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Tectono-metallogenic systems — The place of mineral systems within tectonic evolution, with an emphasis on Australian examples

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ABSTRACT

Tectono-metallogenic systems are geological systems that link geodynamic and tectonic processes with ore-forming processes. Fundamental geodynamic processes, including buoyancy-related processes, crustal/lithospheric thinning and crustal/lithospheric thickening, have occurred throughout Earth's history, but tectonic systems, which are driven by these processes, have evolved as Earth's interior has cooled. Tectonic systems are thought to have evolved from magma oceans in the Hadean through an unstable "stagnant-lid" regime in the earlier Archean into a proto-plate tectonic regime from the late Archean onwards. Modern-style plate tectonics is thought to have become dominant by the start of the Paleozoic. Mineral systems with general similarities to modern or geologically recent systems have been present episodically (or semi-continuously) through much of Earth's history, but most of Earth's present endowment of mineral wealth was formed during and after the Neoarchean, when proto- or modern-style plate tectonic systems became increasingly dominant and following major changes in the chemistry of the atmosphere and hydrosphere. Changes in the characteristics of some mineral systems, such as the volcanic-hosted massive sulphide (VHMS) system, reflect changes in tectonic style during the evolution towards the modern plate tectonic regime, but may also involve secular changes in the hydrosphere and atmosphere.

Whereas tectono-metallogenic systems have evolved in general over Earth's history, specific tectono-metallogenic systems evolve over much shorter time frames. Most mineral deposits form in three general tectono-metallogenic systems: divergent systems, convergent systems, and intraplate systems. Although fundamental geodynamic processes have driven the evolution of these systems, their relative importance may change as the systems evolved. For example, buoyancy-driven (mantle convection/plumes) and crustal thinning are the dominant processes driving the early rift stage of divergent tectono-metallogenic systems, whereas buoyancy-driven processes (slab sinking) and crustal thickening are the most important processes during the subduction stage of convergent systems. Crustal thinning can also be an important process in the hinterland of subduction zones, producing back-arc basins that can host a number of mineral systems. As fundamental geodynamic processes act as drivers at some stage in virtually all tectonic systems, on their own these cannot be used to identify tectonic systems. Moreover, as mineral systems are ultimately the products of these same geodynamic drivers, individual mineral deposit types cannot be used to determine tectonic systems, although mineral deposit *associations* can, in some cases, be indicative of the tectono-metallogenic system.

Ore deposits are the products of geological (mineral) systems that operate over a long time frame (hundreds of millions of years) and at scales up to the craton-scale. In essence, mineral systems increase the concentrations of commodities through geochemical and geophysical processes from bulk Earth levels to levels amenable to economic mining. Mineral system components include the geological (tectonic and architectural) setting, the driver(s) of mineralising processes, metal and fluid sources, fluid pathways, depositional trap, and post-depositional modifications. All of these components link back to geodynamic processes and the tectonic system. For example, crustal architecture, which controls the spatial distribution of, and fluid flow, within mineral systems, is largely determined by geodynamic processes and tectonic systems, and the timing of mineralisation, which generally is relatively short (commonly <1 Myr), correlates with local and/or far-field tectonic events.

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The geochemical characteristics of many mineral systems are a consequence of their geodynamic and tectonic settings. Settings that are characterised by low heat flow and lack active magmatism produce low temperature fluids that are oxidised, with ore formation caused largely by redox gradients or the provision of external H_2S. The characteristics of these fluids are largely governed by the rocks with which they interact, rocks that have extensively interacted with the hydrosphere and atmosphere, both environments that have been strongly oxidised since the great oxidation event in the Paleoproterozoic. In settings characterised by high heat flow and active magmatism, ore fluids tend to be higher temperature and reduced, with deposition caused by cooling, pH neutralisation, depressurisation and fluid mixing. Again, the characteristics of these fluids are governed by rocks with which they interact, in this case more reduced magmatic rocks derived from the mantle or lower crust.

1. Introduction

In the 1960s, the geoscientific world experienced a scientific revolution in which previous models of basin development and orogenesis were superseded by the new paradigm of plate tectonics. Although the concept of continental drift had been around for many decades (e.g. Wegener, 1912), the development and acceptance of the plate tectonic paradigm were driven by the availability of high quality geophysical data along mid-ocean ridges (Vine and Matthews, 1963) and the conceptualisation of processes that can drive plate motion (Holmes, 1929; Dietz, 1961a; Hess, 1962). Today, long-term global positioning system deployments worldwide constrain continental drift rates and directions confirming, beyond doubt, plate tectonics as the pre-eminent conceptual template within which all Earth processes must be interpreted. The concept that the Earth's lithosphere is divided into a number of plates that move relative to each other is central to the plate tectonic paradigm. Importantly, tectonic processes, such as subduction or spreading, occur along the margins of these plates, along which relative plate motion occurs. Although tectonic activity in the oceanic lithosphere is focused along narrow zones, tectonic activity on continental lithosphere tends to be less focussed.

Historically, many mineral deposits have been classified largely based upon local features, such as the characteristics of the ores, the alteration assemblages, the host rocks and the ore fluids, and genetic considerations such as the relative timing of mineralisation (e.g. Lindgren, 1933). As this style of classification was largely empirical, it could readily be applied to mineral exploration. However, as exploration has become more difficult and sophisticated, easily discovered (outcropping) deposits have been progressively found and exploration has been forced to undercover areas, particularly in more mature exploration destinations such as Australia and Canada. In recent years, this challenge has been partially met by increasing exploration within known mineralised districts, thereby adding resources to existing deposits. But this has led to decreases in greenfield exploration, both in absolute and relative terms over the last fifteen years (Schodde, 2011).

