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Review

A review of Australia's Proterozoic mineral systems and genetic models

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ABSTRACT

Australian Proterozoic rocks host significant mineral resources, some of which are amongst the largest in the world with about 50% of the value of Australian mineral production from iron and gold. Australia's Proterozoic mineral systems reviewed in this contribution include: (1) iron-formations or banded iron-formations (BIFs); (2) orogenic and intrusion-related systems; (3) orthomagmatic ore systems; (4) mineral systems associated with anorogenic magmatism; (5) rift-related stratiform and stratabound sedimentary-hosted; and (6) uranium deposits. These mineral systems formed in intraplate, plate margin, back-arc rift and collisional tectonic settings. The Hamersley Basin is endowed with largest Fe resources in the world, which are time equivalent (ca. 2400 Ma) of the Transvaal Group BIF in South Africa. The origin of BIF and of granular iron-formation (GIF) remains a contentious issue with models invoking subaqueous hydrothermal discharges in lakes and/or ocean basins or in Red Sea type brine pools. In all cases a density and oxic–anoxic stratified system is required to enable precipitation of Fe^{3+}. Orogenic and intrusion-related ore systems are very common in the Proterozoic rocks of Australia, with examples from the Pine Creek, Granites–Tanami and Arunta orogens in the North Australian Craton (NAC), and the Capricorn Orogen in the West Australian Craton (WAC). These deposits reflect collision and accretion events between ca. 1800 and 1790 Ma. Orogenic Au lodes are generally, but not always, temporally associated with granitic rocks, but a genetic relationship remains elusive. Orthomagmatic Ni–Cu–PGE and Fe–Ti–V ore deposits in mafic–ultramafic systems are present in the Halls Creek Orogen (NAC) and the ca. 1080 Ma Giles mafic–ultramafic intrusions in the Musgrave Complex (Paterson Orogen). Mineral systems associated with anorogenic magmatism encompass a wide range of hydrothermal deposits of which the economically most important are the Fe oxide–copper–gold or IOCG ore systems, such as the ca. 1580 Ma world-class Olympic Dam in the South Australian Craton (SAC). In the same group are the Abra Pb–Zn–Ag–Ba(–Cu–Au–W) (Capricorn Orogen) and the world-class Telfer Au–Cu (Paterson Orogen). The latter has been one of the largest Au producers in Australia. During 1100 and 800 Ma alkaline rocks, including carbonatites and diamondiferous lamproites, were emplaced in the NAC, SAC and WAC. The 1180 Ma Argyle lamproite pipe in the NAC is the world's largest diamond producer. Studies elsewhere suggest that these alkaline rocks are the distal expression of mantle plume events. Stratiform and stratabound sedimentary-rock hosted giant and world-class Zn–Pb–Ag sulfide deposits developed between ca. 1700 and 1500 Ma in the McArthur River-Mount Isa and Broken Hill rift systems. These deposits are all hosted in metamorphosed siliciclastics or organic-rich shales and associated with clastic–evaporitic successions and bimodal igneous activity. Conceptual models of ore genesis propose discharge of hydrothermal fluids along major basin faults, syn-sedimentary exhalations of these fluids in oxygen deficient pools and bacterial sulfate reduction in order to produce H_2S and precipitate sulfides. An unusual and large non-sulfide Pb carbonate ore deposit, Magellan, is hosted in clastic rocks of the ca. 1800 Ma Earaheedy Group. The lack of sulfides suggest that the deposit is related to paleoweathering processes, which induced oxidation and mobilization of Pb. Uranium ore systems, apart from U contained in IOCG deposits, include the unconformity stratabound deposits in the Pine Creek Orogen with the world-class Jabiluka as the main representative. We conclude that the several giant and world-class ore systems in Australia's Proterozoic were formed during intraplate tectonothermal and rifting events. Orogenic lodes were formed during collision and accretion of arc terranes that led to the amalgamation of the NAC, SAC and WAC.

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1. Introduction

Precambrian (Archean and Proterozoic) Australia is represented by terranes and tectonic units west of the Phanerozoic "Tasmanides" (Direen and Crawford, 2003). The Archean to Paleo-Mesoproterozoic cratonic framework of Australia consists of the West Australian Craton (WAC), North Australian Craton (NAC), and South Australian Craton (SAC; Myers, 1990; Myers et al., 1996). The WAC and NAC are separated by the Proterozoic Paterson Orogen, the WAC and SAC are separated by the ca. 1300–1100 Ma Albany-Fraser Orogen, with the ca. 1080 Ma Pinjarra Orogen on the western margin of the WAC (Fig. 1). Proterozoic rocks in southeastern Australia and Tasmania are not considered in this review because they are probably fragments of terranes that were accreted to the Australian plate during the Phanerozoic (Myers et al., 1996).

An understanding of the geodynamic evolution of Proterozoic Australia is of paramount importance in order to examine and study the timing of the tectonic events that contributed to the formation of ore deposits and identify those factors that lead to the development of conceptual models of ore systems. Similarly, an understanding of ore-forming processes are essential for successful exploration and the discovery of new deposits. The tectonic models that best explain the geodynamic evolution of Proterozoic Australia include both convergent margin and intraplate tectonics (Tyler et al., 1998; Betts et al., 2002, 2003; Tyler, 2005; Betts and Giles, 2006). Plate tectonic models involving convergent margins are well documented for the Halls Creek Orogen, the Rudall Complex, the Albany-Fraser Orogen, Glenburgh Orogen, Arunta Orogen, and the northern margin of SAC (Li, 2000; Bagas, 2004; Betts et al., 2002; Betts and Giles, 2006 and references cited therein). Intraplate tectonics could be linked to mantle plume dynamics at ca. 1800, ca. 1500, ca. 1080 and ca. 830 and 755 Ma, all of which correspond with major igneous and rift forming events (Pirajno et al., in preparation). Mantle plumes are upwellings of hot mantle material that are typically associated with flood volcanism on the earth's surface, dyke swarms, sills complexes and layered intrusions at depth (Ernst and Buchan, 2003). Although felsic magmatism is usually associated with plate margins above subduction zones, igneous products of mantle plumes also include a variety of silicic rocks, so that the full magmatic range is typically bimodal in composition (Ernst and Buchan, 2003). Examples in Australia of mantle plume-related felsic magmatism include the ca. 1800 Ma granites in the Pine Creek Orogen (Wyborn et al., 1992), the ca. 1590 Hitalba granite suite in the SAC (Betts et al., 2002), and the ca. 1080 Ma felsic volcanic rocks and A-type granites of the Musgrave Complex (Glikson et al., 1996).

The geodynamic evolution of Proterozoic Australia has been discussed in some detail by Myers et al. (1996), Betts et al. (2002), Giles et al. (2004), and Betts and Giles (2006). Tyler (2005) provided an overview of the Proterozoic in Australia, and Veevers (2000) treated in detail the evolution of the Australian plate from the Neoproterozoic through to the Phanerozoic. An excellent volume details the Proterozoic geology, mineralization, and tectonic evolution of the SAC (Drexel et al., 1993).

The geodynamic history of the NAC, WAC and SAC began with the collision of the NAC and WAC, between ca. 1830 and ca. 1760 Ma during the Capricorn–Yapungku–Stafford–Early Strangways orogenies, along the Capricorn–Rudall–Arunta orogenic belts (Bagas, 2004; Betts and Giles, 2006). This also resulted in the inception of intracratonic back-arc basins on the NAC. In the next "snapshot", between ca. 1760 and ca. 1620 Ma, much of the tectonic activity is related to the collision of the Gawler Craton, coming in from Antarctica, with the NAC (Kimban-Late Strangway orogenies). This also produced the inversion of the Leichhardt Superbasin in the Mount Isa Inlier and renewed magmatism in the Arunta Orogen at about ca. 1640 Ma. Using new geochronological data, Şener et al. (2005) suggested that the amalgamation of the NAC, SAC and WAC was completed by ca. 1750–1720 Ma. The next stages saw the development of the McArthur River-Mount Isa, Broken Hill, Georgetown rift systems, accompanied by large-scale bimodal magmatism (e. g. Gawler Range Volcanics and Hiltaba Event in SAC; Giles, 1988). These rift systems are considered as being intracratonic back-arc extensional structures, which were formed as a result of regional and far-field tectonic stresses (Betts et al., 2002). Compressional events with further crustal or plate adjustments are recorded in the Paterson Orogen during the ca. 750 Ma Miles Orogeny and the ca. 550 Ma Paterson–Petermann Orogeny (discussed below).

Finally, anomalous high heat flow is recorded in the SAC and the NAC of central Australia, where average surface heat flow is in the order of ca. 80 mW m^{-2} (McLaren et al., 2003). The source

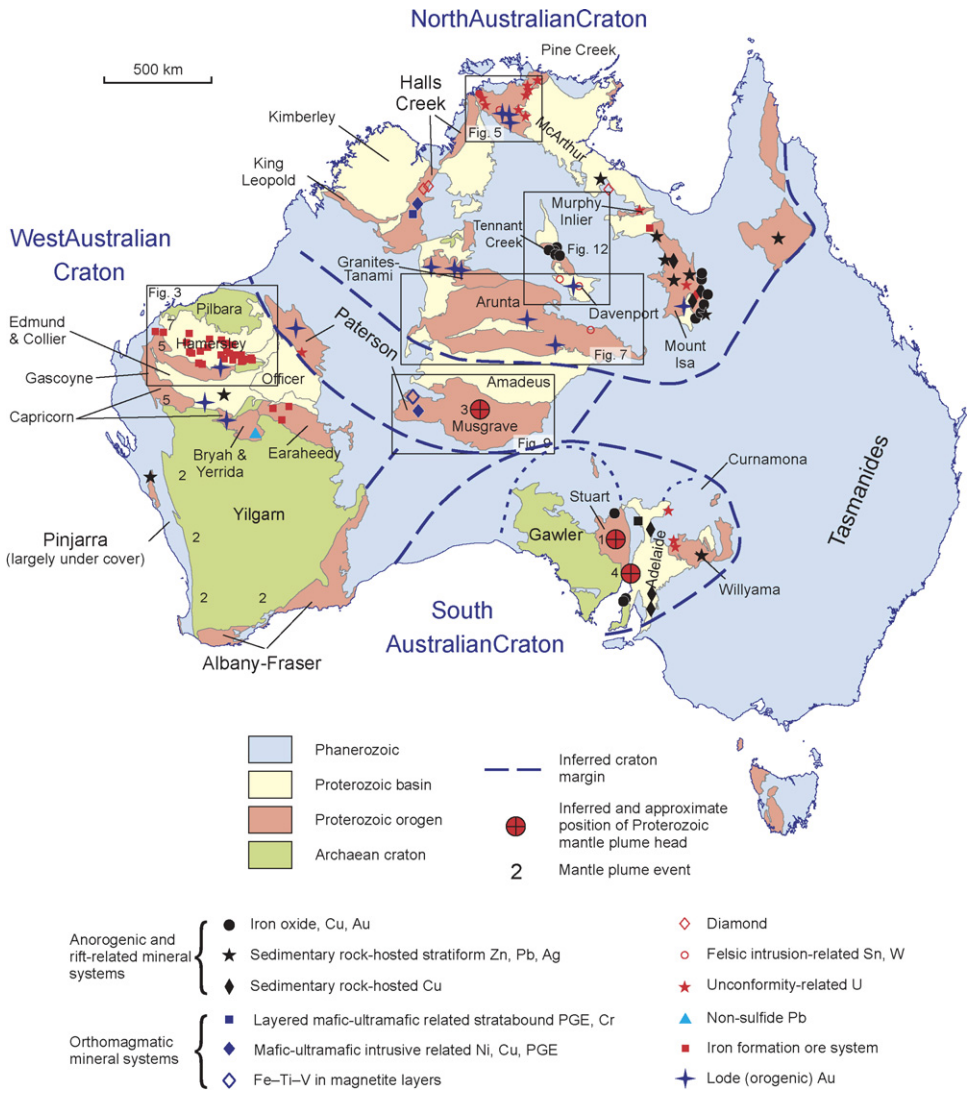


Fig. 1. Simplified map of Australia with Phanerozoic, Proterozoic and Archean terranes and distribution of selected mineral systems of Proterozoic age (based on and modified from Jaques et al., 2002). Proterozoic mantle plume events (see also Table 1): (1) ca. 1580 Ma Hitalba-Gawler Range; (2) ca. Marnda Moorn dyke swarm; (3) ca. 1080 Ma Warakurna; (4) ca. 830 Ma Gairdner dyke swarm; (5) ca. 755 Ma Mundine dyke swarm.

of this anomalous heat flow, which has been termed the Central Australian Heat Flow Province (McLaren et al., 2003), is related to high K, Th and U contents of high heat producing (HHP) Proterozoic granites (see Table 2 in McLaren et al., 2003). One of the outcomes of this anomalous high heat flow is the presence of amagmatic hydrothermal systems, and associated surface manifestations such as hot springs in the Curnamona Province (Brugger et al., 2005).

In Table 1 we present the age of the major mineral systems and associated geological events discussed in this contribution. Fig. 2 illustrates the tectono-thermal and orogenic events that affected some of the key tectonic units of the WAC, NAC and SAC between 1800 and 1500 Ma.

2. Australia's Proterozoic mineral systems

Proterozoic rocks in Australia host considerable mineral resources, some of which are amongst the largest in the world (e.g. Telfer Au, Olympic Dam Fe–Cu–Au–U; McArthur River-Mount Isa Zn–Pb–Ag). About 50% of the value of mineral production comes from iron and gold (Jaques et al., 2002) (Table 2). Solomon et al. (2000) provided a detailed and comprehensive review of the Proterozoic in Australia and associated mineral deposits. Wyborn et al. (1994) and Jaques et al. (2002) have reviewed the Australian Proterozoic mineral systems, and Tyler et al. (1998) conducted similar studies, but focussed on Western Australia.

The concept of a mineral system is analogous to that of a petroleum system (Wyborn et al., 1994; Jaques et al., 2002), but owing to the nature of ore deposits and host rocks, a mineral system is far more diverse and complex. The formation of an ore deposit requires a source of metals, a mode of transport (usually a hydrothermal fluid, but also can be magma) and a site of deposition or accumulation where commodities become concentrated to enable economically viable extraction during a given period. In a broad sense, a mineral system includes geological and geodynamic factors, at all scales, that control the inception, evolution and preservation of ore deposits. Thus, the study of mineral systems must necessarily integrate: (1) local studies on recognised deposits, including such factors as, location of potential accumulation sites, and the physico-chemical processes leading to deposition with; (2) regional-scale studies including geodynamic (tectonic) controls on timing and location of ore deposits (space–time distribution),

Table 1
Summary of selected Proterozoic events in Australia and associated ore systems

Age (Ma)	West Australian Craton	Paterson Orogen (Northwest)	North Australian Craton	Paterson Orogen (Musgrave Complex)	South Australian Craton
ca. 550	Paterson – Petermann – King Leopold Orogeny				
ca. 650		☺ **Telfer Au-Cu** Mt Crofton Granite Miles Orogeny			
ca. 750	⊕ Mundine Dyke Swarm(5)				
ca. 820		☺ **Magnum Au-Cu** ☺ **Kintyre U**			
ca. 830-800		⊕ Gairdner Dyke Swarm and initiation of the Centralian Superbasin(4)			
ca. 1080			⊕ Warrakurna(3) ☺ **Ni-Cu-PGE-Ti-V**		
ca. 1200	⊕ Marnda Moorn Dyke Swarm(2)				
ca. 1350-1260		Magmatism	☺ **Argyle diamonds** Magmatism Isan Orogeny	Musgravian Orogeny Magmatism	
ca. 1560-1500					
ca. 1600 – 1580					☺ **Olympic Dam IOCG** ⊕ Hiltalba and Gawler Range bimodal magmatism(1)
ca. 1600 - 1465	☺ **Abra Pb-Cu-Ba** Rifting and magmatism Mangaroon Orogeny				
ca. 1655 – 1575			☺ **McArthur River, Mt Isa, Earnest Henry Pb-Zn-Cu-Ag** Rifting and bimodal magmatism		
ca. 1715 – 1645			☺ **Jabiluka U Cannigton Pb-Zn-Ag (Sedex)**		☺ **Broken Hill Pb-Zn-Ag (Sedex** Rifitng and bimodal magmatism
ca. 1700-1750			Late Strangways Orogeny		Kimban Orogeny (Phase III)
ca. 1830-1780	☺ **Orogenic Au, GIF** Capricorn Orogeny	Yapungku Orogeny	☺ **Orogenic Au Pine Creek, Sally Malay Ni** Halls Creek, Early Strangways – Stafford, Shoobridge orogenies; Hart Dolerite intrusion		Kimban Orogeny (Phase II)
ca. 1860 – 1845			☺ **Tennant Creek Au-Pb-Zn-Ag-Bi** Magmatism		Kimban Orogeny (Phase I)
ca. 2300 – 2200	☺ **BIF**				

(⊕) Documented Proterozoic mantle plume event (see also Fig. 1): (1) Drexel et al. (1993), (2) Wingate et al. (2005), (3) Wingate et al. (2004), (4) Zhao et al. (1994) and (5) Wingate and Giddins (2000).

terrane-scale physico-chemical processes that determine how ore deposits are formed; and (3) the evolution of magmas and other energy sources and fluids at the scale of mineralizing systems that influence the location of individual deposits.

In the following sections we provide an overview of selected and economically important mineral systems and draw general conclusions, based on literature and our own observations, on their origin with a view to stimulating further research and perhaps provide useful vectors for mineral exploration in areas of green fields. In our discussion of Australia's Proterozoic mineral systems, which is not intended to (and cannot) be exhaustive, we have considered six major categories: (1) iron-formations; (2) orogenic and intrusion-related systems; (3) orthomagmatic mineral systems; (4) mineral systems associated with rifting and anorogenic magmatism; (5) stratiform and stratabound sedimentary-hosted; and (6) uranium deposits (Fig. 1).

