticlines higher in the sequence appear to have provided the ideal fluid-focusing mechanisms to localize gold-bearing fluids.

On a mine scale, there is a clear lithological control on mineralization, with competency contrasts being very important. Certain sedimentary units, such as turbidites and iron-formations or dolerite sills and large basalt flows, provide not only the greatest competency contrasts but also chemical contrasts that appear to have localized the richer and more continuous ore shoots.

Much of the recent exploration success in Proterozoic terranes is due to new exploration techniques resulting from advances in the understanding of the genesis of gold mineralization in Archean terranes as well as new technology, such as low-level geochemical analytical techniques, remote sensing, and image enhancement of geophysical data. The dominant structural control on Proterozoic mineralization means that a good understanding of regional structures is important. Regional mapping complimented with geophysical interpretation is commonly used to define areas for geochemical testing. The apparent association of gold mineralization with late highly fractionated granites in some districts, for example, in the Pine Creek geosyncline, Tanami Desert, and Telfer, can also be used to target on a regional scale. These granites have geochemical and geophysical properties that allow them to be easily identified in the field. The granites usually have distinct and mappable contact aureoles, with at least hornblende hornfels facies rocks present close to the granite contact.

The consistent element association in Proterozoic lode gold deposits means that pathfinder elements such as Ag, Bi, Sb, and more importantly, As can be used in conjunction with low-level gold assays when testing regional structures.

Proterozoic (Fe)-Cu-Au deposits on the other hand formed in a broader range of crustal and tectonic environments and display a great variety of structural and host-rock controls. In many cases they share their syn- to late orogenic timing with the Proterozoic lode gold deposits, though it is notable that economically significant examples of the two groups tend to be concentrated in different terranes. The influence of magmas appears to have been greater in at least some (Fe)-Cu-Au systems and the associated granites are typically oxidized and include both mafic and felsic varieties. There could be a link to the deep-seated process that produced CO_2-rich, low S melts, but a great deal more research is needed to investigate this further. The higher fluid salinities in (Fe)-Cu-Au systems are reflected in huge masses of pervasively altered rock. Although sodic alteration styles are commonly prevalent regionally, the larger ore systems in particular are hosted specifically within substantial bodies of rock that are Na depleted and K-Fe-(H) enriched.

The primary objective in regional exploration for Cu-Au-(Fe) deposits should be the recognition of large structures that were active and dilational during pluton emplacement. Igneous rocks may be favored hosts for large deposits because their massive character may lead to the development of broad zones of brittle fractures. Feldspar-rich rocks can also aid ore deposition because of the chemically reactive nature of this mineral. Remotely acquired geophysical data, especially magnetic surveys, play a significant role in poorly mapped and/or exposed regions (Baker and Laing, 1998). Magnetic data will also identify hydrothermal magnetite bodies, though experience suggests that many of these will be barren, and judicious application of other geophysical (e.g., radiometric K, EM, IP) and geochemical techniques can help prioritize prospects for drilling (e.g., Anderson and Logan, 1992; Brescianini et al., 1992; Webb and Rowston, 1995). It must also be borne in mind that not all economic deposits of this group have a direct magnetite association (e.g., Brescianini et al., 1992). Proterozoic Cu-Au-(Fe) deposits are so diverse in character that a key factor in prospect evaluation must be the development of a good understanding of the geologic characteristics of the particular system under investigation. Alteration mineralogy and physical properties (e.g., magnetite remanence and conductivity) must be properly documented and mapped, and particular attention must be paid to the structural environment. Geologists must be educated about the different characteristics of these deposits, especially with respect to the range of different relationships that ore may have to the distribution of minerals which produce geophysical responses, and to the fact that the large ore systems may contain huge volumes of pervasively altered rock which for the most part lacks economic grade.

In conclusion, the Proterozoic terranes of the world are underexplored compared to other provinces, but with the recognition and understanding of controls on mineralization it is possible that new discoveries in Proterozoic terranes will rival those in the Archean in the future.

Open Questions and Future Work

The current literature suggests that there is general agreement that Proterozoic lode gold deposits are similar in many ways to Archean late orogenic lode gold deposits. In general, Proterozoic lode gold deposits have many characteristics in common. However, in detail, they appear to be quite different. In comparison to their Archean counterparts, there have been very few multidisciplinary studies of individual deposits from which to build a database that would allow a better understanding of their characteristics.

One of the most important issues when considering the genesis of this type of deposit is the timing of metamorphism, granite intrusion, and mineralization. There needs to be more emphasis placed on geochronological studies, dating granites using SHRIMP, and alteration associated with the mineralizing event.

There has also been little work carried out on the role that granites have played in this style of mineralization. There is evidence that granite fluids were involved in forming some of these deposits (e.g., the Pine Creek geosyncline and Telfer), but it is unclear to what extent the granites provided the metals. It may be that the tectonothermal event related to granite intrusion also provided the heat to cause dewatering of the crust leading to gold mineralization. However, if this is so, why are these parts of the Proterozoic so well endowed in gold? More work is required on the relationship between metamorphism, granite intrusion, and mineralization within a tectonic framework as is

being attempted with Archean deposits (e.g., Kerrich and Cassidy, 1994; Groves et al., 1997)

Key contemporary issues in the genesis of (Fe)-Cu-Au deposits include (1) the respective importance of fluid sources (including roles if any of specialized magmas and evaporites) and mechanisms of metal deposition, (2) the origin of the large associated volumes of sodic-altered rocks, and (3) the nature of the relationship to iron oxide-apatite deposits. There are currently a limited number of published radiogenic isotope studies addressing fluid component sources and clearly much more work of this type could be done. However, as pointed out by Gleason et al. (2000), the interpretation of such studies is constrained by the current knowledge of elemental behavior in alteration systems. Advanced microanalytical technologies that allow the direct determination of fluid inclusion contents have the potential to reveal much about these systems, as do microbeam-based geochronological methods. Detailed petrochemical studies of igneous suites are needed to assess the behavior of ore metals and ligands during magma evolution and also to establish whether oxide melt fractionation has occurred. Potentially much of importance could be learned from the developing field of melt inclusion research (cf. Hurai et al., 1998).

The simple fact remains though that many fundamental aspects of the (Fe)-Cu-Au deposits and their regional geologic context remain to be adequately described. To a large extent, this simply reflects the recent discoveries of many of the deposits and the equally recent recognition that they form a diverse but related group. The economic signficance and concentration of these deposits in Australia has contributed to a major focus of research effort (e.g., nine Ph.D. studies on such deposits and their context in the Cloncurry district alone were completed during the 1990s). Current studies in Brazil, Canada, northern Scandinavia, and the United States, many of them founded on deposit description and the gathering of fundamental metallogenic data, will no doubt provide the foundation for a global understanding of their genesis.

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