Several studies (Huleatt and Jaques, 2005; Jaques et al., 2005; Hronsky and Schodde, 2006) have shown that the largest discoveries are commonly made early in the history of mineral districts. As the search space decreases in brownfield areas, the opportunity for discovery of major new deposits decreases; many of the most significant discoveries in Australia over the last two decades have been made in greenfield areas (e.g. Bronzewing, Nebo-Babel, Tropicana, Nova, Gruyere and DeGrussa). Hence, survival of the mining industry in Australia, and elsewhere, requires continued exploration in greenfield areas that are commonly covered by alluvium, colluvium or younger basins and more difficult to explore using techniques and exploration models developed in regions where orebodies crop out at the surface.

One aspect of mineral deposits that has gained traction over the last few decades is the linkage of mineralisation to tectonic, particularly plate tectonic, processes. Sillitoe (1972) and Hutchinson (1973) were among the first to make this linkage when they proposed that porphyry copper and volcanogenic massive sulphide deposits are associated with subduction. Meyer (1981, 1988) reviewed the temporal distribution of mineral deposits and discussed how this distribution was related to the tectonic evolution of the earth. Sawkins (1984) assessed the relationship between mineral deposits and tectonic settings, demonstrating that many types of deposits are restricted to specific tectonic settings. More recently, Kerrich et al. (2000, 2005) reviewed the distribution of many mineral deposits and found that many have specific associations with tectonic settings. In this contribution we extend the work of Meyer (1981, 1988), Sawkins (1984) and Kerrich et al. (2000, 2005) to examine how mineralising systems fit into the evolution of tectonic systems, and, more importantly, how the character of mineral systems changes within tectonic cycles. However, to provide context to this analysis, we present discussions on how the fundamental geodynamic process of heat flow drives tectonism and mineralisation, and controls the formation and evolution of sedimentary basins, magmatic provinces and metamorphic belts. Then we present an overview of the important features of mineral systems and how they are linked to tectonic processes to form tectono-metallogenic systems. Although this contribution concentrates on the relationship of metallogenesis to tectonics, it has long been established that other factors, most importantly the oxygenation of the atmosphere, have had a major influence on metallogenesis. For further information on these other controls, the reader is referred to Meyer (1981, 1988) and papers in the special issue of *Economic Geology* on secular variations in ore deposits through time (Goldfarb et al., 2010).

Three types of tectono-metallogenic systems, those associated with divergent and convergent margins and in intraplate environments, are then analysed to better understand the linkages between tectonics and metallogenesis. This understanding can be used to better predict and explore for mineralisation in under-cover greenfield areas. This contribution is not meant to be a comprehensive description of mineral deposits in Australia, but a synthesis as to how these deposits relate to each other tectonically and geodynamically. Comprehensive descriptions of Australian mineral deposits can be found in various Australasian Institute of Mining and Metallurgy Monographs (Knight, 1975; Hughes, 1990; Berkman and Mackenzie, 1998) and in Solomon and Groves (2000).

2. Geodynamics: the driver of tectonism and mineralisation

The fundamental driver of all geodynamic and tectonic processes on Earth is the escape of latent and radiogenic heat from the Earth's interior to its surface (Holmes, 1929). However, throughout its history Earth appears to have responded to this heat loss through a number of different processes with a number of end-results, including magma oceans, stagnant-lids and plate tectonics (e.g. Sleep, 2000). Of these, the first two appear to have been most important during the early history of Earth (Hadean to Paleoarchean), with plate tectonics becoming increasingly important from the Mesoarchean (Section 9.1). As the vast majority of mineral deposits formed during the Neoarchean or later, we consider mineral systems in terms of the plate tectonics paradigm below, although we note that other styles of tectonics (e.g. stagnant lid) may have been important, particularly early in Earth's history, and that plate tectonics has changed over the history of the Earth (Section 9.1).

Geodynamic drivers of plate tectonics include buoyancy-driven processes (e.g. sinking of dense, cold lithosphere and mantle convection), which cause lithospheric thinning, and lithospheric thickening. The interaction of these processes produce the four broad tectonic environments presently observed on Earth - divergent margins, convergent margins, transform margins and intraplate settings. The three fundamental processes operate within most of these tectonic environments. For example, although convergent margins are dominated by the subduction of dense crust and often involve crustal thickening, parts of overall convergent systems, such as back-arc basins, are dominated by lithospheric thinning, and entire convergent margins can remain in extension for extended periods (e.g. Collins, 2002a).

Most active and geologically young mineral deposits are spatially and genetically associated with plate boundaries (Hannington et al., 2005; Sillitoe and Perelló, 2005). Similarly, many ancient mineral deposits are closely associated with paleo-crustal boundaries that may have marked plate boundaries in the geological past (e.g. Begg et al., 2010; Groves et al., 2010). Empirically, these relationships suggest that many mineral systems are linked to tectonic processes that operate at plate margins. Importantly these processes appear to have evolved through geological time, and respective changes are reflected in temporal distribution and characteristics of mineral deposits. Some mineral deposits (e.g. salt lake-hosted potash deposits in Australia: Mernagh et al., 2013), however, do not have an association with plate margins or boundaries, suggesting that some mineral systems are associated with intraplate processes.