2.1. Iron-formations

Iron-formations, usually referred to as banded iron-formation (BIF), are chemical-sedimentary units, consisting of Fe oxides (15–30% hematite and/or magnetite) alternating with chert and silica bands (ca. 50% SiO_2). Clastic iron-formations, due to the re-working of BIF, are known as granular iron-formation (GIF). Traditionally, two main types of BIF are considered: Algoma type and Superior type. The former are intimately associated with volcanic rocks and are smaller in extent, whereas the latter have greater extent and are intimately associated with sedimentary rocks of continental margin successions (Gross, 1983). The largest BIF are those of the Superior type, which were mostly deposited between ca. 2400 and 1800 Ma, and reflect the onset of extensive continental shelf environments and the assembly of supercontinents (Rogers and Santosh, 2004). A comprehensive review of

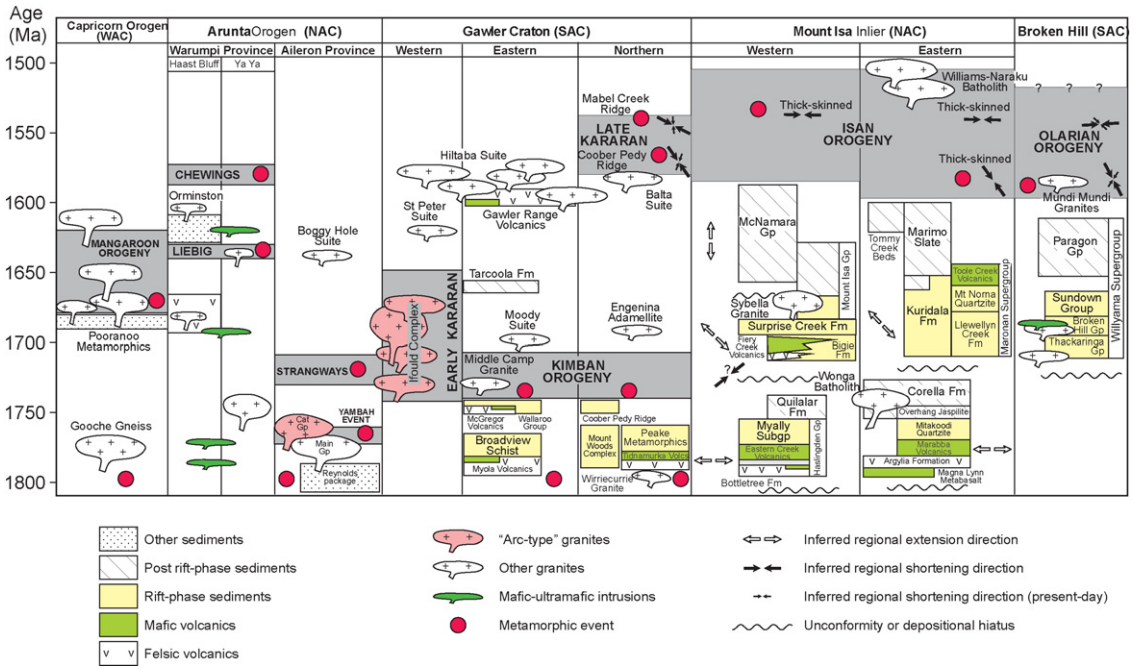


Fig. 2. Space–time diagram of Late Paleoproterozoic to Early Mesoproterozoic events in the selected tectonic units of the WAC, NAC and SAC. Details in text (modified after Sheppard et al., 2005). Mineral systems not shown in this figure, but are listed in Table 1.

iron-formation ore systems can be found in Clout and Simonson (2005).

In Australia, Proterozoic Superior type iron-formations are located in the Hamersley and the Earaheedy basins of Western Australia (Fig. 1; cf. Trendall and Morris, 1983; Morris, 1998; Isley, 1995; Trendall, 2002). BIFs of the Hamersley Basin, in addition to having the largest contained Fe resources in the world, are also the most extensive and best-exposed Precambrian iron-formations known (Fig. 3). The Fe resources (including inferred resources) in the area are estimated at about 32,000 Mt (Townsend and Flint, 1997), with a total production from the major mining concerns of 233 Mt in 2004–2005 (Western Australia Dept. Industry and Resources database), at a value of about A$8.3 billion (2004 dollar value). Excellent reviews of the Hamersley BIF are provided by Morris (1998), Trendall and Blockley (1970), and Trendall (2002).

BIF in the Hamersley Basin is within the ca. 2470–2450 Ma Hamersley Group (Fig. 3), which covers about 40,000 km^2 and is about 2.5 km thick. The Hamersley succession is divided from base to top into the Marra Mamba Iron-formation, Wittenoom Dolomite–Mt Silvia Formation–Mt McRae Shale, Brockman Iron-formation, Weeli Wolli Formation, ca. 2440 Ma Woongarra Rhyolite (a large concordant felsic sill near the top of the Hamersley Group; Trendall, 1995), and Boolgeeda Iron-formation. Economically important are the Marra Mamba Iron-formation and the Brockman Iron-formation, which host most of the iron ore in the region. The Brockman Formation consists of two economically important iron-formation units called the Dales Gorge Member (ca. 2480 Ma; Nelson et al., 1999), and the Joffre Member (ca. 2460 Ma; Pickard, 2002), separated by the Whaleback Shale Member. The Dales Gorge Member contains 17 bands of BIF.

The mineralogy of the Hamersley BIF includes quartz, hematite, magnetite, goethite, stilpnomelane, siderite, dolomite, ankerite, K-feldspar, muscovite, pyrite, clinochlore, talc, kaolinite, apatite, and pyrophyllite (Webb et al., 2003). The BIF are characterised by microbands and macrobands, with the former, according to Trendall and Blockley (1970), possibly representing seasonal (annual) varves. This hypothesis, although disputed, is supported by SHRIMP geochronological based depositional rates, which range from 19 to 225 m/Ma.

The major ore types of the Hamersley BIF can be divided into three groups (Morris, 1998):

- Leached BIF; residual concentrates of iron oxides, with very low P content;
- Martite–goethite ore, with moderate to high P (0.07–0.17%); and
- Martite–microplaty hematite ore, with low P (<0.07%).

Correlations have been made between the Hamersley Group in the Pilbara Craton with the Transvaal Group (Kuruman Iron-formation) in the Kaapvaal Craton of southern Africa (Pickard, 2003). These two basins have similar stratigraphic successions, depositional environments and geotectonic settings (Pickard, 2003), whether or not the Pilbara and Kaaapvaal cratons are considered fragments of the same Vaalbara supercontinent (Cheney, 1996).

Table 2
Iron ore and gold production in Australia and the world in 2003 and projected for 2008

Commodity	2003	2005	2008 (projected)
Iron (Mt)			
World	1106	1313	1173
Australia	195	467	220
Gold (t)			
World	2672	2519	3011
Australia	269	266	323
Copper (kt)			
World	15720	16665	18630
Australia	869	905	1032
Zinc (kt)			
World	9600	10282	10750
Australia	1510	1352	1607

Data from ABARE (2003), Australian Commodities forecasts and issues, June Quarter 2003, vol. 10 (2); and ABARE (2006), Australian Commodities September Quarter 2006; www.abareconomics.com.

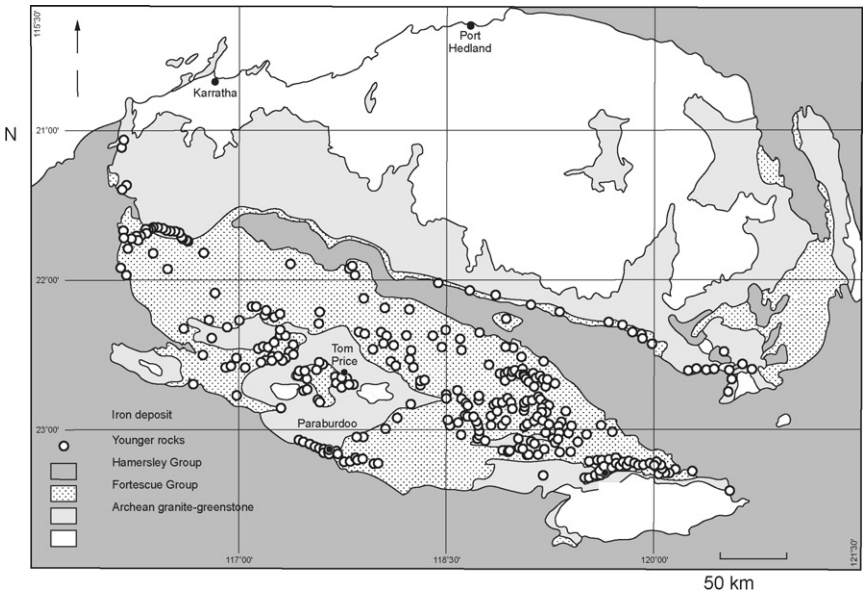


Fig. 3. Distribution of iron ore deposits in the Hamersley Basin (after Cooper and Flint, 2005). Note the abundance of iron deposits in the region.

In the Capricorn Orogen located between the Archean Pilbara and Yilgarn cratons in WAC (Fig. 1), the Earaheedy Basin contains a 5-km-thick succession of shallow-marine clastic and chemical sedimentary rocks (Earaheedy Group) that is divided into the Tooloo and Miningarra subgroups (Pirajno et al., 2004a). The Tooloo Subgroup consists of the Yelma Formation (base), Frere Formation and Windidda Formation (top). The overlying Miningarra Subgroup consists of the Chiall Formation (base), Wongawol Formation, Kulele Limestone, and Mulgarra Sandstone (top). The GIF beds of the Frere Formation, constitute a significant iron resource. In the western parts of the Earaheedy Basin, zones of supergene-enriched iron oxides contain between 21 and 66% Fe (Pirajno et al., 2004a; Pirajno, 2004a). In the Stanley Fold Belt, along the northern margin of the basin, the GIF beds contain platy hematite and magnetite mineralization that is spatially associated with strike-slip and reverse faults. The extent of this hematite-magnetite enrichment is revealed by aeromagnetic images, which indicate a total strike length of approximately 280 km. A model for the origin of GIF and BIF is shown in Fig. 4.

Two contentious issues for the genesis of these formations are the source of the metals, and the amount of oxygen that is necessary

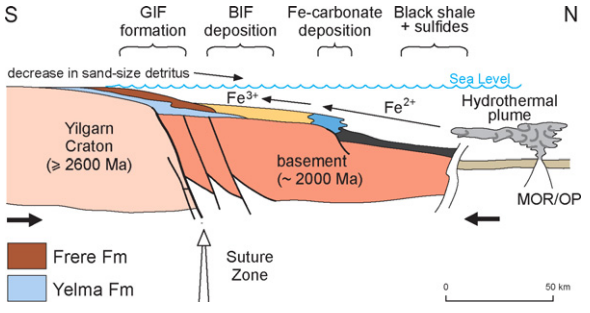


Fig. 4. Conceptual model for the deposition of granular iron-formations (GIF) in the Earaheedy Basin (details in text; after Pirajno et al., 2004b). Deposition of GIF coincided with a decrease in the supply of sand-sized siliciclastic detritus into the basin. GIF beds probably formed in shallow water on a continental shelf as ferruginous peloids formed by accretion in wave and current-agitated iron-rich water, and following reworking are deposited with variable amounts of terrigenous contamination. MOR: mid-ocean ridge; OP: oceanic plateau.

to induce oxidation of the Fe (and associated Mn). Possible sources of the metals are either Fe is derived from the weathering of iron-rich rock types (e. g. continental flood basalts), or Fe (and Mn) is introduced by extensive subaqueous hydrothermal discharges, in lakes or in ocean basins (Solomon et al., 2000), or in Red Sea-type brine pools (Krapež et al., 2003). Both possibilities require a density-stratified system, in which upwelling currents bring the reduced iron from anoxic deeper waters into the oxygenated environment of near surface waters, such as a continental shelf, where the Fe^{2+} is oxidised to Fe^{3+} and precipitated as oxides and carbonates. In summary, the required submarine volcanism and related extensive hydrothermal venting, presence of appropriate depositional systems (e.g. continental shelf), and deposition of black shales, suggest a cause and effect relationship (Barley et al., 1997; Condie et al., 2000; Abbott and Isley, 2001).

The commercial exploitation of the iron-formations is based on their subsequent enrichment to grades of up to 65% Fe. How this enrichment took place is controversial, with theories invoking supergene enrichment (e.g. Morris, 1985, 1998), hydrothermal processes (Barley et al., 1997), or flow of hot basinal fluids (Powell et al., 1999). The supergene model proposes that hematite and magnetite in BIF are enriched by supergene processes to an assemblage of hematite and goethite, and this assemblage is subsequently further modified by burial metamorphism to hematite-rich ore (Morris, 1985, 1998). The hydrothermal model of Barley et al. (1997) is based on fluid inclusion and petrographic studies of iron ore from the Hamersley Basin. In this model, interaction of high temperature fluids hotter than 150 °C, and possibly above 250 °C with BIF units, resulted in an assemblage of magnetite–hematite–siderite. Increasing oxidation at high temperature converted magnetite to martite (hematite) and microplaty hematite, and removed silica. Powell et al. (1999; also see Martin et al., 1998), using oxygen isotope systematics, fluid inclusions and textural data, suggested that regional flow of high-temperature oxidising meteoric fluids took place during orogenic activity at about 2400–2200 Ma. These fluids were expelled from a fold-and-thrust belt, along low-angle faults, towards the northern, less deformed margin of the host Hamersley Basin, resulting in the oxidation of BIF to hematite, goethite and microplaty hematite, with removal of silica. A model that involves both hydrothermal and supergene enrichment was proposed by

Webb et al. (2003), and other models of BIF deposition (Kappler et al., 2005; Konhauser et al., 2002) propose a mechanism for the oxidation of hydrothermal Fe(II) to Fe(III) by anoxygenic photosynthetic bacteria. A possible bacteria-mediated reaction is:

$$4Fe_2 + CO_2 + 11H_2O + \text{light energy} \rightarrow CH_2O + 4Fe(OH)_3 + 8H^+$$

Finally, it should be noted that a correlation between the deposition of Superior type BIF and mantle plumes, largely based on statistical evaluation of geochronogical data, was proposed by Isley and Abbott (1999; Abbott and Isley, 2001).

2.2. Orogenic and intrusion-related ore systems

Orogenic lode-Au deposits are defined by Groves et al. (1998) and Goldfarb et al. (2001) as a distinctive and coherent class of gold deposits, associated in space and time with collisional and accretionary tectonics. Intrusion-related Au deposits have been reviewed by Lang and Baker (2001). Given that most orogenic Au systems are spatially and temporally associated with granitic intrusions, the distinction between the two classes, especially in Proterozoic rocks, is difficult. Therefore, in this paper orogenic Au lodes and intrusion-related Au systems are considered together.

A worldwide episode of orogenic gold deposition took place during ca. 2100–1800 Ma (Goldfarb et al., 2001). In Australia, orogenic Au systems were formed in the NAC, WAC and SAC at ca. 1800–1790 Ma, reflecting collision and accretion tectonics in the three cratonic units (Şener et al., 2005). Host rocks containing orogenic gold deposits are typically associated with deformation and metamorphism at mid-crustal levels, are affected by deep-seated crustal structures, and are commonly spatially associated with granites (Goldfarb et al., 2001). However, a genetic and temporal relationship between the gold mineralization and the granites is not firmly established. In addition, there are examples of Australian Proterozoic orogenic gold deposits that are not spatially associated with magmatism (e.g. in WAC, discussed below).

In the NAC, orogenic Au deposits are present in the Pine Creek Orogen, Granites–Tanami Orogen (Bagas et al., 2007b; also called the Granite-Tanami Complex and Tanami region), Davenport Province of the Tennant Creek Inlier, and Arunta Orogen. Around 1800 Ma, the Pine Creek Orogen, Granites–Tanami Orogen, and Halls Creek Orogen gold mineralization and magmatism were broadly coeval with post-collisional extension following the collision of the NAC with the Kimberley Basin along the Halls Creek Orogen (Huston et al., 2007). This suggests that gold introduction in these two regions may be related to coeval tectonic events, as proposed by Şener et al. (2005). The WAC contains amagmatic orogenic, shear zone-hosted, Au lodes in the Paleoproterozoic Glenburgh and Capricorn Orogens (Fig. 1).

2.2.1. Pine Creek Orogen

The Pine Creek Orogen (Fig. 5) includes Neoarchean basement rocks, exposed as small granitic domes in the Neoarchean Rum Jungle and Nanambu complexes, a Paleoproterozoic (ca.

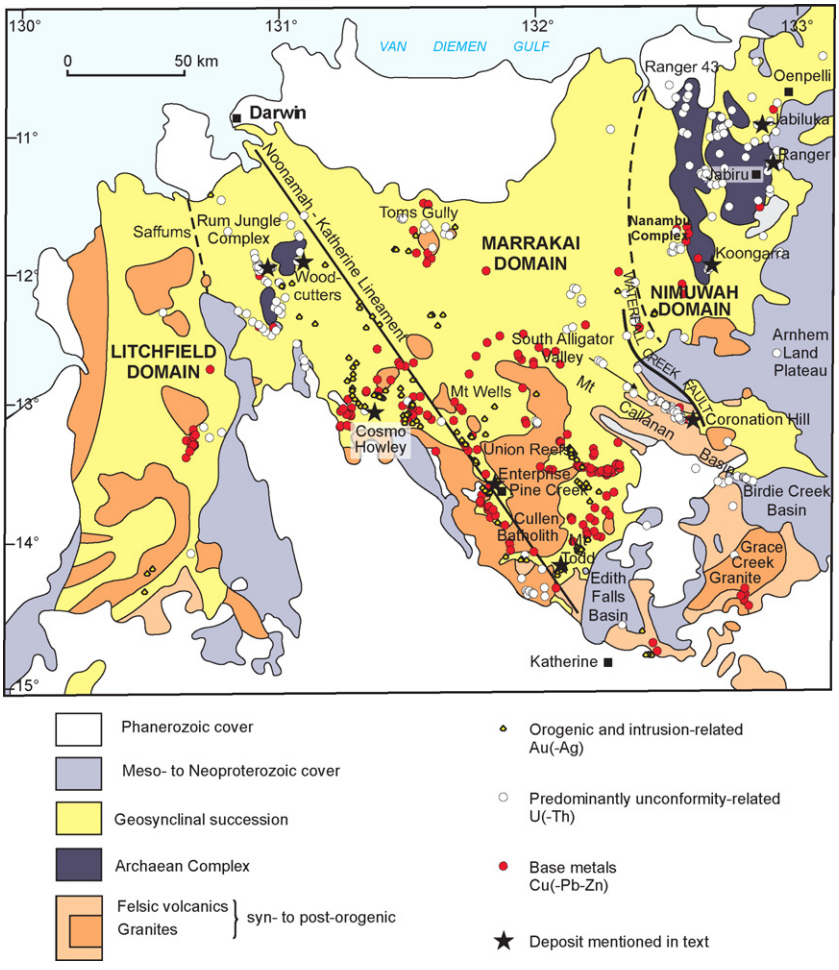


Fig. 5. Simplified geology of the Pine Creek Orogen showing the location of selected mineral deposits (after Needham et al., 1988; Northern Territory Geological Survey, 1999).