An important consequence of a specific geodynamic process operating in several tectonic settings is that mineralising systems driven by that geodynamic process can occur in different tectonic settings. As an example, modern black-smoker deposits, which form as a consequence of new crust formation during rifting, occur both along mid-ocean ridges, an example of a divergent tectonic system, and in rifted arcs or back-arc basins in convergent tectonic settings (Hannington et al., 2005). Some other deposit types seem to be more restricted in time and space. For example, orogenic gold deposits, which are associated with crustal thickening that commonly follows a period of crustal thinning (Bleeker, 2012), are most commonly associated with convergent margins (Groves et al., 1998; Kerrich et al., 2000). Hence, the presence of individual deposit types can be used as an indication of the geodynamic process driving mineralisation, but, more importantly, the presence of a number of different deposit types in close proximity in time or space, or mineral deposit associations, may be indicative of the overall tectonic system. For example, the presence of volcanic-hosted massive sulphide, orogenic gold and/or porphyry Cu deposits within a single tectonic system might indicate the presence of a previous convergent margin.

3. The geodynamic and tectonic settings of sedimentary basins

Sedimentary basins can be subdivided into three categories based on the geodynamic subsidence mechanism that formed them (Fig. 1; after Allen and Allen, 2005). Approximately 70% of the world's sedimentary basins were formed as a consequence of lithospheric thinning (e.g. rift basins), 20% as a consequence of lithospheric loading (e.g. foreland basins) and 10% in response of convection within the earth's mantle (e.g. intra-cratonic basins) (N. White and K. Czarnota, unpublished data). Each basin type is associated with distinct lithospheric architectures, subsidence histories and heat-flow regimes, factors that strongly influence the types of mineral systems hosted by basins. Temporally, one basin type can be stacked on top of another.

3.1. Rift basins

McKenzie (1978) showed that rift basins are characterised by two stages of subsidence (Fig. 1A). An initial rapid phase of subsidence occurs in response to mechanical thinning of the lithosphere which typically lasts ~20 Myrs and may result in sea floor spreading (rift phase; Bown and White, 1995; Newman and White, 1999). During this phase of basin formation, sediments thicken towards extensional faults displaying either graben or half graben geometries (Bosworth et al., 2005). Often an increase in heat flow over the basin occurs as the reduction in the crustal component of heat-flow, resulting from thinning of the radiogenic crust, is offset by an increase in the mantle component of heat-flow. Decompression melting of the mantle is also common late in the syn-rift phase provided elevated mantle potential temperatures and/or high strain rates occur during rifting (White and McKenzie, 1989; Bown and White, 1995).

The second, slower phase of subsidence in rift basins (thermal subsidence phase) is driven by cooling of asthenosphere beneath the thinned lithospheric mantle until the mantle lithosphere attains its pre-rift

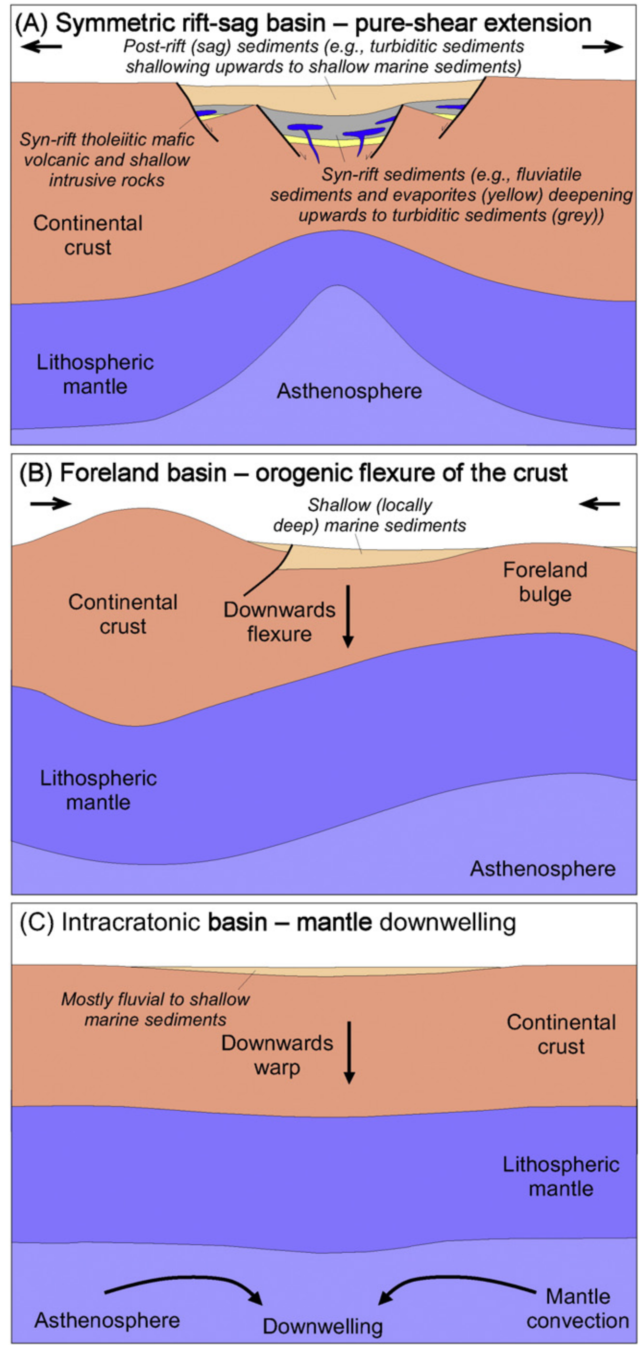

Fig. 1. Schematic diagrams showing geodynamic setting and characteristics of major basin types: (A) rift-thermal subsidence basin, (B) flexural basin, and (C) intracratonic basin.

thickness. The duration and amplitude of this phase of subsidence is a function of the degree the lithosphere has been stretched and its original thickness (Crosby et al., 2010). Sediments show gradual thickening towards the centre of the basin with an absence of faulting apart from small compaction driven structures and onlap relationships to underlying sediments (White and McKenzie, 1988). As the rate of subsidence decreases exponentially, the stratigraphic record becomes more sensitive to high frequency eustatic sea-level oscillations and long-wavelength dynamic topography effects (Van Wagoner et al., 1988).

McKenzie's (1978) pure shear model, whereby the degree of lithospheric thinning is similar at all depths, was challenged for many years by the concept that the degree of lithospheric thinning changes with depth, resulting in an offset between syn-rift and post-rift basin depocentres (Wernicke, 1985). It is now generally accepted that rift basins formed by small degrees of stretching of the lithosphere are a consequence of pure shear, whereas at higher degrees of stretching resulting in passive margin formation, depth dependency is the exception rather than the rule (White et al., 2003; Crosby et al., 2011).

The abundance of rift basins relative to other basins in the geological record is a function of their compensated isostatic state which enables preservations (Allen and Allen, 2005). Basins at dilational jogs along strike-slip faults (pull-apart basins) are small examples of rift basins as they form as a consequence of lithospheric thinning.

3.2. Flexural basins

Flexural basins form by the bending of the lithosphere in response to crustal loading associated with orogenesis. Foreland basins are the main expression of basins resulting from loading. There are two main types: peripheral (on the subducting plate) and retroarc (on the overriding plate: DeCelles and Giles, 1996). Basin width (35–400 km) is controlled by the effective elastic thickness of the lithosphere, and the depth of the basin is controlled by the applied load (Watts, 2001). Deformation of previously extended regions results in narrow basins (e.g. Pyrenees, Spain) whereas intra-cratonic deformation over regions of thick lithosphere results in wide basins (e.g. Ganges Basin, India, DeCelles et al., 1998; Officer Basin, Central Australia, Haddad et al., 2001; Melbourne Zone, SE Australia, Cayley et al., 2002). Both peripheral and retroarc basins characteristically display a wedge shape geometry with depocentres adjacent to the orogenic frontal thrust and low geothermal gradients. As orogenesis progresses and the frontal thrust propagates into the basin a portion of the depocentre is uplifted and eroded. The poor preservation of foreland basins within the geological record is a function of their uncompensated isostatic state. As the crustal load responsible for the flexure is removed, the basin rebounds and is eroded. The orogenic load results in the establishment of a large hydraulic head adjacent to the depocentre thereby promoting fluid flow from the orogenic front into the basin. Foreland basin fill is sediment-dominated, with compositions reflecting the nature of the source-terrane, generally a continental metamorphic and igneous mix (Allen and Allen, 2005).

3.3. Intra-cratonic basins

There is a lack of consensus as to the main driver of subsidence in intracratonic basins that are not related to extensional or thrust faulting. Nevertheless, there is a growing appreciation that convective circulation within the mantle can be a significant driver of sedimentary basin formation and modification. Convective upwelling and down welling within the mantle can result in dynamic topography at the earth's surface with amplitudes of up to 2 km (typically 1 km) over wavelengths of 500 km. This topography can develop at rates of up to 200 m/Myr (Roberts and White, 2010; Czarnota et al., 2013; Winterbourne et al., 2014). Deposition of the Mesozoic to Cenozoic intracontinental Eromanga Basin in eastern Australia and north American intracratonic basins have been linked to the presence of subducted slabs beneath these regions and mantle flow (Matthews et al., 2011; Burgess, 2008). Conversely uplift and erosion above mantle upwelling has been implicated as the subsidence mechanism for the Centralian Superbasin following waning of mantle upwelling and related mafic magmatism (Zhao et al., 1994). Recently, plate-margin interactions have been implicated in the evolution of adjacent intraplate regions, with continent-wide tilts influencing intracratonic basin formation/evolution and causing regional-scale submergences and uplifts (e.g. Sandiford, 2007). Given most intracratonic basins form over regions of thick lithosphere it is unlikely that variations in sub-plate mantle temperature result in a difference in heat flow at the Earth's surface. Most intra-cratonic basins are in-filled by fluvial to shallow marine sediments with distinct transgression and regression cycles bounded by significant unconformities extending over multiple cratonic blocks (Sloss, 1963; Walter et al., 1995). Total stratigraphic thicknesses are typically low, on the order of a few kilometres – yet a stratigraphic record spanning over 500 Myrs may be preserved (Burgess, 2008).

3.4. Lithology and implications to mineral systems

Sedimentary basin fill is controlled by rates of formation of accommodation space, eustatic sea-level variation and sediment supply. Given these controls, sedimentary facies stacking trends in themselves are non-diagnostic of the geodynamic or tectonic setting, although certain sediment stacking trends are more prevalent in certain basin types. For example, the rate of syn-rift subsidence often outstrips sediment supply resulting in under-filled depression and the deposition of deepening upwards sedimentary successions including black shales and/or evaporites depending on paleogeography (Prosser, 1993). Foreland basins are often associated with fining upward and then coarsening upward successions (Sinclair et al., 1991). In order to deduce the geodynamic control on basin formation, it is necessary to combine architectural, stratigraphic and subsidence information.