2000–1870 Ma) succession of sedimentary, pyroclastic and rare volcanic rocks, ca. 1800 Ma succession of clastic, pyroclastic and volcanic rocks and late Paleo- to Mesoproterozoic (ca. 1650 Ma) subhorizontal platform clastic rocks of the Katherine River Group in the McArthur Basin (Fig. 5).

Rocks in the Pine Creek Orogen were deformed by a series of compressional events between ca. 1870 and 1780 Ma. This prolonged period of deformation and metamorphism that includes the ca. 1870 Ma continental arc-type magmatic Nimbuwah Event, the ca. 1850 Ma Maud Creek Event, and the ca. 1800 Ma Shoobridge Event (Table 1; Needham et al., 1988). The Nimbuwah Event was followed by emplacement of granites during ca. 1870–1860 Ma (Needham et al., 1988), and the development of a NW-SE trending rift around the Alligator River Valley near the eastern margin of the Marrakai Domain (Fig. 5). The rift was filled with clastic, pyroclastic, and volcanic rocks of the El Sherana Group. The Pine Creek Orogen was then folded and faulted during the Maud Creek Event followed by deposition of the Edith River Group in graben or half-graben structures initiated by extensional faulting (Johnston, 1984; Stuart-Smith et al., 1993).

The 1835–1800 Ma Cullen Batholith in the Marrakai Domain and Grace Creek Granite to the east are younger than the Edith River Group (Needham et al., 1988). Numerous mineral deposits are located around the contact aureole of the batholith (Bagas, 1983; Stuart-Smith et al., 1988, 1993), and include structurally controlled quartz-veined deposits, greisens, and pegmatite-related or skarn deposits (Fig. 5). Recent geochronological data, based on SHRIMP U–Pb dating of hydrothermal monazite from the Tom's Gully Au deposit yielded an age of 1780 ± 10 Ma (Rasmussen et al., 2006), which is younger than the granitic rocks. However, these authors have also suggested that the Au mineralization may be related to the Shoobridge Event. A model of ore genesis for the Pine Creek deposits, taking into account the strong spatial association with the Cullen Batholith, is shown in Fig. 6.

The Pine Creek Orogen includes hydrothermal vein and stockwork deposits (containing Sn, W, Au, Ag, Pb, Zn, Cd, Cu, Bi, As, U, and Mo), massive sulfide deposits (containing Au, Ag, Cu, Pb, and/or Zn), alluvial Au and Sn deposits, residual massive Fe- and Mn-oxide deposits, and unconformity-related U(–Au–PGE) deposits (Stuart-Smith et al., 1993). Base-metal deposits in the orogen are like the Au deposits in that they are structurally controlled, have similar age relationships, and appear to have formed shortly before deposition of Au, in a mineral system characterised by metallogenic zoning from W and base-metal deposits close to granites to Au deposits further away (Fig. 6).

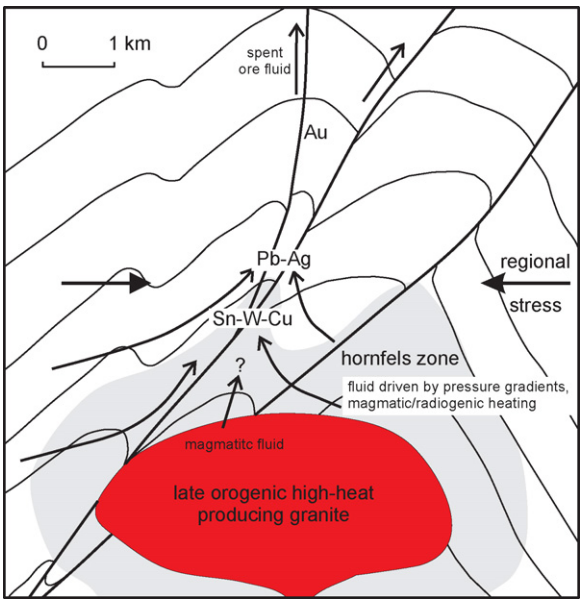


Fig. 6. Genetic model for the hydrothermal ore systems in the Pine Creek Orogen during the ca. 1800 Ma Shoobridge Event (modified after Bagas, 1983; Ahmad et al., 1999; Solomon et al., 2000). The zonation from Sn–W–Cu to Pb–Ag to Au away from the granite is controlled by temperature and metal solubility. Most of the metals (except Au) are likely to have been sourced from the granite, whereas the Au was probably sourced from the country rocks.

Several orogenic Au deposits are located around the Cullen Batholith in the central Marrakai Domain, and some deposits are located in the western Litchfield Domain and eastern Nimbuwah Domain (Fig. 5). The deposits in the Nimbuwah Domain are invariably associated with uranium.

Orogenic Au deposits in the Marrakai Domain are controlled by faults and shear zones trending north–northwest or northeast (e.g. the Enterprise deposit and the Mount Todd or Yimuyn Manjerr deposit), and by ironstones located in hinge zones of anticlines (Figs. 5 and 6; e.g. Cosmo Howley). The Pine Creek goldfield is about 1 km wide and 6 km long and includes 15 mines (Ahmad et al., 1999). The Pine Creek Shear Zone is a complex zone of faults that extend SE through the Cullen Batholith in the vicinity of the Yimuyn Manjerr gold mine, and lies in the Noonamah-Katherine Lineament Zone (Simpson et al., 1980). Many of the deposits are Au-bearing quartz veins forming saddle reefs, concordant and discordant veins, and stockworks along axial planes of anticlines associated with the Nimbuwah Event (Stuart-Smith et al., 1993). Most formed after the Nimbuwah Event, and many lie near to or outside the contact metamorphic aureoles of granites, but none are known to be hosted by granite. Many of the deposits are probably related to fluid movement through structurally prepared sites during low-grade (greenschist facies) metamorphism associated with deformation during the ca. 1800 Shoobridge Event and the later stages of the emplacement of the Cullen Batholith (Stuart-Smith et al., 1993).

The Enterprise deposit is in the southeastern end of the Pine Creek goldfield, in the Pine Creek Shear Zone (Fig. 5). The deposit is located in a tight anticline, and forms quartz saddle reefs and minor discordant quartz veins in faults and shears. The quartz veins contain free gold and gold associated with arsenopyrite. Other sulfides include pyrite, pyrrhotite, marcasite, chalcopyrite, galena, sphalerite, bismuthinite, tetrahedrite, and covellite (Ahmad et al., 1999).

Orebodies included in the Yimuyn Manjerr deposit are hosted by hydrothermal fractures and normal faults (Hein, 2003) and comprise discontinuous, steeply dipping, sheet-like veins hosted by the contact aureole southeast of the Cullen Batholith in the southeastern extension of the Pine Creek Shear Zone (Ahmad et al., 1999). The ore zones are discontinuous quartz–pyrrhotite–pyrite- or arsenopyrite–pyrite-dominant systems, which locally form stockworks containing fine gold (<60 μm) inclusions in sulfides or free gold (Ahmad et al., 1999; Hein, 2003). Other minerals include chalcopyrite, marcasite, bismuth, bismuthinite, galena, sphalerite, cubanite, talnakhite, hedleyite, and loellingite (Ahmad et al., 1999). The deposits are similar to other orogenic gold deposits within the orogen, but also have affinity to the thermal aureole gold style (Hein, 2003).

The Cosmo Howley Au deposit consists of quartz veins on the steeply dipping limbs and hinge of the Howley Anticline located in the high-temperature aureole of the Cullen Batholith (Fig. 5). The deposit is hosted by upper greenschist to lower amphibolite facies carbonaceous pelite, metasiltstone, ironstone containing recrystallised chert nodules and bands, and metadolerite sills (Ahmad et al., 1999). Gold is concentrated in structurally controlled sites developed at or near the contact between ironstone and carbonaceous units, forms inclusions in arsenopyrite and pyrite hosted by

both concordant and discordant quartz veins. Associated ore minerals include chalcopyrite, pyrrhotite, galena, and sphalerite (Ahmad et al., 1999). Matthäi et al. (1995) suggested that reduction of late magmatic brine by mixing with metamorphic fluids near the carbonaceous pelite was responsible for localization of gold. Heat from the prolonged intrusive and cooling history of younger ca. 1800 Ma granites in the Cullen Batholith, coupled with pre-existing structures, allowed long-lived hydrothermal systems to channel fluids from granite and metamorphic sources (Klominsky et al., 1996).

2.2.2. Granites–Tanami Orogen

The Neoarchean to Paleoproterozoic Granites–Tanami Orogen is located in the northwestern part of the NAC (Myers et al., 1996; Fig. 7). The complex is situated between the Halls Creek Orogen (Griffin and Gray, 1990) to the northwest, and the Arunta Orogen to the southeast (Fig. 1). Neoproterozoic and Phanerozoic sedimentary rocks conceal the contacts between these orogens.

The Granites–Tanami Orogen comprises mostly tightly folded greenschist-facies Paleoproterozoic metasedimentary and metavolcanic rocks. Isolated inliers of Neoarchean (ca. 2510 Ma) leucocratic orthogneiss and schistose metasedimentary rocks also crop out in the Northern Territory (Page et al., 1995). The Tanami Group (Blake et al., 1975) is divided into the ca. 1847–1840 Ma Dead Bullock and ca. 1840 Ma Killi Formations, and is regionally intruded by pre-tectonic dolerite sills that locally have peperitic upper and lower margins (Crispe and Vandenberg, 2002). Older (ca. 1864 Ma) metasedimentary, mafic volcanic, and intrusive rocks have recently been recognised at the Bald Hill deposit in the western part of the complex (Bagas et al., 2007a).

Orogenic lode-Au deposits in the Granites–Tanami Orogen are concentrated around the Granites and Tanami Au province in the Northern Territory and includes the world-class Callie deposit, which has a resource of 188 t or over 6 Moz or Au (Huston et al., 2007; Fig. 7). Significant mineralization is also located at the Coyote and Bald Hill deposits in Western Australia (Fig. 7). Gold deposits in the Granites–Tanami Orogen are hosted by a variety of rock types, including carbonaceous siltstone, iron-formation, basalt, dolerite, and turbiditic sedimentary rocks. Most deposits are associated with relatively late oblique strike and reverse faults, and many are localized where the late faults intersect anticlines. These structures provided the environment that allowed pressure fluctuations and the focussing of ore into chemically reactive rocks, producing large, high-grade deposits such as Callie (Huston et al., 2007). Huston et al. (2007) suggested that Au deposition was controlled by varied mechanisms that include competency gradients, decarbonation and sulfidation of host rocks, fluid mixing, and fluid pressure fluctuation. Another factor could include local perturbations in the temperature gradients that influence the movement of mineralized fluids in structurally prepared sites. The larger deposits appear to be located where many factors that control Au mineralization coincide (Huston et al., 2007).

Preliminary analyses of hydrothermal xenotime from the Coyote deposit and the Callie deposit suggest that gold was deposited ca. 1810–1790 Ma, which overlaps the ages of granites in the region (Bagas et al., 2007a; Huston et al., 2007). Similarities in ages of granites and deposits might indicate a genetic link, but this does not unequivocally demonstrate that the source of the mineralization is the granites.

Most of the gold in the Granites–Tanami Orogen is hosted by carbonaceous siltstone and iron-formation in the Dead Bullock Formation, which are extensively hydrothermally altered (Huston et al., 2007). At Callie, carbonaceous siltstone is decarbonized and bleached in ore zones (Smith et al., 1998), which appears to have accompanied gold deposition. Decarbonization leads to reduction of the ore fluid and the production of CO_2(g) and H_2(g) (Huston et al., 2007). Deposits hosted by iron-formation are generally strongly sulfidic in the ore zones, with local development of semi-massive sulfide (Smith et al., 1998), which suggests that Au deposition accompanied the loss of sulfur from the ore fluids (cf. Phillips and Groves, 1983). Other deposits, such as in the Tanami goldfield, Au deposition was probably controlled by fluid mixing or boiling. However, a number of deposits, such as the turbidite-hosted Coyote deposit, are not characterized by obvious wall-rock alteration, and ore-related alteration is restricted to narrow (<20 mm) chlorite- or biotite-rich selvedges along quartz-vein margins (Huston et al., 2007).

At Coyote, ore lenses are localised in brittle faults along the limbs of an anticline where Au is mostly hosted by bedding-parallel veins within metasiltstone only a few centimetres from contacts with metawacke, which is interpreted to represent competence gradients during mineralization. Mineralized quartz veins are typically associated with chlorite(-sericite-biotite) alteration selvedges that mostly extend less than 10 mm into the host metasiltstone. Gold deposition was late in the paragenesis of quartz veining, probably in response to fluid pressure fluctuations during local extension, synchronous with granite emplacement (Bagas et al., 2007a).

In summary, geological, geochemical and structural data suggest that gold deposition in the Granites–Tanami Orogen was structurally controlled, influenced by fluid-rock interactions in chemically reactive rocks, or controlled by fluctuations in fluid pressures due to the structural setting. The interaction of these three factors is probably the direct cause of the variety in the style of gold deposits in the region. Many of these deposits are temporally linked to the emplacement of ca. 1800 Ma granites (Huston et al., 2007).

2.2.3. Arunta Orogen

The Arunta Orogen (Figs. 1 and 8), located on the southern margin of the Precambrian NAC, is a major orogenic belt of approximately 200,000 km^2. The Orogen comprises two distinct east–west-oriented tectonic provinces (Aileron and Warumpi of Close et al., 2002, 2003; Scrimgeour, 2003). The ca. 1865–1710 Ma Aileron Province can be further subdivided into the Northern and Central Tectonic zones of Shaw et al. (1984) based on the geochronological framework of Black et al. (1993) and Collins and Williams (1995). The 1800–1500 Ma tectonic history of the orogen is summarized in Fig. 2. Throughout the Arunta Orogen are mafic–ultramafic intrusions emplaced during ca. 1810 Ma (Stafford Event), ca. 1780 Ma (Yambah Event), ~1690 Ma, and ~1635 Ma (Fig. 2; Hoatson et al., 2005; Claoué-Long and Hoatson, 2005). These intrusions, associated with major fault zones, consist of mafic granulites interleaved with gabbroic sheets and stacked dolerite sills, ultramafic bodies, and amphibolite sheets (Hoatson et al., 2005). These authors suggested that some of these intrusions may host Ni–Cu and sulfide deposits.

The Northern Tectonic zone is the most extensive and comprises a predominantly greenschist to granulites facies turbiditic successions (Stewart et al., 1984; Collins and Shaw, 1995). In the Central Tectonic zone of the Aileron Province an assemblage of mafic and felsic meta-igneous rocks is interlayered with, and locally overlain by, subordinate pelitic and calcareous metasedimentary rocks. Similar rocks also form a minor component of the Northern Tectonic zone (Black and Shaw, 1995; Collins and Shaw, 1995). A platform succession, comprising quartzite, shale, and carbonate characterises the Warumpi Province (formerly the Southern Tectonic Province; Shaw et al., 1984; Stewart et al., 1984). Granitic rocks have been dated at 1879 ± 11 Ma in the Arltunga Nappe Complex, which is thought to be part of the Central Tectonic zone thrust onto the Warumpi Province (Zhao and Cooper, 1992; Zhao and McCulloch, 1995; Zhao and Bennett, 1995). The Warumpi Province is mostly younger than the other provinces of the Arunta Oro-

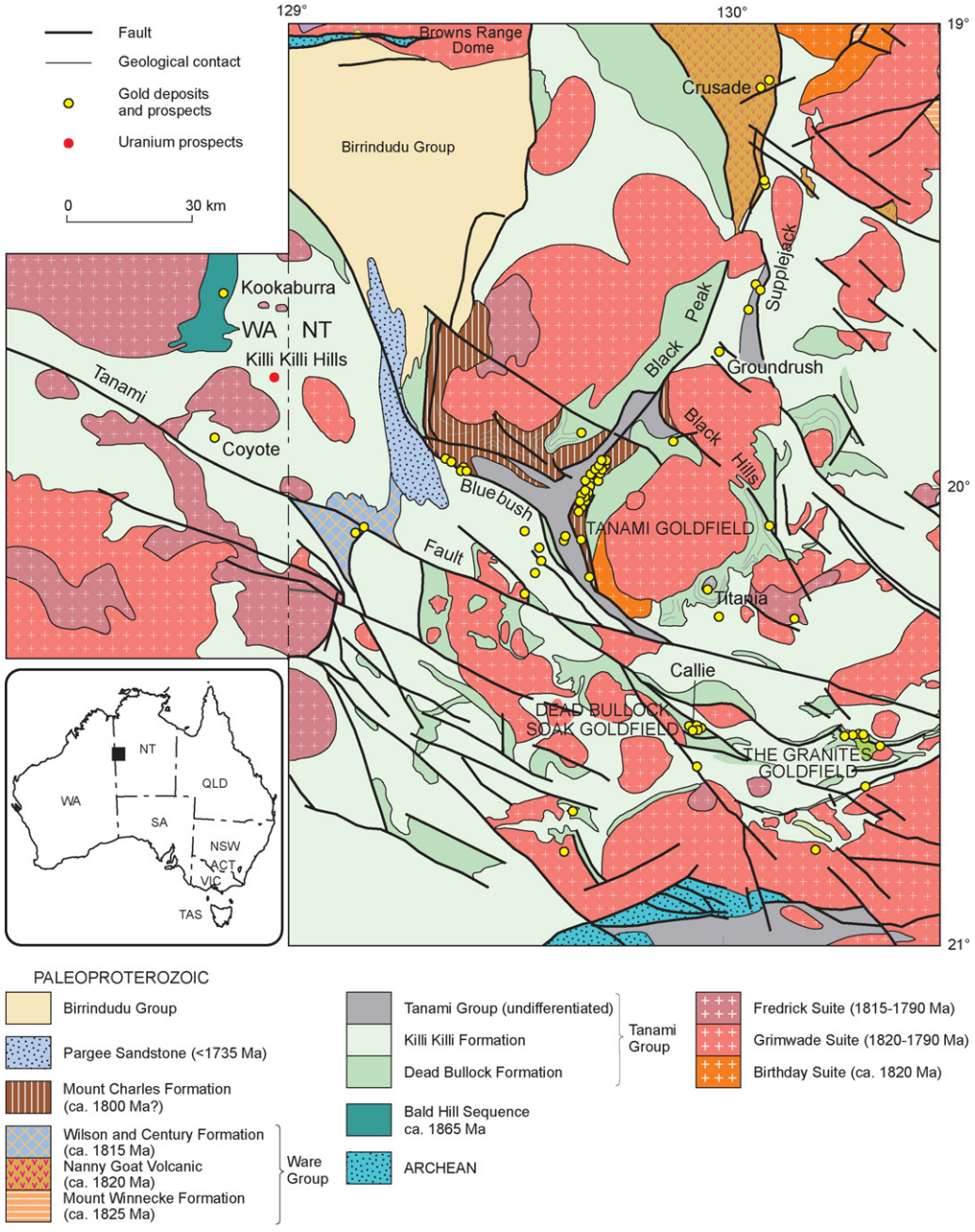


Fig. 7. Simplified Paleoproterozoic geology of the Granites–Tanami Orogen and distribution of gold deposits (modified after Huston et al., 2007).