The type of basin exerts a significant control on the characteristics of basinal mineral systems and the likelihood of preservation. Anomalously high heat flow, which can drive fluid flow through convection, is mostly restricted to the rift phase of rift basin evolution. Fluid flow drivers in the thermal subsidence phase of rift basins, foreland basins and intra-cratonic basins are generally topographic, as these are typically associated with lower heat flow, although the sediments in these basins can act as a thermal insulator which may drive convective fluid flow provided the basement is enriched in radioactive elements (Majorowicz et al., 1984; Hand and Sandiford, 1999). In all basins, compaction is a significant mechanism driving hydrocarbon and basinal brine flow. Finally, the architecture and fluid flow pathways will also be influenced by the type of basin. Although faults will be developed in all basin types, commonly reflecting basement architecture, extensional faults are more common and most effectively facilitate vertical fluid flow during basin formation and later reactivation.

As discussed above, the preservation potential of a basin depends to a large extent upon the tectonic environment in which they form. Rift basins have a higher potential for preservation, particularly relative to foreland basins. As a consequence, deposits formed in rift basins also have a higher preservation potential.

4. The geodynamic and tectonic settings of igneous rocks

On modern Earth, igneous rocks are known from most geodynamic and tectonic environments (e.g. Wilson, 1989); these include rift zones (oceanic or continental lithosphere), convergent zones (i.e. oceanic and continental subduction settings and continental collision zones), and intraplate environments (oceanic or continental). Within these geodynamic environments, geochemical signatures are often ambiguous, especially for felsic magmatism, although a number of geochemical discrimination schemes and geochemical proxies for specific geodynamic settings have been widely employed in studies of ancient mafic and felsic rocks (e.g. Pearce and Cann, 1973; Pearce and Norry, 1979; Winchester and Floyd, 1977; Wood et al., 1979; Pearce, 1982, 1986; Pearce et al., 1984).

Mantle-derived magmatism is produced by a variety of processes: passive upwelling (mid-ocean ridge, back-arc basins: Langmuir et al., 1992; Klein, 2003), active upwelling, such as mantle plumes (within-plate: White and McKenzie, 1989), and flux melting in the (hydrated) metasomatised mantle wedge (e.g. subduction and early back-arc settings: Tatsumi and Eggins, 1995; Taylor and Martinez, 2003; Stern, 2010). Compositions are dependent on a range of variables, including the degree of melting, pressure and temperature of melting, source composition, volatile content, as well as post-melt modification by processes like fractionation, assimilation, contamination, and mixing (McKenzie and Bickle, 1988; Langmuir et al., 1992; Tatsumi and Eggins, 1995; Taylor and Martinez, 2003).

Nevertheless, a number of generalisations can be made. Regions of moderate and larger degrees of mantle melting (mid-ocean ridges, plume-related(?) oceanic plateaux and continental flood basalt provinces) are dominated by tholeiitic compositions (e.g. Wilson, 1989) with increasingly more Mg-rich (parental) melts (including picritic and ultramafic compositions) in plume-related magmatism, especially during, though not confined to, the Archean (Arndt, 2003; Herzberg et al., 2007, 2010; Kerr and Mahoney, 2007). These melts tend to be more homogeneous (although still variable, e.g. Klein and Langmuir, 1987; Kerr and Mahoney, 2007; Sobolev et al., 2007) within the oceanic environment, relative to continental flood basalt provinces where crustal contamination, high-level fractionation and/or involvement of mantle lithosphere may occur (e.g. Hergt et al., 1989). Smaller degrees of partial melting are characterised by more alkaline compositions (e.g. in continental rifts, intraplate environments and ocean islands: McKenzie and Bickle, 1988; Farmer, 2003; Jung et al., 2013; Price et al., 2014). Unusual melts, such as carbonatites, kimberlites, lamproites, utrapotassic melts (e.g. leucitites), are typically small volume melts of asthenosphere and/or possibly metasomatised lithospheric mantle (e.g. Farmer, 2003; Foley et al., 2012) — certainly they have a strong association with continental lithosphere (Mitchell, 1995; Tappe et al., 2006). It should be borne in mind that these are, of course, generalisations and numerous examples exist of associated tholeiitic and alkaline magmatism in flood basalt provinces, intraplate and ocean islands settings (Farmer, 2003; Clague and Dalrymple, 1987; Ernst and Bell, 2010; Price et al., 2014).

The major exception to the above generalisations is in convergent margin settings where partial melting is largely driven by fluid-fluxed melting (i.e. introduction of water and other volatiles into the mantle by the subducting slab: Gill, 1981; McCulloch and Gamble, 1991; Stern, 2002; Tatsumi and Eggins, 1995). Rocks in such settings are characterised by a consistent geochemical signature, including enrichment in the large ion lithophile elements (LILEs; especially the light rare earth elements (LREE)) and depletion in Nb, Ta and Ti relative to other incompatible elements (Tatsumi and Eggins, 1995). The enrichment in LILEs is commonly thought to reflect metasomatism of variably-depleted mantle wedge material by components derived from dehydration or partial melting of the subducting slab (and accompanying sediments; e.g. McCulloch and Gamble, 1991; Tatsumi and Eggins, 1995), whereas the depletion of Nb, Ta and Ti is thought to relate to either sequestration in rutile (and other Ti-bearing minerals) and/or perhaps prior melt-depletions (e.g. Gill, 1981; Woodhead et al., 1993; McCulloch and Gamble, 1991). Such rocks are typically calc-alkaline, though other rock types do occur.