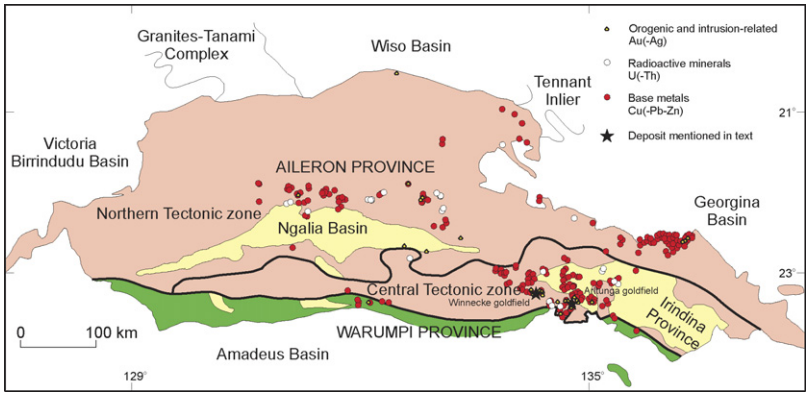


Fig. 8. Simplified geology of the Arunta Orogen and distribution of some mineral deposits (modified after Shaw et al., 1984; Northern Territory Geological Survey, 1999).

gen. It is characterised by amphibolite-facies quartzofeldspathic gneisses that are unconformably overlain by siliceous and aluminous metasedimentary rocks. The succession has been intruded by voluminous granite and widespread ca. 1080 Ma dolerite-dyke swarms that are probably part of the Warakurna LIP (Shaw et al., 1984; Black and Shaw, 1995; Collins and Shaw, 1995; Wingate et al., 2004). Most of the province is younger than about 1680 Ma, was not affected by the earlier orogenies in the Aileron Province, and may have been united with the rest of the Arunta Orogen at around 1640 Ma (Black and Shaw, 1995).

There are several sub-economic occurrences of Au, Cr, Cu–Pb–Zn, Ni, Sn, W, Mo, Ta, mica, and semi-precious gems in the Arunta Orogen. These may include VHMS (e.g. Warren et al., 1995), Broken Hill-type (e.g. Cu–Pb in the east; Mackie, 1984), or skarn (e.g. W–Mo in the east; Freeman, 1990; Hussey et al., 2004) deposits.

The contact between the Arunta Orogen and the Neoproterozoic to Paleozoic Amadeus Basin to the south has been deformed during the ca. 400–300 Ma Alice Springs Orogeny, and contains the Arltunga and Winnecke goldfields (Fig. 8). The deposits consist of quartz veins containing auriferous pyrite, and minor chalcopyrite and galena. The deposits are hosted by the Central Tectonic Province of the Arunta Orogen and Neoproterozoic sandstone (Heavitree Quartzite) in the basal part of the Amadeus Basin (Ahmad et al., 1999). The age of the Au occurrences are not certain; Warren et al. (1975) suggested that Au was deposited in the Arunta Orogen and remobilized into rocks included in the Amadeus Basin during the Alice Springs Orogeny (cf. Dunlap and Teyssier, 1995).

2.2.4. West Australian Craton

In addition to the early Paleoproterozoic iron deposits discussed above, the WAC contains amagmatic orogenic, shear zone-hosted, Au lodes in the Paleoproterozoic Bryah and Padbury groups, and other units of the fold-and-thrust belt, such as the Peak Hill Schist. The Peak Hill Schist is a mylonitic tectonic unit which, although it may be part of the southwestern tip of the Archean Marymia Inlier (Pirajno and Occhipinti, 2000), is considered together with the Bryah and Padbury groups as a coherent tectono-metamorphic domain. The mineralization is dominated by Au-only and Au–Cu lode deposits hosted in high-strain brittle–ductile and/or ductile shear zones in greenschist facies metasedimentary and/or metavolcanic rocks (Bryah and Padbury groups), or along contact zones between rock units. Mines have been developed at Peak Hill, Harmony, and Fortnum, and other economically significant lode deposits include Labouchere and Horseshoe (Fig. 1; Pirajno and Occhipinti, 2000).

The orogenic lode-Au in the Peak Hill Schist, Bryah and Padbury groups were formed during late stages of compressional deformation, followed by extension and retrograde regional metamorphism. The fluids were focused along thrust planes and shear zones, which acted as conduits of high permeability in a hydrothermal regime. The mineralization is associated with relatively narrow zones of hydrothermal alteration characterised by pyritisation and alkali metasomatism, with albite and biotite being the most abundant hydrothermal mineral phases. Other alteration minerals include chlorite, white mica, and tourmaline (Pirajno and Occhipinti, 2000). The mineralizing fluids were of low to moderate salinity and were H_2O–CO_2 rich (Thornett, 1995). The source of the metals is unknown, but it is possible that they were derived from either the volcano-sedimentary rocks within the basin or the underlying Archean basement.

The Egerton mineral field is hosted by an inlier of the Paleoproterozoic Gascoyne Complex within the Edmund Basin (Fig. 1). Mineralized veins consist of quartz–pyrite or quartz–pyrite–carbonate. The Hibernian gold deposit is sited within the Hibernian Shear Zone between gabbro and quartz–muscovite schist. Hydrothermal alteration, characterised by the presence of chlorite, epidote, carbonate, sulfides, and silicification, is located within or on the margins of a mafic intrusive (Pirajno, 2004a). The origin of this mineralization is not known, but may be related to deformation and metamorphic events associated with the Mangaroon Orogeny in the Gascoyne Complex (Fig. 2; Sheppard et al., 2005).

In the Paleoproterozoic Ashburton Province just south of the Pilbara Craton, the Mount Olympus orogenic Au system has a resource of about 30 t of gold. The Mount Olympus deposit is structurally controlled and is hosted by clastic and carbonate rocks of the <2455 Ma McGrath Formation. Gold is associated with quartz–sericite alteration, and was precipitated at 350 °C and 1–2 kbar (Young et al., 2003). Şener et al. (2005) dated alteration-associated phosphate minerals by SHRIMP U–Pb and obtained an age of 1738 ± 5 Ma, which is interpreted as the minimum age of an Australia-wide orogenic event that affected the WAC, NAC and SAC.

2.3. Orthomagmatic sulfides and oxides

In the Halls Creek Orogen, a number of layered mafic–ultramafic intrusions contain orthomagmatic, as well as hydrothermal, Ni–Cu–PGE, Co, Cr, V, Ti, Cu, Au mineralization (Hoatson and Blake, 2000). Basal sulfide segregations at ca. 1840 Ma Sally Malay have a resource of 3.8 Mt @ 1.8% Ni, 0.7% Cu, 0.1% Co and 0.2 ppm PGE (Hoatson and Blake, 2000). The Panton Sill contains stratabound PGE-bearing chromitite layers with a resource of 2 Mt @ 6.02 ppm PGE and Au (Hoatson and Blake, 2000). The Sally Malay sill-like lopolithic intrusion and the Panton intrusion are classified as orthomagmatic of tholeiitic association derived from the differentiation of tholeiitic magma that was subjected to varying degrees of crustal contamination (Sproule et al., 1999; Hoatson et al., 2006). Ore genesis is related to dynamics of magma flow from staging chamber to conduit, with physical traps being critical to sulfide precipitation.

The Musgrave Complex, in central Australia forms part of the Paterson Orogen and covers an area of approximately 140,000 km^2 (Glikson et al., 1996). The complex consists of a Paleo- to Mesoproterozoic basement of high-grade quartzofeldpathic metamorphic rocks including rocks of sedimentary or volcanosedimentary protolith. These have been intruded by mafic–ultramafic and felsic rocks of various ages and multiply deformed and metamorphosed between ca. 1600 and ca. 550 Ma. At ca. 1080 Ma the complex was intruded by the Giles mafic–ultramafic intrusions and temporally associated felsic volcanic rocks and alkaline granites, prior to further high-grade metamorphism (Glikson et al., 1996; Smithies et al., 2004). The Giles mafic–ultramafic intrusions are part of the ca. 1080 Ma Warakurna LIP (Wingate et al., 2004; Morris and Pirajno, 2005). The Warakurna LIP is the result of a widespread thermal event that in the Musgrave Complex also produced bimodal volcanic rocks, A-type granitic rocks, and mafic dyke swarms (Alcurra dyke swarm). The bimodal volcanic rocks of the Bentley Supergroup (including the Tollu and Cassidy groups), represent the highest level of the magmatic system.

The Warakurna LIP is estimated to cover about 1.5×10^6 km^2 and has a total west to east distance in excess of 2400 km (Fig. 9). It extends from the sill complexes of the Mesoproterozoic Edmund Basin (Bangemall Supergroup; Martin and Thorne, 2004), through to the Glenayle sill complexes of the Collier and Earaheedy basins in the WAC (Martin and Thorne, 2004; Pirajno et al., 2004b), to the Mann and Musgrave ranges in southern central Australia. The formation of the Warakurna LIP is postulated to be the result of a mantle plume that impinged onto the base of the lithosphere at about 1100 Ma (Fig. 9; Morris and Pirajno, 2005).

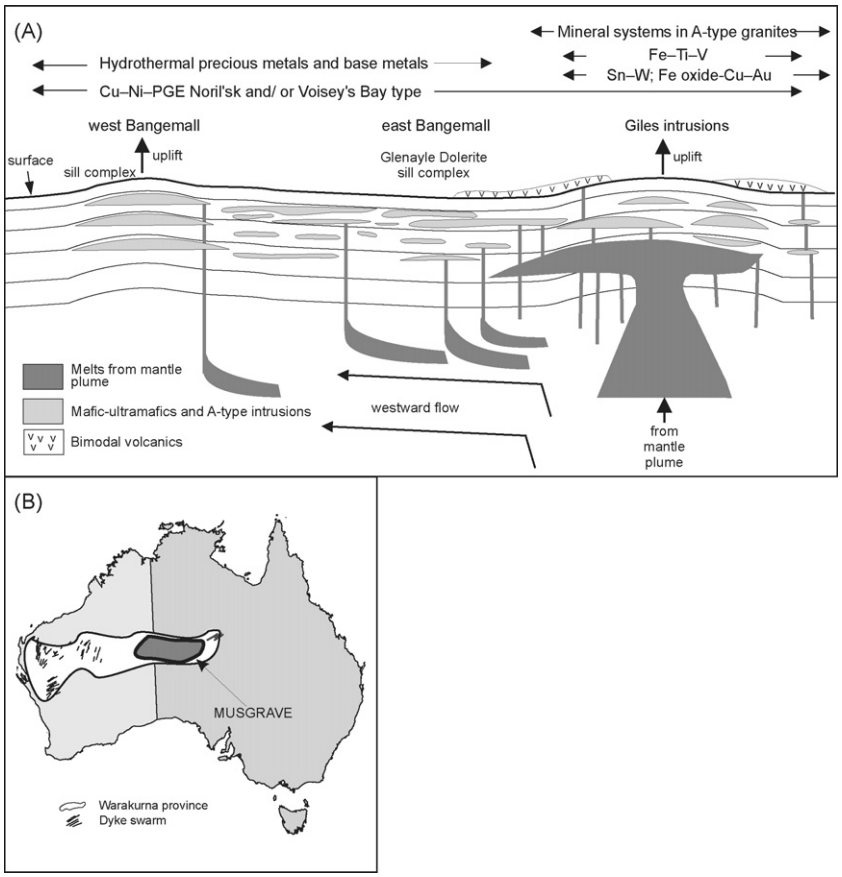


Fig. 9. (A) Schematic longitudinal section (not to scale) depicting the interpreted regional architecture of the Warakurna LIP that is characterised by a series of sill complexes (west and east Bangemall Supergroup) and the layered Giles mafic–ultramafic intrusions. It is assumed the Giles mafic–ultramafic intrusions were derived from a mantle plume that impinged onto the lithosphere at about 1080 Ma. In places, the plume manifested itself with the eruption of bimodal volcanics in the Musgrave Complex, whereas westward flow of mantle melts resulted in the emplacement of sill complexes in the Edmund and Collier basins. The model also shows the distribution of potential magmatic and hydrothermal mineral deposits across the Warakurna LIP (after Morris and Pirajno, 2005). B) Extent of the Warakurna LIP and position of the Musgrave Complex.

The Giles mafic–ultramafic intrusions consist of at least 20 sheet-like intrusions that extend roughly for 550 km along an E–W trend, with a total north–south extent of over 100 km (Fig. 10). Aspects of the geology of the intrusions are discussed in numerous publications, including Nesbitt et al. (1970), Daniels (1974), Goode (1977, 1978), Ballhaus and Glikson (1989, 1995), Gray and Goode (1989), and Glikson et al. (1996; and references cited therein). The Giles intrusions comprise gabbro, anorthosite, troctolite,

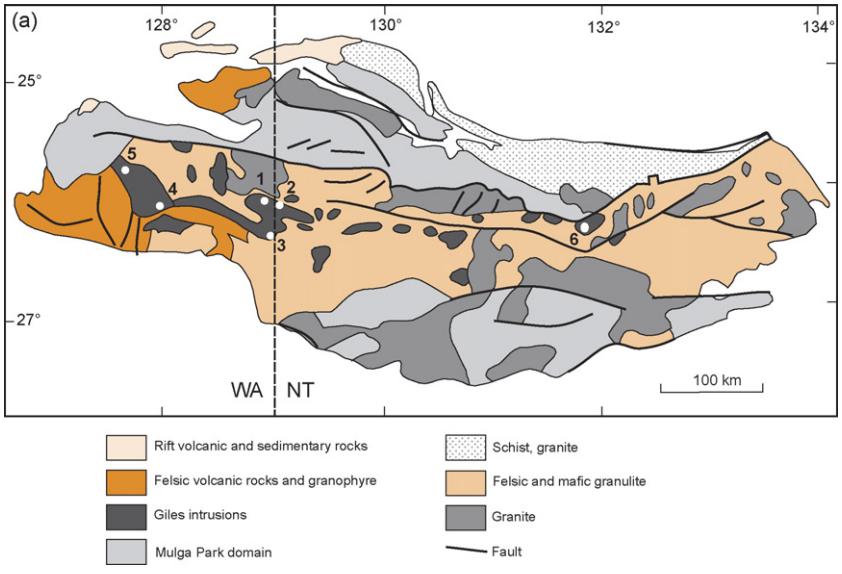


Fig. 10. Simplified geological map of the Musgrave Complex (adapted from Conor et al., 2002), showing distribution of the Giles mafic–ultramafic intrusions and associated principal ore deposits; (1) Wingellina laterite Ni; (2) Claude Hills Ni; (3) Bell Rock Fe–Ti–V; (4) Nebo-Babel Ni–Cu–PGE; (5); Zen-Canaan group Fe–Ti–V; (6) Mt Caroline Ni.

gabbronorite, norite, clino- and orthopyroxenite, dunite, and peridotite. Granophyres are present as dykes and as masses up to 60 m thick and locally cap layered gabbro.

Economically significant orthomagmatic mineral systems in the Musgrave Complex, reviewed by Pirajno et al. (2006), include: stratiform vanadiferous magnetite deposits; disseminated to semi-massive Ni–Cu–PGE sulfides hosted by a layered gabbronorite intrusion at Nebo-Babel (Seat et al., 2005; Baker and Waugh, 2005; Seat et al., 2007); and Mt Caroline in South Australia (Fig. 10), where Ni–Cu disseminations are hosted in olivine gabbro (Woodhouse and Gum, 2005).

The Nebo-Babel Ni–Cu–PGE deposit is located approximately 30 km west of the Cavenagh Range (Fig. 10). In early 2002, WMC (now BHP-Billiton) announced a drill intersection of 26 m @ 2.45% Ni, 1.78% Cu and 0.09% Co at Nebo-Babel. To-date this deposit represents the largest Ni sulfide discovery since Voisey's Bay in Canada (Seat et al., 2005, 2007). The Nebo-Babel deposits were discovered by conventional deflation lag geochemical sampling (Baker and Waugh, 2005). Little is known about the geology of the deposits and the only information, at the time of writing, is provided by Seat et al. (2005, 2007) from whom the following summary is taken. The Nebo-Babel mineralization is hosted in a tube-like mafic intrusion, with a cross-section of about 1 × 0.5 km, emplaced within granulite facies intermediate to granitic gneiss country rocks. Nebo and Babel comprises two mineralized zones of the same intrusion, offset by a fault that cuts the tube-like intrusion into western and eastern sectors. Sulfide mineralization is massive at Nebo (eastern sector) and net-textured and disseminated at Babel (western sector). Mineralogical and geochemical trends suggest that the intrusion may be overturned.