These other rocks can include tholeiites (though still with an arc signature) that are largely confined to island arc settings (Tatsumi and Eggins, 1995), alkaline rocks, such as shoshonites, which are often either located further behind the arc or late in the history of the arc (including post-collisional; Morrison, 1980), and boninites, which are derived from metasomatised, highly-depleted mantle and thought to be largely generated in supra-subduction zone settings (Crawford et al., 1989).

Back-arc regions can be dominated by tholeiitic magmatism without an arc signature; however, even these often have hybrid signatures, between mid-ocean ridge and arc basalts (e.g. Taylor and Martinez, 2003). Importantly, rocks derived in arc settings are often more strongly oxidised than other mantle melts (Gill, 1981; Parkinson and Arculus, 1999; Kelley and Cottrell, 2009; Richards, 2015), a feature which has a bearing on a number of mineral associations. The calc-alkaline signature is not, however, unique, as it is carried by the majority of continental crust rocks and any involvement of such crustal material will impart this signature to some degree. This is a problem in general with mantle-derived magmatism, that is identifying compositional effects due to crustal contamination versus those that may be primary. This includes potential contributions from the mantle lithosphere (e.g. Hergt et al., 1989).

Felsic magmatism is more ambiguous than mantle-derived magmatism, and very few generalisations can be made about the associated geodynamic environment: felsic magma compositions are largely non-unique. This mostly reflects the strong primary controls continental crust has over the geochemistry of generated melts: the geochemistry of the magma will largely mirror the type and nature of the source rocks (e.g. immature versus more mature metasedimentary rocks, intermediate or mafic igneous rocks, high-grade gneisses) which can be mixed (e.g. Collins, 1996) and therefore not unique to the geodynamic environment. This is invariably the case even where substantial mixing with mantle-derived melts can be demonstrated. Despite this, some generalisations can be made with regard to broad geodynamic (but not tectonic) environment, often on the basis of the relative timing, chemistry, and distribution of magmatism (Collins, 2002b), and the temperature and/or pressure of magma generation. For example, Sr-undepleted, Y-depleted compositions, such as those found in many Archean trondjemite–tonalite–granodiorite (TTG) and in adakitic rocks in some younger arc settings (e.g. Martin et al., 2005; Gromet and Silver, 1987; Defant and Drummond, 1990), are commonly attributable to partial melting at pressures high enough to stabilise garnet and destabilise plagioclase. This process is thought to occur in either strongly thickened or underplated crust or via partial melting of a subducting slab (e.g. Gromet and Silver, 1983; Defant and Drummond, 1990), typically in a convergent environment. Similarly, higher temperature felsic rocks, often indicated by A-type (elevated HFSE) and/or alkaline composition, are associated with extensional environments such as rifts or back-arcs or post-collisional settings (e.g. Eby, 1990).

With regard to felsic magmatism, the associated mantle-derived rocks are often more diagnostic of geodynamic and tectonic environment, and may help to distinguish, for example, felsic rocks generated in an arc to those in the extensional back-arc, both settings which are important for specific mineral systems. In some regards, even the general igneous associations can be informative: bimodal associations with tholeiitic rocks are suggestive of extension and rifting, and linear belts of calc-alkaline rocks of mafic to intermediate and felsic compositions are more indicative of arc settings. Finally, it should be noted that, volume-wise, three tectonic environments dominate mafic and felsic magma generation: mid-ocean ridges; subduction settings including back-arc zones; and intraplate environments where large igneous provinces form. Only the latter two are likely to be preserved in the rock record.

5. The tectonic setting of metamorphic rocks

There are several types of metamorphism that can play a role in the formation of mineral systems. Orogenic metamorphism, as defined by Miyashiro (1973) (dynamothermal metamorphism of Winkler, 1979), includes metamorphism occurring over a large area including deformation-related, burial and ocean-floor metamorphism. Orogenic metamorphism is characteristic of convergent margin orogenic belts, where deformation accompanies recrystallisation and formation of penetrative fabrics such as foliation in phyllites, schists and gneisses. Orogenic metamorphism can last for millions of years and can include several distinct metamorphic events (e.g., Goscombe et al., 2009; Czarnota et al., 2010). Importantly, metamorphism in orogenic belts is linked to polyphase deformation and recrystallisation. Rocks subjected to orogenic metamorphism extend over large belts (100–1000-km-long and 10–100-km-wide; Bucher and Frey, 1994). According to Bucher and Frey (1994) temperature and pressure can range from 150 to 1100 °C and 2–30 kbars for crustal rocks, respectively. The temperature gradients range between 5 and 60 °C/km vertically. Pressure–temperature time 'paths' for orogenic metamorphism are highly variable, reflecting a diversity of orogenic metamorphic and related processes (e.g. Stuwe and Sandiford, 1995). These can include lithospheric thickening and contraction followed by heating (typically an overall clockwise P–T–t path), but perhaps more significantly lithospheric thinning, extension and decompression melting (complex P–T–t paths: Coney, 1980; Lister and Davis, 1989; Collins, 2002b), and heating associated with subduction, followed by thermal relaxation (Bucher and Frey, 1994).