Both Nebo-Babel in the Musgrave Complex and Sally Malay in the Halls Creek Orogen have some similarities with the Voisey's Bay Ni–Cu–Co deposit in Canada (see Naldrett, 2004). The essence of such systems is that sulfide deposition is controlled by fluid dynamics in a magma conduit system (Naldrett, 1997; Li et al., 2001). A model for such deposits (Naldrett, 2004) suggests that magma rises to a lower (staging) chamber where the magma interacts with roof material and becomes sulfide-saturated with some separation of a sulfide liquid. A second pulse enters the staging chamber, forcing out the initial magma along a feeder conduit into an upper chamber. The second pulse of undepleted magma entrains sulfides and transports them along the feeder into the base of the upper chamber. As the second magma flows through the system, interaction occurs leading to upgrading in chalcophile elements and precipitation of sulfides in structural traps.

2.3.1. Vanadiferous titano-magnetite and ilmenite

In the western part of the Musgrave Complex, there are layers and lenses of V-bearing titano-magnetite and ilmenite hosted in troctolite and gabbro of the Giles mafic–ultramafic intrusions at Jameson Range and Blackstone Range, and at Bell Rock, with an estimated resource of 100 Mt @ 1% V_2O_5 (Daniels, 1974; Pirajno et al., 2006). At the Jameson Range, magnetite bands are present in Zones 2 and 4 of the intrusion (Daniels, 1974). Zone 2 in the northeast of the Range contains between 20 and 50 vol.% of opaques (probably all titano-magnetite) in the ultramafic layers, with the V_2O_5 tenor estimated at about 1.4 wt% (Daniels, 1974). Zone 4 has at its base a band of titaniferous magnetite, which has been traced along strike for about 19 km (Daniels, 1974). At one locality this band reaches a thickness of 15 m. Analyses of samples from this band yield an average of 0.9 wt% V_2O_5 and 23.4 wt% TiO_2 (Table 35 of Daniels, 1974). Zone 4 has more titaniferous magnetite layers towards the upper parts of the sequence, where one of these bands can be traced intermittently for 37 km, with thicknesses of up to 61 m at an average of 0.79 wt% V_2O_5.

2.4. Mineral systems associated with anorogenic magmatism

In this section, we consider mineral systems that are associated in space and time with magmatic activity in anorogenic environments. Anorogenic magmas host or are associated with a great variety of ore deposits, which include both magmatic and hydrothermal types. Such magmas are enriched in incompatible elements (e.g. Ti, P, Y, Nb, K, Th, U, F, Ba, REE) and produce peraluminous and peralkaline granites, which contain greisen or late-magmatic sub-solidus Sn, W, Zn, Cu, U, Nb mineralization (Pirajno, 1992 and references cited therein). In this section we also take a brief look at ore systems associated with alkaline rocks.

Hydrothermal ore deposits in this category include the economically important stratabound polymetallic and Au–Cu deposits and the ore systems of the Fe oxide–Cu–Au(–REE) style (IOCG; Porter, 2000 and references cited therein). In the first, we include two significant deposits, Abra in the Edmund Basin in WAC and the world-class Telfer in the northwestern part of the Paterson Orogen. The IOCG style mineral systems (Hitzman et al., 1992; Oreskes and Hitzman, 1993; Davidson and Large, 1994, 1998; Williams et al., 2005) are represented in Australia by giant ore deposits such as Olympic Dam and Ernest Henry, and deposits in the Tennant Creek Inlier. The general theme of the IOCG deposits is the enrichment in Fe, P, F and the widespread alkali metasomatism in the host rocks (Hitzman et al., 1992; Oreskes and Hitzman, 1993; Williams and Skirrow, 2000). Typically, IOCG hydrothermal systems form in shallow crustal environments (4–6 km) and are possibly the expression of volatile-rich, alkaline magmas (Hitzman et al., 1992). Their occurrence is especially widespread in the time span of approximately 1800–1100 Ma and they appear to be linked to planetary-scale rifting events and the assembly and breakup of supercontinents, such as Rodinia (Unrug, 1997).

In Australia, the period between 1800 and 600 Ma (and especially between 1700 and 1600 Ma) spans important metallogenic events (Fig. 2) that include giant hydrothermal ore systems (Solomon et al., 2000). Examples are the stratabound Pb–Zn–Ag–Ba(–Cu–Au–W) deposit at Abra in the Edmund Basin of the WAC (discussed below), the IOCG Olympic Dam-style breccia complexes in the Curnamona Province of the Gawler Craton (Cross et al., 1993; Williams and Skirrow, 2000; Skirrow et al., 2002), the Au–Cu deposit at Telfer in the Paterson Orogen, Zn–Pb–Ag SEDEX deposits in the McArthur River-Mount Isa-Cloncurry districts of the NAC (discussed below; Williams, 1998a, b), and IOCG deposits in the Tennant Creek Inlier of the NAC (Skirrow, 2000).

Continental reconstructions at 1600–1300 Ma (Moores, 1991; Karlstrom et al., 1999) show a geographic connection between the Gawler, NAC and WAC with Laurentia. It is possible that at least some of these hydrothermal ore systems could be part of a global metallogenic event (e.g. Thorkelson et al., 2001), linked to rift or rifted continental margin settings with thick sedimentary successions, continental mafic and anorogenic felsic magmatism.

2.4.1. West Australian Craton; Capricorn and Paterson Orogens—Abra and Telfer

The Abra stratabound sedimentary rock-hosted Pb–Zn–Ag–Ba(–Cu–Au–W) deposit, discovered in 1981 is located in the Edmund Basin (Mesoproterozoic Bangemall Supergroup) (Fig. 1). This polymetallic deposit contains approximately 150 Mt of mineralized rock, beneath a 200–500 m cover of rocks in the lower part of the Edmund Group (Abra Mining Ltd., 2006a, b, and www.AbraMining.com.au). The Abra polymetallic deposit is described by Vogt and Stumpfl (1987), Boddington (1990), Collins and McDonald (1994), and Vogt (1995). The deposit is hosted in dolomitic siltstone and shale and is characterised by a funnel-shaped stringer (feeder) zone, which is overlain by stratiform

ore.The stratiform ore includes an upper 'Red zone' and a lower 'Black zone', containing approximately 50 Mt of ore. The Red zone is dominated by jasperoidal material, hematite, barite, carbonate, and magnetite, with sphalerite and galena bands. The Black zone consists of hematite, magnetite, carbonate, barite, and sulfides, with bands of chloritised quartz–siltstone–jasper jig–saw fit breccias and stockworks of galena, barite and pyrite. Apart from galena and chalcopyrite, other ore minerals include tetrahedrite, sphalerite and scheelite, but in subordinate amounts. The underlying stringer (feeder) zone is estimated to contain about 150 Mt of base metals, including 0.13 g/t Au. The stringer zone consists of a stockwork of carbonate–quartz veins that cut through an alteration envelope of chlorite and siderite, which is enriched in silica and Fe, and depleted in Al, Ca and Rb in relation to unaltered rocks. Boddington (1990) noted that the upper part of the stringer zone contains higher Ba–Ag–Pb than the lower parts, which is enriched in Cu–Au. Vogt and Stumpfl (1987) and Collins and McDonald (1994) linked the genesis of the Abra deposit to ca. 1640 Ma rift-related tectonics and felsic magmatism. Vogt and Stumpfl (1987) also suggested that the source of Pb and Ba was the associated arkosic sediments. The possible association of the deposit with anorogenic felsic magmatism in a rift setting, together with Fe-rich alteration patterns raises the possibility that the Abra deposit may be a variant of an IOCG deposit, which are typically associated with hematite, magnetite and chloritic alteration (Hitzman et al., 1992; Hitzman, 2000; Haynes, 2000). Alternatively, the Abra mineralization could represent a Zn–Pb sedimentary exhalative (SEDEX) vent-distal system comparable to HYC (see below and Large et al., 1998), or perhaps the syn-rift clastic-sedimentary rock hosted stratiform barite-rich ore systems in the Selwyn Basin in northwesten Canada (Goodfellow et al., 1993).

The northwestern part of the Paterson Orogen has been subdivided into the Paleo- to Mesoproterozoic Rudall Complex basement, the Neoproterozoic Tarcunyah Group of the Officer Basin, and the Neoproterozoic Throssell Range and Lamil groups of the Yeneena Basin (Fig. 11; Bagas et al., 1995, 1999). The contact area between the Throssell Range and Lamil groups is marked by a major crustal feature known as the Camel-Tabletop fault zone, which is one of a series of structures that extend southeast to the Musgrave Complex of central Australia. Southwest of the ca. 650 Ma Telfer gold deposit, this fault zone contains sill-like, intermediate to mafic bodies that intrudes the Throssell Range and Lamil groups. Gabbro from the area has a SHRIMP zircon age of 816 ± 6 Ma (Reed, 1996). In

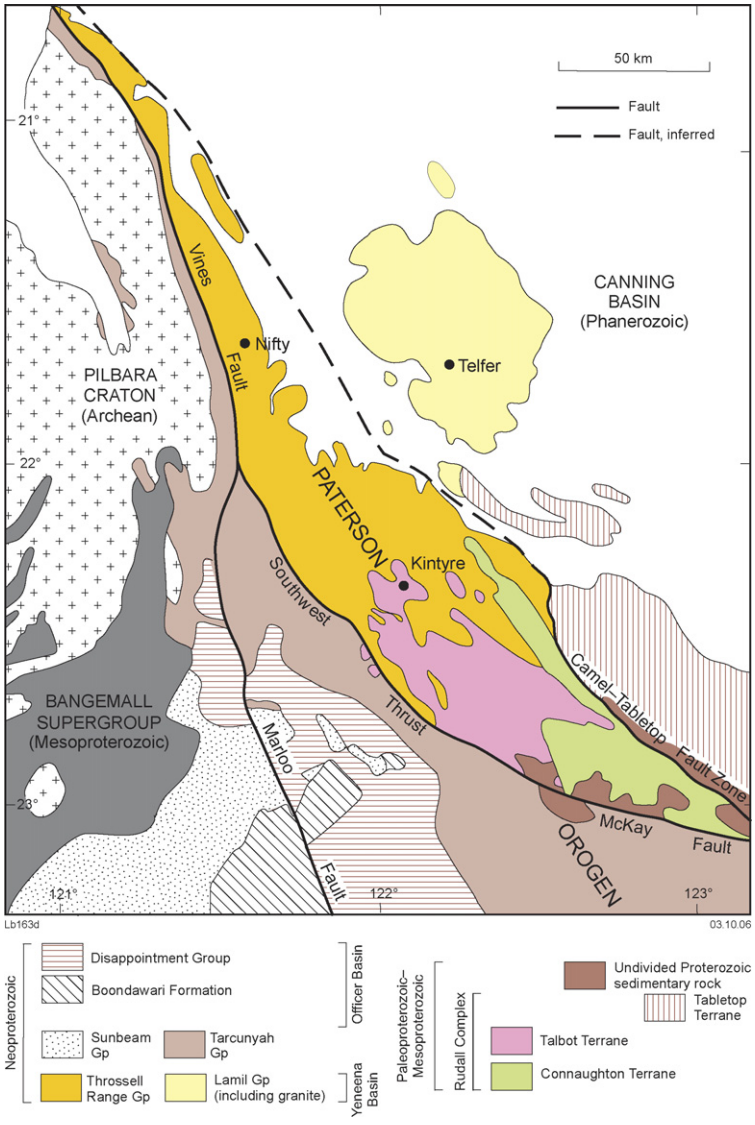


Fig. 11. Simplified regional geology of the northwest Paterson Orogen (adapted from Bagas, 2004).

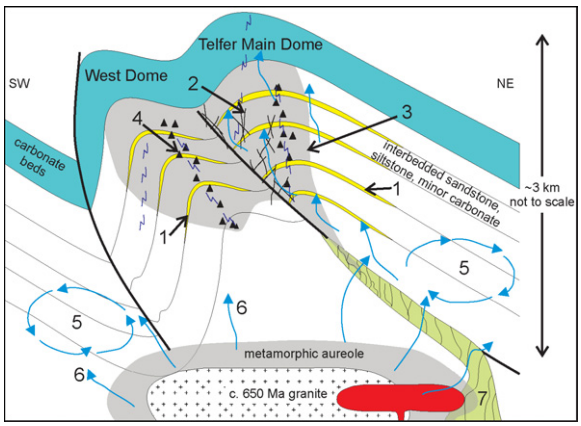


Fig. 12. Model for the Telfer mineral systems (modified after Dimo, 1990). (1) Stratabound, quartz–carbonate–sulfide reefs; main (ca. 650 Ma) Au deposit type. (2) Quartz–pyrite–chalcopyrite stockworks, and fracture-hosted sheeted veins. (3) Silica-dolomite alteration with disseminated Au, pyrite, chalcopyrite, and galena. (4) Breccia mineralization. 5: Mixed (formational and lesser magmatic) convective fluids (ca. 650 Ma; late to post-Miles Orogeny). 6: Metamorphic and magmatic fluids (ca. 650 Ma). 7: Early (?ca. 820 Ma) Cu–Au–Ag mineral system hosted by subvertical quartz veins in ca. 820 Ma (Reed, 1996) intermediate and mafic sills (pre- to early Miles Orogeny).

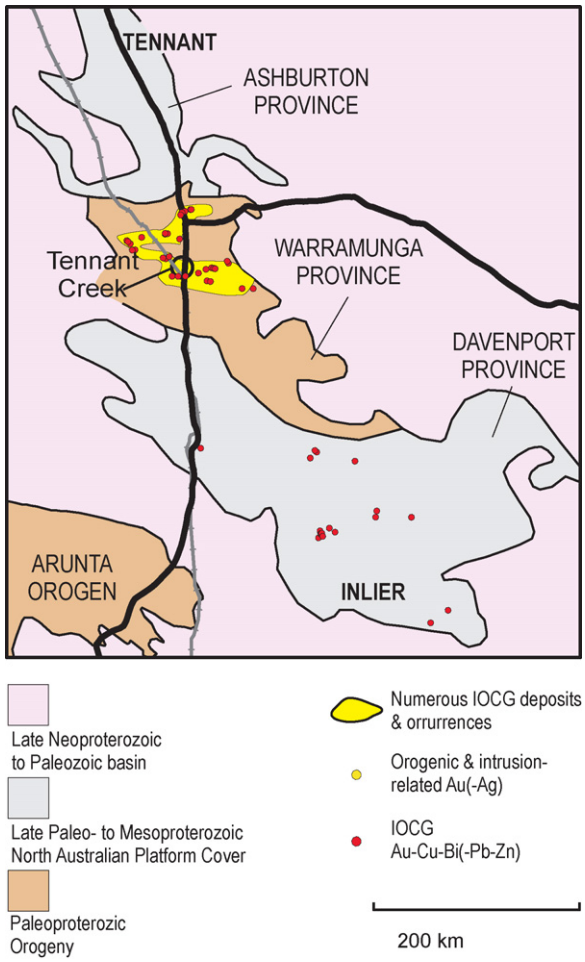


Fig. 13. Simplified geology of the Tennant Creek Inlier showing the location of some mineral deposits (modified after Blake and Page, 1988; Northern Territory Geological Survey, 1999).

the poorly exposed Yeneena Basin about 100 km north of Telfer, gabbroic sills of probably similar age, host Cu–Au–Ag mineralization in quartz–sulfide veins at the Magnum prospect (Gindalbie, 1999). This suggests that there might be at least two Au systems in the Yeneena Basin; one at ca. 820 Ma (related to intermediate and mafic intrusions), and the other at ca. 650 Ma (discussed below). These early intermediate and mafic intrusions may be linked to a ca. 800 Ma mantle plume associated with the Gairdner dyke swarm that intrude the SAC and parts of the WAC in the region of the Musgrave Complex (Zhao et al., 1994; Fig. 12).

The genesis of the Telfer Au–Cu deposit has been explained by either a syngenetic exhalative model, based on the observations that the mineralization is stratabound in the Malu Formation of the Lamil Group and pyrite is usually laminated (Turner, 1982), or epigenetic models (Goellnicht et al., 1989; Dimo, 1990; Rowins et al., 1997). The epigenetic models emphasize the importance of structural controls on mineralization that is hosted by a series of vertically stacked, stratabound quartz–carbonate–sulfide reefs, and centred on the anticlinal hinge zone of the Telfer Dome (Fig. 12). Other styles of mineralization include stockworks, pods, and vein systems in faults (Dimo, 1990). The stratabound quartz–sulfide reefs are laterally extensive, and are locally named the Middle Vale reefs, and the E-, M-, and I-Series reefs (Dimo, 1990). The reefs are several metres thick, mostly conformable, and hosted by calcareous or carbonaceous siltstone interbedded with massive sandstone of the Malu Formation. The reefs are typically linked by stockworks of quartz–sulfide veins and sheeted vein sets. The sulfide minerals are mainly pyrite and chalcopyrite, and form either aggregates or disseminations in carbonate and argillic veins in the quartz reefs. Gold is found in the pyrite as small inclusions and in fractures, and is generally associated with small amounts of chalcopyrite and trace amounts of pyrrhotite (Dimo, 1990). Wallrock silica–dolomite(–sericite–tourmaline–rutile–xenotime–monazite) alteration is restricted to narrow zones at the reef margins, which contain disseminations of fine-grained gold with disseminated euhedral to subhedral pyrite, and minor chalcopyrite and galena. Joint sets in the core of domal structures host the stockwork veins, which include quartz–sulfide veins and laminated quartz–carbonate–sulfide veins. The quartz veins are up to 0.2 m thick and contain up to 10 g/t Au; the laminated veins are up to 0.3 m thick with grades up to 160 g/t Au. High-grade mineralization is due to supergene enrichment that extends to 300 m below the surface (Dimo, 1990). This zone is extensively kaolinitic, has montmorillonitic alteration, and veined with clay. The weathered reefs contain chalcocite, limonite, goethite, hematite, unidentified iron oxides, and numerous other secondary minerals (Dimo, 1990). Fluid inclusion studies show that the ore fluids were rich in H_2O–CO_2–CH_4–NaCl, and reached temperatures between 225 and 450 °C (Dimo, 1990; Goellnicht et al., 1991; Rowins et al., 1997). Goellnicht et al. (1989, 1991) suggested that the mineralizing fluid was a mixture of fluids derived from magma and host or basin rocks, and mineralization is related to a distal Au-halo of a giant porphyry Cu–Au system. This indicates that the mineralization event may have been about 650 Ma, which is the approximate age that the post-tectonic, highly fractionated I-type granitic rocks intruded the Lamil Group (Dunphy and McNaughton, 1998).