Ocean-floor (seafloor) metamorphism includes the transformations in the oceanic crust near the mid-ocean ridges (Miyashiro et al., 1971). Ocean floor metamorphic rocks are largely mafic and ultramafic in composition and moved laterally (several 1000 km) by ocean-floor spreading. This type of metamorphism is characterised by a lack of penetrative foliation but accompanied by extensive vein formation as a result of heated seawater in convecting hydrothermal cells. A direct result is extensive fluid-rock reactions between ocean water and seafloor rocks. Thus, an overlap between seafloor metamorphism and hydrothermal alteration is noted. Temperature and lithostatic pressures range between 150 and 500 °C and <3 kbars which corresponds to a temperature gradient of 50–500 °C (vertical or horizontal; Bucher and Frey, 1994).

A subset of regional metamorphism is burial metamorphism (Coombs, 1961) for low temperature regional metamorphism affecting largely sediments and interlayered volcanic rocks in basins without influence of orogenesis or magmatic intrusions. Rocks affected by burial metamorphism largely retain the original rock fabric and thus lack schistosity. This type of metamorphism cannot be sharply distinguished from deep-seated diagenesis (cf. Angerer et al., 2015).

Contact metamorphism takes place in country rocks near the contact with intrusive or extrusive igneous bodies. Metamorphic changes are the result of heat and materials emanating from the magma, and locally accompanied by deformation connected with the emplacement of the igneous body (Bucher and Frey, 1994). The zone of contact metamorphism is termed a contact aureole with its width varying between several metres to kilometres. Rocks in contact metamorphic aureoles do not display foliation. The temperature and lithostatic pressure varies from 150 to 750 °C and from a few hundred bars to 3 kbars, respectively, which corresponds to a temperature gradient of 100 °C/km or higher (horizontal; Bucher and Frey, 1994).

Cataclastic metamorphism or dynamic metamorphism is confined to the vicinity of faults and involves purely mechanical forces causing crushing and milling of the rock fabric by high strain rates under high shear stress at relatively low temperatures (Passchier and Trouw, 2005). The resulting cataclastic rocks include fault breccia, gouge and pseudotachylite.

Impact metamorphism was coined by Dietz (1961b) in which shock waves and resulting changes in the mineralogy of the country rock are caused by the hypervelocity impact of a meteorite. Meteorite impact is thought to have been the trigger for the emplacement of the Sudbury Irruptive and its associated Ni-Cu-PGE deposits in Sudbury, Canada (Dietz, 1964; Naldrett, 1981; and references therein).

6. What is a mineral system?

It has long been known that mineral deposits with quite diverse characteristics are linked either in space or in time. For example, many intrusive-related mineral districts are zoned. A good Australian example of such a zoned system is the Zeehan field in western Tasmania (Waller, 1904; Twelvetrees and Ward, 1910; Both and Williams, 1968: Fig. 2A). This district is zoned mineralogically relative to the ~360 Ma Heemskirk Granite (Black et al., 2005), to which the deposits appear to be spatially and genetically linked. The inner zone is dominated by Sn, and changes outward to zones dominated by pyrite and then by siderite. In the outer zone, the iron gangue mineralogy changes from pyrite- to siderite-dominated. The Bingham Canyon district in Utah, western USA is also strongly zoned (Fig. 2B). The core of this system is the giant Bingham Canyon porphyry Cu-Au-Mo deposit, with an intrusive-centred metal zonation of Mo → Cu → Zn–Pb (John, 1978). These two districts illustrate that temporally- and genetically-linked deposits can be characterised by quite different metallogenic, alteration and textural characteristics.

In many intrusion-related systems, systematic zonation provides evidence that deposits with quite different metallogenic characteristics form during single mineralising events. The systematics of such systems are important, as they have the potential to provide vectors for targeting mineral deposits (e.g. Sillitoe, 1993; Corbett and Leach, 1998). Evidence for linked metallogenesis at the mineral province scale, although not as compelling, suggests that individual mineralising events can produce mineral deposits with quite different characteristics. For example, existing geochronological data suggest that unconformity-related U deposits in the Paterson Province of Western Australia overlap in time with shale-hosted Cu deposits and Zn–Pb prospects (Cross et al., 2011; Huston et al., 2010a) and may be genetically linked. Moreover, it is well established in the Eastern Goldfields orogenic gold province that gold deposits with highly variable characteristics formed from the same mineral system (Groves, 1993; McCuaig and Kerrich, 1998), and Sillitoe and Perelló (2005) have documented systematic cratonic-scale patterns in mineral deposits that have formed along the Andean margin in South America (Fig. 2C). All-in-all, several lines of evidence suggest that mineral deposits with different metallogenic characteristics can form during individual mineralising events. Understanding the relationships between mineral deposits that form coevally and cogenetically can provide not only a better understanding of mineralising processes, but, more importantly, can be predictive during mineral exploration (McCuaig and Hronsky, 2014).