2.4.2. North and South Australian Cratons; iron oxide–Cu–Au deposits (IOCG)

Australian examples of IOCG deposits include Olympic Dam in the Gawler Craton of South Australia, Ernest Henry, Starra in the Cloncurry district (Mount Isa), and a number of deposits in the Tennant Creek Inlier (Figs. 1 and 13). Several other deposits of the same class are present in the Curnamona Province of South Australia (e.g. Portia, Kalkaroo, White Dam, Mundi; Williams and Skirrow, 2000). These regions are characterized by extensive

Na–Ca(–Cl)- or K-metasomatism, and intrusions that are similar in age to the ore deposits (Barton and Johnson, 1996; Williams et al., 2005).

The Olympic Dam deposit contains ore reserves greater than 600 Mt averaging 1.8% Cu, 500 g/t U_3O_8, 0.5 g/t Au, and 3.6 g/t Ag (Reynolds, 2000) and a global resource of 31 810 Mt @ 1% Cu, 0.5 g/t Au and 400 g/t U_3O_8 (Williams et al., 2005). Most of the Precambrian rocks in the area consist of juvenile Proterozoic crust thought to be developed by lateral accretion of magmatic arcs (Ferris et al., 2002). These rocks are intruded by the Mesoproterozoic Hiltaba Suite of granites. Ferris et al. (2002) proposed that the magmatism associated with the suite was related to an intracontinental back-arc extension, whereas Pirajno (2000, 2004b) suggested that the magmatism was related to mantle plume activity.

The Olympic Dam deposit is hosted by the Olympic Dam Breccia Complex (ODBC) in the ca. 1590 Ma Roxby Downs Granite of the Hiltaba Suite, which is a coarse-granite syenogranite with A-type affinities (Creaser, 1989; Reeve et al., 1990; Johnson and Cross, 1995). ODBC and the Roxby Downs Granite form a basement high that is buried 300 m beneath subhorizontal Neoproterozoic and Cambrian sedimentary rocks of the Stuart Shelf (Reeve et al., 1990). Reeve et al. (1990) give detailed descriptions of the ODBC. The deposit is characterized by a number of large Cu-, Au-, Ag-, U-, REE-bearing hematite–quartz dyke-like breccia bodies. Copper is present as chalcopyrite, bornite and chalcocite, gold and silver form native metals, the main U minerals are coffinite, pitchblende and brannerite, and the REE minerals are monazite, xenotime, bastnasite and floencite. The abundance of REE correlates with increasing hematite content of the breccia. The breccia bodies are up to 100 wide, form a NW-trending zone that is around 5 km long and 1.5 km wide, and typically consist of brecciated granite at the edge of the deposit, passing into a heterolithic breccia and hematite–quartz microbreccia in central portions. The heterolithic breccia contains fragments that are generally less than 100 mm across. The fragments consist of hematite (some of which is bedded), crushed and altered granite in a matrix of quartz–hematite–sericite–siderite–chlorite, fluorite, siderite, barite (some of which is laminated), pyrite, and volcaniclastic rocks. The heterolithic breccia grades into the hematite microbreccia and fine-grained massive hematite–quartz (Cross et al., 1993; Haynes et al., 1995). The formation of the breccias implies large-scale movements of high-pressure fluids at a shallow depth (Fig. 14). Furthermore, the presence of sedimentary barite and hematite in the upper parts of the deposit suggests a phase of exhalative activity (Haynes et al., 1995).

Hydrothermal alteration assemblages grade from chlorite–hematite (after biotite–magnetite) in fractured granite, to sericite–chlorite–hematite, to sericite–hematite-only in the centre of the breccia bodies (Davidson, 2002). The extensive Fe-metasomatism was accompanied by fluorite, barite, and REE-bearing mineral phases, and the high-iron content of the ore suggests that the mineralizing fluid was saline (e.g. McKibben and Hardie, 1997). Collapse of upper parts of the breccia system was accompanied with deposition of sulfides (Haynes et al., 1995). Two sources of the mineralizing fluids have been recognized (Oreskes and Einaudi, 1992). An early fluid interpreted to be of magmatic origin and produced most of the magnetite in the ore system. This fluid had a high $\delta^{18}O$ values of 10‰ and a high temperature of around 400 °C. Later meteoric or seawater fluids deposited hematite, with $\delta^{18}O$ of <9‰, temperatures between 200 and 400 °C and salinities ranging from 7 to 42 wt% NaCl equivalent (Oreskes and Einaudi, 1992).

Davidson (2002) suggested that the mineralization at Olympic Dam is probably related to decompression of metalliferous, saline fluids in regions of lower permeability along highly permeable faults and breccias, in regions with higher geothermal gradients, and in over-pressured regions below fault seals (Fig. 14). Haynes et al. (1995) suggested that fluid mixing was the dominant process for the origin of ores, and involved magmatic and deep meteoric fluids mixing with cooler, near-surface, oxidized meteoric water. In this model, the groundwater carried in solution Cu–Au–U–S, which were introduced into the breccia and mixed with hotter water, which in turn introduced Fe–F–Ba–CO_2.

Skirrow (2006) suggested that IOCGs around Olympic Downs are characterized by: high paleogeothermal gradients associated with high-temperature K-rich A-type granites that were coeval with mafic and ultramafic intrusives; a regional-scale Na(–Ca–Fe) alteration up to 10 km wide; a regional to local-scale K–Fe(–carbonate) alteration and Cu–Au mineralization; local-scale hematitic alteration and Cu–Au(–U) mineralization; and hypersaline high-temperature brines (associated with magnetite alteration) and lower salinity lower temperature brines containing various concentrations of CO_2.

The Ernest Henry deposit is in the Cloncurry district at the eastern margin of the Mount Isa Inlier in Queensland, where 1760–1660 Ma metasedimentary (clastic–carbonate–evaporite) and bimodal metavolcanic rocks are intruded by 1550–1500 Ma alkaline to subalkaline A type granitic rocks. The rocks in the region are characterised by extensive Na and Na–Ca alteration, which is broadly synchronous with the granite intrusions (Mark et al., 2000). The Cu–Au ore at Ernest Henry deposit is the largest in the district, with a resource of 167 Mt at 1.1 Cu% and 0.54 g/t Au (Ryan, 1998; Mark et al., 2000), and it has a complex metal association of Cu–Au–Co–Ag with variable amounts of As–Se–Mo–Sn–Sb–Te–LREE–W–Hg–Bi (Williams, 1998a,b). The ore is hosted in a structurally controlled hydrothermal breccia vein or pipe-like body that has gradational contacts with a crackle breccia (Mark et al., 2000, 2005; Ryan, 1998). An important feature of the Ernest Henry mineral system is the complexity of multiple generations of veins, brecciation events and hydrothermal alteration that formed at temperatures between 400 and 550 °C (Mark et al., 2000). These can be considered in terms of regional Na–Ca alteration, pre- and early mineralization and alteration, Cu–Au mineralization, and post-Cu–Au mineralization events (Mark et al., 2000). Na–Ca metasomatism is characterised by assemblages containing albite, actinolite, titanite, quartz, magnetite, overprinted by K-feldspar, hematite, epidote, and quartz (Williams, 1998a; Mark et al., 2000). Alteration associated with the mineralization typically consists of biotite–magnetite and garnet–K-feldspar–biotite–pyrite assemblages. Mineral species, including ore minerals, are magnetite, hematite, chalcopyrite, pyrite, pyrrhotite, barite, fluorite, titanite, garnet, molybdenite, arsenopyrite, sphalerite, galena, monazite, scheelite, and REE-rich fluorocarbonate. In the weathering profile, ore minerals are cuprite, native Cu, chrysocolla, chalcocite covellite, and bornite (Williams, 1998a,b).

The Paleoproterozoic (ca. 1870 Ma) Warramunga Province in the Tennant Creek Inlier (Fig. 13) contains Au–Cu–Bi deposits associated with massive magnetite–quartz or hematite–quartz (ironstone) bodies included in the Tennant Creek goldfield. The Paleoproterozoic (ca. 1830–1815 Ma) Davenport Province in the southern part of the Tennant Creek Inlier (Fig. 13) contains a variety of small mineral occurrences including W, Au, Sn, Cu, Pb–Zn, Ni, Ta and Nb. The Tennant Creek mineral field is hosted by the lower greenschist facies clastic sedimentary and volcaniclastic rocks (ca. 1870 Ma Warramunga Formation). Pre- to syn-orogenic monzogranite, granodiorite, and porphyry sills that were predominantly intruded during 1870–1830 Ma (McPhie, 1993; Stolz and Morrison, 1994). In this period, a deformation event was accompanied by

greenschist facies metamorphism (Donnellan et al., 1995; Ahmad et al., 1999). The province is also intruded by late- to post-orogenic (late Paleoproterozoic) granite, porphyry, and lamprophyre (Ahmad et al., 1999).

Most IOGC deposits in the Tennant Creek mineral field are associated with ironstone bodies that are irregular in shape, flattened, ellipsoidal and pipe-shaped with long axes pitching steeply in an early (S_1) cleavage plane and formed syn- to post-cleavage (Rattenbury, 1992; Ahmad et al., 1999; Skirrow and Walshe, 2002). The ironstones are discordant to bedding and many form "lines of lodes" that trend east (Ahmad et al., 1999). Of the 700 or more ironstone bodies recorded in the Tennant Creek goldfield, less than 200 are known to contain mineral occurrences, and only 25 have recorded productions of over 100 kg of Au (Ahmad et al., 1999). Skirrow and Walshe (2002) have modelled the Tennant Creek Au–Cu–Bi deposits and suggested that the ores formed at temperatures of around 350–400 °C and consisted of a range of mineral systems from a reduced end-member to an oxidised end-member. The former being dominated by pyrrhotite, magnetite and pyrite, with salinities in the range of 3–10% NaCl eq., S isotope compositions of 0–5‰, and $CH_4(g)+N_2(g)$ in the fluids. In contrast, the oxidised end-member is dominated by pyrite, hematite, and magnetite; with salinities in the range of 12–20% NaCl eq.; S isotope compositions of −15 to +5‰; and $CO_2(g)+N_2(g)$ rich fluids. Mixing of the reduced with oxidised fluids engendered high Au-grades.

The main minerals in the ores of the Tennant Creek deposits are magnetite, chalcopyrite, pyrite, hematite, gold, bismuthinite, bismuth sulfosalts, chlorite, quartz, talc, dolomite, sericite, pyrrhotite, marcasite, bornite, galena, sphalerite, bismuth, cobaltite, uraninite, scheelite, stilpnomelane, minnesotaite, greenalite, actinolite, and cosalite (Large, 1975; Large and Mumme, 1975; Solomon et al., 2000). Gold is concentrated in magnetite–chlorite(–muscovite) zones in the ironstone bodies at their base, footwall or along their contact with the host rocks (Ahmad et al., 1999).

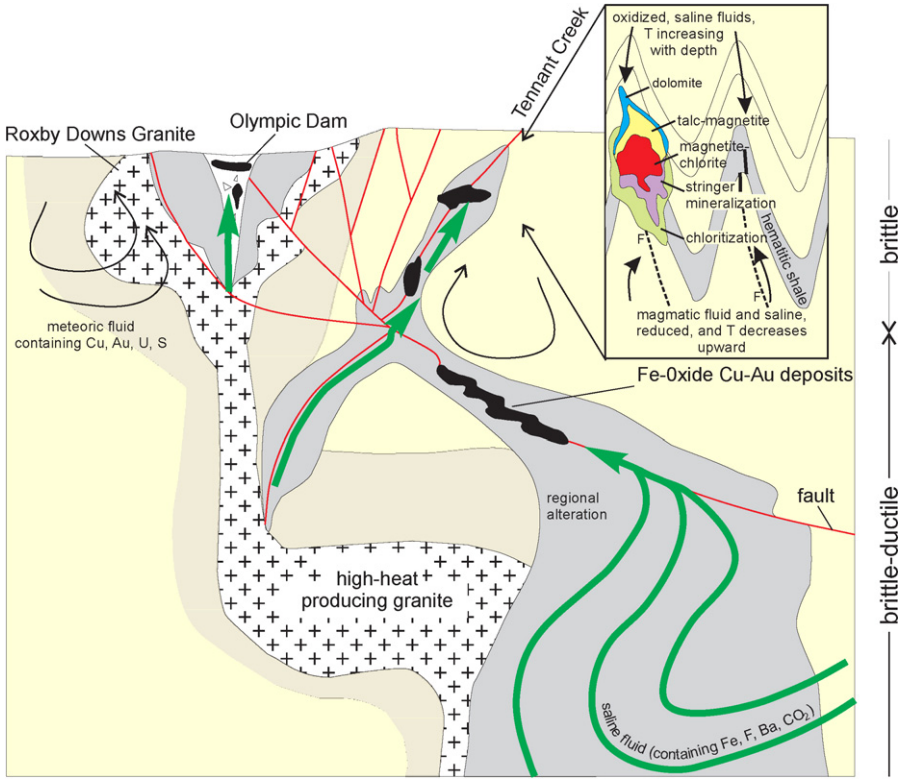
A conceptual model depicting the possible origin and paths of ore fluids for IOCG systems in the Gawler Craton and Tennant Creek Inlier is shown in Fig. 14.

2.4.3. *Alkaline rocks, lamprophyres, carbonatites, kimberlites, lamproites*

Alkaline ($K_2O/Na_2O = 1–3$ wt%) rocks in Australia span an age back to the Archean (Jaques et al., 1985). In Halls Creek Orogen and the Kimberley Basin, Proterozoic alkaline rocks include a number of carbonatite, lamprophyre, and kimberlite and lamproite intrusives ranging in age from about 1100 to 800 Ma. The 1180 Ma Argyle AK1 pipe, by far the most economically important deposit in the region, is a typical diatreme of olivine lamproite pyroclastic rocks and associated dykes with inferred diamond resources (at 2000) of about 145 Mt @ 2.9 cts/t (Hassan, 2000). The ca. 800 Ma Cummins Range carbonatite intrusion consists of both sovite (calcite dominant) and beforsite (dolomite dominant). This carbonatite contains zircon, baddeleyite, titanite, monazite, pyrochlore, columbite, and allanite as accessory minerals and has anomalous amounts of REE, P_2O_5, and Ta (Hassan, 2000). However, the grades are too low and at present this carbonatite is uneconomic.

Jaques et al. (1985) reported on several other Proterozoic alkaline complexes and rocks in the Pine Creek Orogen, McArthur River Basin, and Arunta Orogen. To-date, none of these has economic potential.

Australian kimberlites, lamproites, and several alkaline intrusions between ca. 1100 and ca. 800 Ma are sited around the margins of the Kimberley Basin and Proterozoic orogens on the northern



Model for the formation of Fe oxide Cu-Au deposits at different crustal levels
(adapted from Davidson, 2002, Ahmad et al., 1999 and Solomon et al., 2000)

Fig. 14. Conceptual model for the formation of Fe oxide-Cu–Au deposits at different crustal levels (adapted after Large, 1975; Wedeking et al., 1989; Davidson, 2002; Solomon et al., 2000).

and eastern margins of the Yilgarn Craton, and they may relate to episodes of intraplate magmatism (O'Neill et al., 2005). These alkaline rocks may correlate with the ca. 1080 Ma Warakurna LIP (Wingate et al., 2004) and the 827–800 Ma Mundine and Gairdner thermal events (Zhao et al., 1994; Wingate and Giddins, 2000) that affected parts of the WAC, NAC and SAC. These thermal events have been attributed to the impingement of mantle plumes (Pirajno, 2004b; Pirajno et al., in preparation). Carbonatites, kimberlites, and other alkaline complexes have isotopic signatures (He, Os, Sr, Nd, Pb, O) similar to ocean-island basalts (e. g. Nelson et al., 1988), and as such may be considered as part of plume magmatism. They may represent distal expressions resulting from the channelling of plume material along pre-existing lithopheric breaks, small degrees of melting of enriched/metasomatised lithosphere and/or crustal melts. A link with mantle plumes has been invoked by a number of workers (e.g. Bell, 2001; Schissel and Smail, 2001). Ebinger and Sleep (1998) suggested that a mantle plume might focus its flow towards craton-mobile belt boundaries, where small volumes of melt can be produced by decompression at depths of >150 km. Indeed, several of the alkaline rocks are emplaced within the Halls Creek and King Leopold orogens, or along the margins of the Kimberley Basin (Fig. 1). Although a connection with mantle plumes is by no means certain, it may be that thermal perturbation of anomalously hot mantle triggers melting, particularly in region of thinned or weakened lithosphere, even if these are a long distance away.

2.5. Stratiform and stratabound sedimentary rock-hosted mineral systems of intracratonic rifts

Rift basins developed mainly in the NAC between ca. 1800 and 1660 Ma (Betts et al., 2003 and references cited therein), in which exhalative shale-hosted Pb-Zn-Ag deposits were formed (Fig. 15). The McArthur River-Mount Isa Inlier rift-system of the NAC extends for more than 1000 km along a north–northwest trend. These continental rift-related basins are characterised by bimodal (felsic–mafic) magmatism and thick clastic sedimentary sequences with widespread alkali metasomatism. Crustal thinning, extension and bimodal magmatism suggest some form of mantle upwelling, perhaps due to a mantle plume or to convective removal of continental lithosphere (crustal delamination; Gibson et al., 2004). Shale-hosted giant stratiform sulfide deposits formed in this tectonic-depositional environment between 1640 and 1575 Ma (Betts et al., 2003). Similarly, in the SAC, the Olary-Willyama Block contains sedimentary and bimodal volcanic rocks in the ca. 1730–1700 Ma Willyama Supergoup, which hosts the stratiform Broken Hill massive sulfide deposits (Gibson et al., 2004; Page et al., 2005).

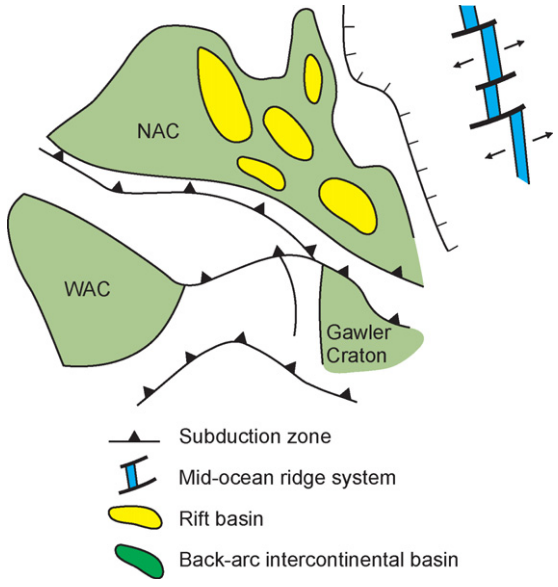


Fig. 15. Schematic and not-to-scale sketch showing the development of intracontinental basins in the NAC in a back-arc tectonic setting (based on Betts et al., 2002; Betts and Giles, 2006).

The McArthur River-Mount Isa rift system includes some of the largest massive sulfide deposits in the world, such as Mount Isa, Hilton, HYC (Here's Your Chance), George Fisher, Lady Loretta, and Dugald River (cf. Williams, 1998a, b). Similar to these, but of different age is the ca. 1595 Ma Century deposit. Prior to mining, Mount Isa contained about 150 Mt @ 7% Zn, 6% Pb and 150 g/t Ag; HYC contained 237 Mt @ 9.2% Zn, 4.1% Pb, and 41 g/t Ag; Century contained 118 Mt @ 10.2% Zn, 1.5% Pb and 36 g/t Ag (Betts et al., 2003; Large et al., 1998, 2005; see also Forrestal, 1990; Perkins and Bell, 1998; Painter et al., 1999; Economic Geology, 1998; Australian Journal of Earth Sciences, 2000, 2006). A description of these deposits is beyond the scope of this paper, for which the reader is referred the above-cited references. However, important features of these giant ore systems, as listed by Large et al. (2005), include: (1) ores located at or near basin-scale synsedimentary faults, along which fluids were channelled; (2) hosted in organic-rich black shale and siltstone; (3) bedding-parallel synsedimentary sulfides; (4) stacked ore lenses separated by Fe–Mn carbonate-bearing rocks; (5) metal zonations, vertical with upward decreasing Zn/Pb ratios, and lateral with increasing Zn/Pb ratios away from feeder structures; (6) stratabound Fe–Mn alteration haloes; and (7) lack of footwall stringer zones or vent systems. A conceptual genetic model for the McArthur River and Mount Isa ore systems is schematically presented in Fig. 16.

The Century Zn–Pb–Ag deposit may be unique in the Mount Isa rift-system. Broadbent et al. (1998) considered it as a new end-member of the hydrothermal sedimentary ore deposit family. The following is summarised from these authors. The Century mineralization is Zn-rich, is hosted by deep-water siliciclastics and consists of two major zones or "blocks" that are separated by a fault. The mineralization transgresses the stratigraphy, in that the higher grade zones are associated with host rocks that have thicker black shale units, indicating a replacement origin. The deposit is spatially associated with regional-scale lode mineralization, mainly veins and breccias within fault zones. The main ore mineral is sphalerite, which is present as a porous form associated with abundant pyrobitumen, and a non-porous form with low pyrobitumen content. The intergrowth of the sphalerite with hydrocarbons, suggests that thermochemical sulfate reduction (reaction of deeper sulfate and metal-bearing fluids with liquid and gaseous hydrocarbons), took place. The proposed reactions are:

$$CH_4 + SO_4^{2-} \rightarrow H_2S + HCO_3^- + OH^-$$

$$HCO_3^- + Me^{2+} \rightarrow MeCO_3 + H^+$$

and

$$H_2S + Me^{2+} \rightarrow MeS + 2H^+$$

where "Me" is a metal, such as Pb, Zn and Cu.

The genetic model suggested by Broadbent et al. (1998) proposed that sulfate-bearing hydrothermal solutions originated from the deeper parts of the basin and were channelled along crustal fault systems. The movement of fluids is related to phases of basin inversion and regional deformation. The hot fluids reacted with organic-rich shales, producing gas and liquid hydrocarbons. This

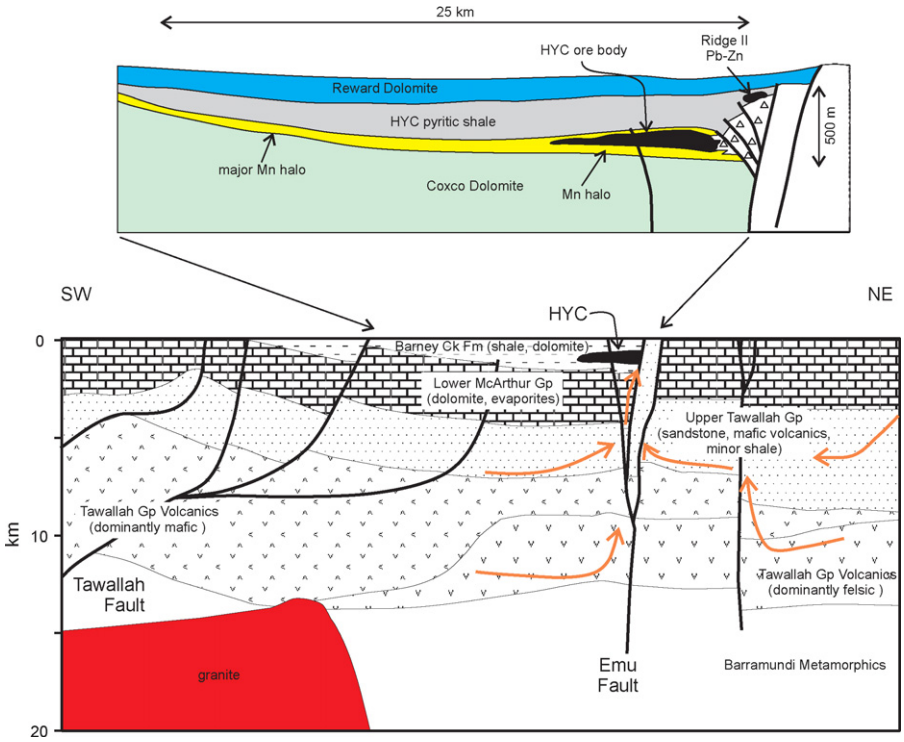


Fig. 16. Conceptual model, after Large et al. (1998) and Large et al. (2005), to schematically illustrate the development of shale-hosted massive sulfides in the McArthur River-Mount Isa rift system; the example here refers to the HYC ore deposit. Saline fluids may derive from evaporate beds, mixing with magmatic fluids from subjacent granite plutons. The cross-section of this figure is based on seismic reflection data from Rawlings et al. (2004).

led to the thermochemical reactions, which resulted in the precipitation of sulfides, as shown in the equation above. Sulfur isotope systematics revealed progressively higher $\delta^{34}S$ values, from 5–10 to 20–25‰, reflecting a closed-system hydrothermal reservoir with enrichment in heavy sulfur over the life of the mineralization event. The sulfate-bearing fluids may have derived from evaporite rocks in the deeper parts of the basin. Broadbent et al. (1998) suggested that the Century ore might represent a link with Mississippi Valley-type (MVT) deposits.

Plimer (1986) suggested that the Mount Isa mineralization is related to convective hydrothermal systems generated in a failed rift coeval with maximum tectonic activity and highest geothermal gradient. The hot saline brines, with a high metal-carrying capacity, would have leached metals from the enclosing sedimentary rocks, and exhaled in oxygen-deficient pools. Large et al. (2005) discussed three current genetic models, namely: (1) Zn–Pb–Ag ore lenses formed by diagenetic replacement below the sea floor, with Cu being added at a later stage by syntectonic replacement; (2) Cu as well as Zn–Pb–Ag formed by exhalative processes and/or syndiagentic replacement on and below the sea floor, with fluids being channelled through a feeder fault; (3) the orebodies were formed by "syntectonic replacement" during the Isan orogeny, but this model is not favoured by Large et al. (2005), who pointed out the lack evidence, such as the isotopic data, which support the synchronicity of the mineralization and sedimentation.

The Broken Hill Pb–Zn–Ag stratiform and stratabound ore systems are some of the largest in the world. The stratiform orebodies form several horizons in the succession and include the classic Broken Hill type (Parr and Plimer, 1993; Walters, 1996; Solomon et al., 2000), with Pb and Zn sulfides accompanied by assemblages containing quartz–gahnite, garnet–quartz, quartz–fluorite and banded iron oxides. The stratabound deposits are characterised by base-metal sulfides hosted by quartz–amphibolite (Solomon et al., 2000). The Broken Hill Main Lode constitutes one of the largest concentrations of Pb and Zn in the world and extends for more than 7 km to a depth of over 850 m, in which six individual lodes are present, possibly replicated by deformation, with some dominated by galena and others by sphalerite (Solomon et al., 2000). Interestingly, whereas Fe sulfides are somewhat scarce, Mn silicates, apatite and fluorite are locally abundant. This led Plimer (1986) to suggest a vertical and horizontal metal zonation, possibly from a series of feeder systems (see below and Fig. 17). In the Mount Isa Inlier, the ca. 1678 Ma Cannington Pb–Zn–Ag deposit contains ca. 44 Mt @ 11.6% Pb, 4.4% Zn and 538 g/t Ag is a Broken Hill-type system (Solomon et al., 2000 and references cited therein). Much of the ore at Cannington consist of veins and breccias hosted in amphibolite facies rocks, mostly quartzo-feldspathic gneiss and amphibolite. The nature of the mineral assemblages of the Cannington deposit suggest a Zn-skarn environment in which ore fluids became reduced by interaction with graphitic metasedimentary rocks (Solomon et al., 2000).

The bimodal magmatism that affected the rift succession of the Willyama Supergroup was probably the result of crustal thinning and mantle upwelling (Gibson et al., 2004), and this would have provided the heat energy for hydrothermal convection. Plimer (1986) examined the Broken Hill massive sulfide deposits in terms of characteristic metal associations and zonations, and the associated lithologies. He attributed part of the fluid system to be mantle-generated fluids that were rich in CO_2, F, P, Mn, Fe, Pb, Zn, etc. The Broken Hill Pb–Zn–Ag–Cu deposits are spatially associated with a host of other hydrothermal mineral deposits such as stratabound Cu–Au, W, epigenetic Pb–Ag, Cu–Ag, Cu–Ni–Fe and Pt deposits, and pegmatitic Sn–Ta, U, Nb, Be and Sn deposits. By contrast, the Mount Isa sulfide deposit includes Pb–Zn–Ag ores and brecciated sediment-hosted Cu ores, and there may be a spatial and genetic association with MVT (carbonate-hosted) Pb–Zn deposits.

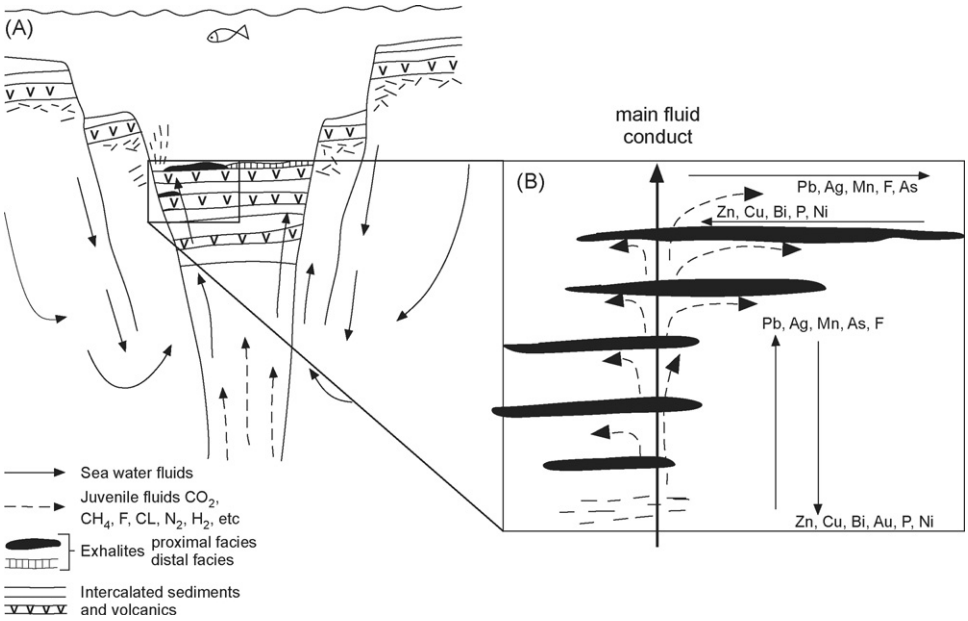


Fig. 17. (A) Hydrothermal circulation in a deep rift basin, in which magmatic and/or mantle fluids rise along faults and mix with sea or lake waters, to exhale in subaqueous environments to form massive sulfide deposits and associated exhalites; (B) horizontal and vertical zonation from a conduit(s) resulting in a series of stacked ore lenses (modified after Plimer, 1986; Large et al., 1998; Large et al., 2005).

Sawkins (1989) considered the formation of the "giant" Pb–Zn–Ag rich exhalative sediment-hosted massive sulfide deposits of Broken Hill, McArthur and Mount Isa as having a link with anorogenic felsic magmatism. He related the large abundance of Pb in these deposits with the availability of Pb-rich material, due to abundant feldspar-bearing rocks, such as crustal granites, dacitic–rhyodacitic volcanics, and extensive pyroclastics of anorogenic tectonic settings and their derived clastic sedimentary units. Betts et al. (2003) concurred that felsic rocks were a potential source of metals and further suggested that anoxic and organic-rich conditions in the sedimentary environment provided the necessary conditions for the formation and preservation of the massive sulfide deposits. Giles et al. (2002) and Betts et al. (2003), preferring the far-field back-arc continental rift setting as a tectonic model, suggested that long-lived lithospheric extension, high heat flow and a change from rift basin to sag basin over a protracted period of time (ca. 140 Ma) were fundamental factors in producing the necessary conditions for the genesis of shale-hosted massive sulfides.

A genetic model for the Broken Hill sedimentary exhalative sulfide deposit is presented in Fig. 17 (Plimer, 1986), which depicts an idealised time–space evolution of mineralizing systems in a rift structure, from deeper and higher temperature to shallower and lower temperature. In this model, exhalative stratabound/stratiform massive sulfide deposits are formed by the discharge of metalliferous brines rich in Cu, Pb, Zn, Ag, and Ba. The Broken Hill ore systems differs from the Mount Isa SEDEX deposit in that they are hosted in high-grade metamorphic rocks, do not exhibit bedding-parallel sulfides, have high concentrations of Mn, P and F, the absence of organic-rich lithologies, the presence of a magnetite association, and garnet-bearing alteration haloes (Large et al., 2005).

2.5.1. Non-sulfide Pb ore

The stratabound Magellan Pb deposit is hosted by outliers of the Earaheedy Group in the southeast of the Paleoproterozoic Yerrida Basin in the WAC (Fig. 1). The mineralization was accompanied by silicification and sericitic alteration of the host sandstone and silicified stromatolitic dolomite of the Yelma Formation close to, or at the unconformity with the underlying Windplain Group of the Yerrida Basin (Pirajno et al., 2004b and references cited therein). Magellan is an unusual mineral deposit containing cerrusite, plattnerite (PbO_2), coronadite ($PbMn_8O_{16}$), and pyromorphite as ore minerals (McQuitty and Pascoe, 1998). Its discovery was announced in 1993, with resources estimated at approximately 220 Mt @ 2.2% Pb (proven and probable ore reserves total 8.5 Mt @ 7.12% Pb). Immature wacke is one of the host rocks, and is characterised by pervasive sericitic and kaolinitic alteration. The lead minerals are paragenetically late and form replacement minerals of the matrix of the host rock. Trace element analyses of ore materials indicate anomalous abundances of Ba (1000–1828 ppm), Mn (1900–3672 ppm), and Cu (257–400 ppm). No sulfides or other metals are present. Homogenisation temperatures of primary fluid inclusions in quartz range from 180 to 220 °C, with salinities of 9–15 wt% NaCl eq. (McQuitty and Pascoe, 1998). It is possible, that this unusual type of deposit is associated with the migration of low-temperature late basinal fluids along permeable rocks such as sandstone.

However, the lack of sulfides and the unique presence of oxide minerals would suggest that the deposit might be related to palaeoweathering processes, under physico-chemical conditions, which were conducive to the oxidation and subsequence mobilization of Pb, which could have been sourced from the breakdown of K-feldspar during weathering and then mobilised as a soluble complex in groundwater (Biørlikke and Sangster, 1981). Alternatively, and perhaps more likely, the Pb carbonate could have resulted from the intense and deep weathering of a precursor MVT deposit, in which the sulfides were completely weathered out with removal of the more mobile elements such as S, Fe, and Zn. The Pb was re-precipitated as a carbonate due to the presence of abundant Ca in the degrading carbonate rocks (Pirajno, 2004a; Pirajno and Burlow, in preparation). Whereas non-sulfide Zn ore systems are well known (Hitzman et al., 2003), most are characterised by various combination of Cu, Zn, and Pb oxides and/or carbonates. To our knowledge, only one other deposit in Italy has been reported

to contain Pb carbonate ore hosted in sandstone (Fadda et al., 1998).

2.5.2. Stratiform epigenetic Cu–Zn

Nifty is a stratabound epigenetic deposit that is hosted by carbonate rocks and shale of the Broadhurst Formation in the Throssell Range Group. Copper is the only metal recovered from the deposit, but minor sphalerite, galena, and silver, and traces of gold and uranium minerals are also present (Dare, 1994). There are three types of Cu mineralization in the Nifty area: primary chalcopyrite; secondary, silicified carbonate-hosted Cu; and secondary shale-hosted mineralization (Dare, 1994). Current production is from the secondary (oxide) mineralization, which is mined as an open-cut operation. There is also a large, undeveloped, Cu sulfide resource estimated to be about 1.9 Mt of copper equivalent. The project has a total resource of about 150 Mt of ore grading 1.3% Cu.

The Nifty Cu deposit is hosted by structures included in the Miles Orogeny (regional D_{3-4}). Hydrothermal fluids were focused along early (D_4) thrust faults in a syncline, where chalcopyrite precipitation was controlled mainly by changes in pH accompanied by a slight decrease in temperature (Anderson et al., 2001). The mineralization and alteration styles at Nifty are similar to those of the copper orebodies at Mount Isa (Anderson et al., 2001). There is no evidence for the source of the copper.

2.6. Unconformity-related uranium

In the Pine Creek Orogen, major stratabound U deposits are hosted by partly carbonaceous greenschist to amphibolite facies metasedimentary rocks, and all deposits are within 100 m of the unconformably overlying late Paleoproterozoic Kombolgie Formation. These deposits are structurally controlled, and most are characterised by chlorite alteration in breccia or fractured rocks in faults or shear zones that are subparallel to the foliation and layering in host rocks (Ferguson and Goleby, 1980). Uranium tenor generally decreases with depth and does not extend above the unconformably overlying Kombolgie Formation, except at Jabiluka (Solomon et al., 2000). Most of the deposits are close to Neoarchean gneissic granite, which, where not leached, many have elevated U contents (McAndrew and Finlay, 1980).

The world class Jabiluka U deposit is located in the Alligator Rivers Uranium field in the Nimbuwah Domain (Fig. 5). There are two mineralized areas, Jabiluka I and Jabiluka II, with the latter being the largest with a resource of 31 Mt @ 0.53% U_3O_8 and 4.6 Mt @ 3.07 g/t Au (Polito et al., 2005). The ores are hosted by shallow-to-steeply dipping graphitic units of chlorite-biotite-muscovite schist in the Neoarchean Nanambu Complex (Fig. 5). The host rocks are unconformably overlain by shallowly dipping, coarse-grained sandstone of the Kombolgie Formation.

Uranium ore is located in semi-brittle shears and breccia that is subparallel to the units in the Nanambu Complex and developed at the hinge zones of fault-related folds (Polito et al., 2005). Uraninite (UO_2) veins are intimately associated with chlorite, sericite, hematite, quartz, and illite veins, which extend in to the basal 2 m of the Kombolgie Formation (Gustafson and Curtis, 1983). Post-diagenetic alteration has also taken place above the unconformity (Gustafson and Curtis, 1983). This suggests that there were periods of remobilization, and that the initial mineralization was early in the history of the Kombolgie Formation.

Electron microprobe and X-ray diffraction analysis of syn-ore illite and chlorite indicates a mineralization temperature of 200 °C and fluid inclusion studies suggest that the mineralizing brines were saline (Polito et al., 2005). Isotopic (U–Pb and $^{207}Pb/^{206}Pb$) studies indicate that uraninite first precipitated at ca. 1680 Ma, which is coincident with the $^{40}Ar/^{39}Ar$ ages of 1683 ± 11 Ma age of brine migration out from the Kombolgie Formation (Polito et al., 2005). This age appears to be about 60 million years younger than the age for the Ranger deposit.

At the Ranger deposit, a Paleoproterozoic sequence of volcanic and metasedimentary rocks unconformably overlie Neoarchean granitic gneiss of the Nanambu Complex. The Paleoproterozoic rocks are folded, faulted and sheared, and crosscut by east-trending ca. 1870–1860 Ma granite dykes and pegmatite veins, and gently dipping N–NE trending mafic dykes of the ca. 1690 Ma Oenpelli Dolerite (Hein, 2002). Regional metamorphism is greenschist facies and contact metamorphism is hornblende–hornfels facies. The rocks were deformed during the Maud Creek Event, and E–W extensional deformation (normal faulting) before or at the time of deposition of the Kombolgie Formation (Hein, 2002). The initial deposition of the U is dated at ca. 1740 Ma (Ludwig et al., 1987; Maas, 1989).

Several small U(–Au) deposits are located in the Paleoproterozoic Murphy Inlier bordering the southern McArthur Basin (Ahmad et al., 1999). The inlier comprises a greenschist facies sequence of wacke, siltstone, and shale that is intruded by ca. 1840–1820 Ma granite and unconformably overlain by ca. 1773 Ma felsic volcanics (Sweet et al., 1981; Ahmad et al., 1999). These rocks are unconformably overlain by conglomerate and sandstone included in the McArthur Basin. The mineral deposits are located in the basal sandstone and conglomerate in the McArthur Basin and consist of pitchblende-uraninite and minor brannerite with traces of galena, pyrite, marcasite, chalcopyrite, bornite and gold (Ahmad et al., 1999). The U mineralization followed a regional sericite–illite–chlorite–haematite alteration (Ahmad et al., 1999). The mineralization was either leached from the basal units in the McArthur Basin, presumably by oxidized solutions, and deposited in reducing environments, or leached from the basement and deposited in the McArthur Basin (e.g. Solomon et al., 2000).

The ca. 800 Ma Kintyre U deposit is the only mineable mineralization in the Paleoproterozoic Rudall Complex to date. Jackson and Andrew (1990) described the deposit as an unconformity-related, vein-type deposit, and compared it to the East Alligator River uranium deposits of the Northern Territory. The mineralization at Kintyre is hosted by chlorite(–carbonate)–quartz schist and chert in the hinge zone of an east–northeasterly plunging F_2 antiform, is close to the unconformably overlying Neoproterozoic Coolbro Sandstone of the Throssell Range Group in the Yeneena Basin, and is early in the history of the basin development (Maas and Bagas, in preparation). At Kintyre pitchblende veins contain chlorite, dolomite, ankerite, and calcite, with accessory bismuthinite, chalcopyrite, bornite, and galena, and locally significant gold and platinum group elements (Jackson and Andrew, 1990).

2.6.1. Associated hydrothermal PGE deposits

The Coronation Hill deposit in the Pine Creek Orogen is similar to unconformity-related uranium deposits and is located in strongly folded, metamorphosed, and hydrothermally altered Paleoproterozoic rocks that are unconformably overlain by the haematitic sandstone, sedimentary breccia, and conglomerate of the Kombolgie Formation. Both the basement rocks and the Kombolgie Formation are traversed by long-lived N- to NW-trending faults. The deposit includes Au–Pt–Pd(–Se–Sb–U) and has a resource estimate of 5.4 Mt grading 4.31 g/t Au, 0.65 g/t Pd, and 0.19 g/t Pt (Carville et al., 1990).

The ore at Coronation Hill consists of Au–PGE–U veins in steeply dipping zones in the basement rocks, and disseminated Au–PGE ore at or below the unconformity in felsic volcanics, quartz diorite, and

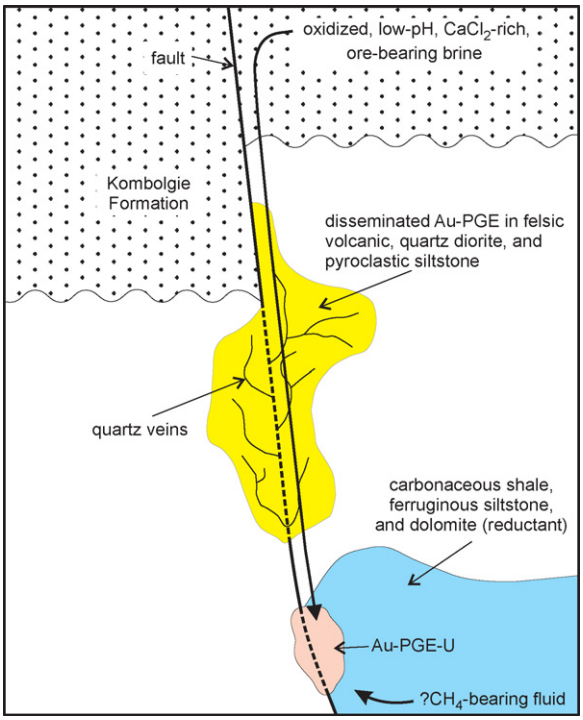


Fig. 18. Model for the formation of the Coronation Hill type of hydrothermal PGE ore system (modified after Mernagh et al., 1994). Oxidized, low-pH, CaCl2-rich, ore-bearing meteoric brine in a neutral-pH cover sequence flows structures in to basement rocks where the brine interacts with feldspathic or calcareous rocks. The increase of pH and the reduction in fO_2 in the fluid leads to precipitation of Au–PGE. Mixing with high pH? CH4-bearing fluid fluids from below the unconformity causes precipitation of Au–PGE–U.

pyroclastic siltstone (Fig. 18). The ore in the subvertical zones is structurally controlled in the form of tabular lenses that are up to 35 m wide are hosted in the N–NW-trending faults. The host rocks in this type of ore are metamorphosed igneous intrusions (quartz-feldspar porphyries and diorite), chloritic volcaniclastic rocks, and calcareous and carbonaceous sedimentary units (Mernagh et al., 1994, 1998). Detailed studies by Mernagh et al. (1994) indicate that the mineralization took place when highly oxidized and calcium-rich brines originating from the Kombolgie Formation reacted with feldspathic and reducing rocks and fluids in the basement at around 140 °C (Fig. 18).

3. Discussion and conclusions

As mentioned in "Section 1", Australia's Proterozoic mineral systems are best understood within the framework of the geological and geodynamic factors that control the genesis and preservation of the deposits. Australia's Proterozoic terranes were formed in intracratonic, convergent plate margin, or collisional tectonic settings. Figs. 4, 9 and 19 illustrate the main tectonic settings of Australia's Proterozoic mineral systems.

The iron-formations of the Hamersley Basin probably constitute the largest endowment of Fe in the world. Barley et al. (1997) proposed to link flood-basalt volcanism, with large input of Fe^{2+} into the ocean and the origin of the banded iron-formations of the Hamersley Basin. Epsilon Nd signatures ($\varepsilon Nd = +1.0 \pm 0.5$) of BIF from the Hamersley Basin confirm volcanic and/or hydrothermal mantle sources (Alibert and McCulloch, 1993). Isley (1995), and Isley and Abbott (1999) also suggested a link between plume-related igneous activity and the deposition of iron-formations during the Archean and Proterozoic. They used the geochronological record of plume activity between 3800 and 1600 Ma, as indicated by the presence in the geological record of komatiites, flood basalt, mafic dyke swarms, and mafic–ultramafic layered intrusions. Condie et al. (2000) proposed that the world-wide occurrence of passive margins, platform sediments and iron-formations at around 1900 Ma could be indicative of a superplume event (see also Barley et al., 1997; Condie et al., 2000).

Intraplate tectonothermal events involve magmatic underplating and extensional tectonics, driven by mantle dynamics in the form of deep mantle plumes or asthenospheric upwelling in rift settings associated with far field tectonic stresses. The predominance of bimodal magmatic systems in several of the Australia's Proterozoic terranes (Pine Creek; McArthur River-Mount Isa, Broken Hill) and the presence of dyke swarms and large igneous provinces support intracratonic tectonic scenarios (Figs. 9 and 19). Arc-terranes and their lateral accretion have been recognised in regions such as the Arunta Orogen and Musgrave Complex (Betts et al., 2002; Wade et al., 2006), the Gascoyne Complex (a remnant of an Andean-style arc terrane, which was extensively deformed and metamorphosed during the ca. 2000–1960 Ma Glenburgh Orogeny; Sheppard et al., 2005), and the Halls Creek Orogen (Griffin et al., 2000). Şener et al. (2005) suggested Cordilleran style tectonics at ca. 1740 Ma that eventually led to the amalgamation of the NAC, SAC and WAC, was accompanied by the widespread development of orogenic Au lodes, such as those of Mount Olympus in the WAC, and the Pine Creek and Tanami Au provinces in the NAC. However, the ca. 1800 Ma age of lode deposits in the NAC and the evidence for collision between the WAC and NAC during the ca. 1800–1765 Ma Yapungku Orogeny (Smithies and Bagas, 1997; Bagas, 2004) indicate that this model needs to be refined. Betts and Giles (2006) and Betts et al. (2002) envisaged a north-dipping subduction system or zone on the southern margin of the NAC, with concomitant continental back-arc extension, which would have formed the McArthur-Mount Isa rift-system (see Figs. 15 and 19).

The plate margin and subduction-related tectonics of Australia's Proterozoic are lacking in calc-alkaline volcano-plutonic rocks and related mineral systems (e.g. porphyry and epithermal deposits), which are typically found in these tectonic settings. It is possible that these mineral systems may be absent due to erosion and/or metamorphism-deformation; the latter perhaps preventing recognition. In the Halls Creek Orogen, for which a subduction tectonic setting followed by collision is well established (Griffin et al., 2000), low-sulfidation adularia-sericite epithermal Au has been recently discovered in the 1855 Ma Whitewater Volcanics (Northern Star Resources Annual Report, 2005; www.nsrltd.com/projects.html).

The thermal impact of mantle plumes and associated magmatism cannot be underestimated (Fig. 9). In addition to direct generation of magmas, thermal anomalies associated with mantle plumes constitute powerful heat sources in the crust. Plumes may be responsible for the inception of crustal-scale hydrothermal circulation and high-T and low-P metamorphism, which may result in a wide range of ore deposits in hotspot-related rift systems (Pirajno, 2000, 2004b). An example, discussed in Pirajno (2004a), is provided by the numerous sub-economic hydrothermal veins present throughout tectonic units of the Capricorn Orogen, which are indicative of widespread crustal-scale hydrothermal circulation in brittle to brittle–ductile structures. High geothermal gradients linked to continental mafic magmatism may have favoured large-scale circulation of hydrothermal fluids, which would have been responsible not only for the emplacement of the vein and lode deposits but also in the modification and remobilisation of earlier syngenetic and/or epigenetic deposits into structurally prepared sites (e.g. see Figs. 9 and 12).

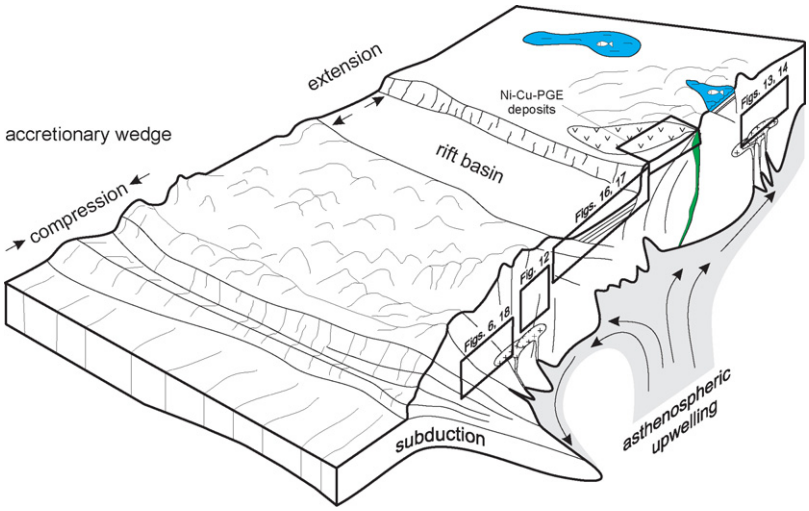


Fig. 19. Collision and back-arc rift settings of Proterozoic mineral systems in Australia. The suggested setting for the genetic models shown in Figs. 6, 12–16 and 17 are indicated in this model. Ni–Cu–PGE orthomagmatic deposits are related to mafic/ultramafic intrusions and continental flood basalts (see also Fig. 9). The model is schematic and not to scale and is only designed to provide a holistic view. Other ore systems, such as iron-formations are shown in Fig. 4.

IOCG deposits appear to be linked to extensional settings with episodes of anorogenic bimodal and alkaline magmatism. In IOCG systems, the unifying factors are the enrichment in iron, phosphorous, fluorine and widespread alkali metasomatism in the host rocks (Hitzman et al., 1992; Oreskes and Hitzman, 1993). Typically, IOCG hydrothermal systems form in shallow (4–6 km deep) crustal environments and are the expression of volatile-rich, alkaline magmas (Hitzman et al., 1992). As mentioned above, several IOCG deposits cover the time span of approximately 1800–1100 Ma and they appear to be linked to planetary-scale rifting events and the assembly and breakup of supercontinents. The assembly of the Proterozoic supercontinent Rodinia, may have acted as a large insulation blanket on sublithospheric mantle flow, resulting in accumulation of heat, rise of mantle diapirs, melting, rifting of the continental crust, and finally the inception of regional-scale hydrothermal systems at shallow levels in the crust. An alternative view is that of Barton and Johnson (1996) who suggested that Fe oxide–Cu–Au–REE–U mineralization is common in both Proterozoic and Phanerozoic extensional environments. Although these authors did not dispute the link with continental hotspots, they proposed that the saline fluids responsible for these deposits are sourced from evaporites. They cited examples of Holocene hydrothermal Fe oxides formed from evaporitic sources and the correlation of the IOCG of Mesozoic age with zones of low-latitude as revealed by continental reconstructions. The evaporitic source model of Barton and Johnson (1996) suggests that circulation of hydrothermal fluids is caused by magmatic heat, that igneous rocks are the source of the metals, and that the metal transport is effected by chlorides supplied by evaporitic deposits (see Fig. 14). In addition, the widespread Na-alteration associated with the deposits is also related to the evaporites, which supplied the large amounts of sodium to the hydrothermal fluids.

The world-class stratiform massive sulfide deposits in the McArthur River-Mount Isa rift-system may ultimately owe their genesis to mantle-generated heat. Genetic models for the McArthur River-Mount Isa and Broken Hill sedimentary-hosted stratiform and stratabound deposits have been discussed in detail by Large et al. (2005) and Plimer (1986). The host rocks comprise packages of metamorphosed evaporite–carbonate–clastic rocks and the sulfide mineralization is clearly associated with organic-rich siltstone and shale. A widely accepted model proposes the activity of hypersaline brines that discharge in a reduced environment (e.g. bottom of anoxic basin or lake; Large et al., 2005). This model suggests bacterial mediation with biogenic sulfate reduction being a major contributor of H_2S leading to sulfide precipitation. The range of $\delta^{34}S$ values from −13 to +30‰ in the ores (Large et al., 2005) support sulfate reduction. The alteration patterns are characterised by regional-scale Na and K metasomatism, phyllosilicate alteration, Fe-Mn carbonates, and silicification (Large et al., 2000). Regional alkali-feldspar (microcline, adularia and albite) metasomatism is associated with major fault zones along which the giant base-metal deposits are situated. Hydrothermal fluids and heated basinal brines were driven by the geothermal gradient in the rift basins with metals sourced from mafic and felsic volcanic and volcaniclastic rocks adjacent to sandstone aquifers (Large et al., 2005).

Australia's Proterozoic mineral endownment is probably one of the most varied and richest in the world. Major advances in the understanding of these mineral systems have been made in the last 30 years, in terms of tectonic settings and mechanisms of ore genesis, largely due to precise age dating using the U–Pb, K–Ar and Ar–Ar systems. Nevertheless, much remains to be done, especially in view of the ever increasing demand of metals provided by Australia's vast mineral resources.

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