Mineral deposits form through the coincidence of favourable geological processes within a given spatial setting and at a specific geological time, generally over a very tight time window. By analogy with the concept of a petroleum system (Magoon and Dow, 1994), processes that form a mineral deposit can collectively be termed a "mineral system". Wyborn et al. (1994) defined a mineral system as "all geological factors that control the generation and preservation of mineral deposits". Since this initial definition, the concept of a mineral system has evolved and a number of different formulations of this concept have been developed (e.g. Wyborn et al., 1994; Knox-Robinson and Wyborn, 1997; Hagemann and Cassidy, 2000; Jaques et al., 2002; Barnicoat, 2007; McCuaig et al., 2010; Huston et al., 2012; Angerer et al., 2014; McCuaig and Hronsky, 2014). However, all formulations, in one manner or another, include: geological (including geodynamic and architectural) setting; the driver; timing and duration of mineralisation; the source(s) of the mineralising fluid and its components; the pathways along which fluids (including magmas) flow; the depositional (or trap) site; and post-depositional modifications. A good example of the application of a mineral systems approach is that of Hitzman et al. (2005) for sediment-hosted Cu deposits.

Although the mineral system concept was developed mainly as a conceptual framework for hydrothermal mineral deposits, most components can also be applied to orthomagmatic mineral deposits and even deposits formed during erosional processes. In the context of mineral system analysis applied to (ultramafic and mafic) magmatic mineral systems, keys to the genesis of mineralisation are lithospheric architecture, which helps focus and self-organise magmatic plumbing to maximise and focus magma flux and volume (cf. Barnes and Fiorentini, 2012; Mole et al., 2014), and, most importantly in the case of komatiitic magmas, the availability of crustal sulphur reservoirs that

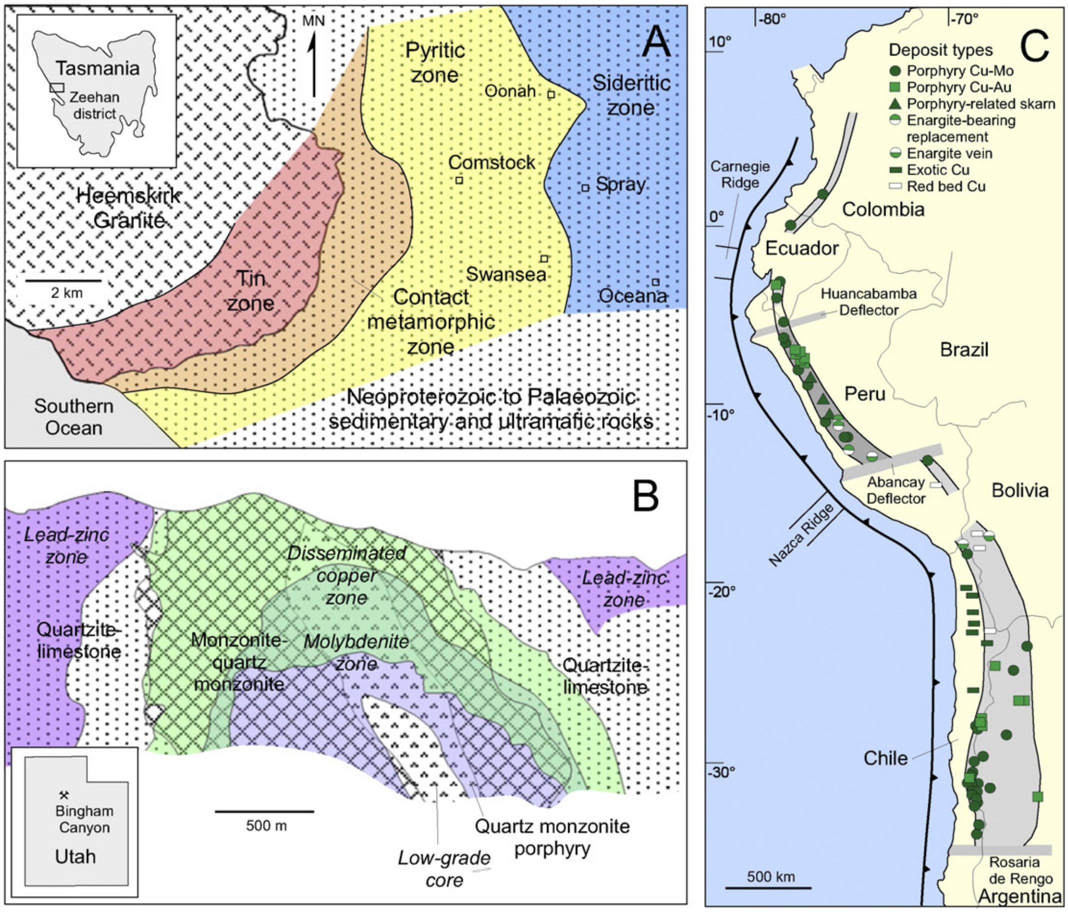

Fig. 2. Diagrams showing mineral and metal zonation in mineral fields and deposits: (A) regional zonation in Fe–S–C–O gangue minerals relative to the Heemskirk Granite in the Zeehan, Tasmania Sn–Pb–Zn district (modified after Twelvetrees and Ward, 1910), (B) cross section showing deposit-scale mineral and metal zonation at the Utah orebody, Bingham Canyon, Utah (modified after John, 1978); and (C) continental-scale zonation of mineral deposits that formed during the Miocene to early Pliocene (23–4 Ma) along the Andean margin of South America (modified after Sillitoe and Perelló, 2005).

magmas can thermo-mechanically erode and assimilate upon emplacement. In contrast to other mineral systems, it is unclear what the role of metal source is in the genesis of orthomagmatic Ni–Cu–PGE systems beyond the need for mafic and/or ultramafic magmas (cf. Hoatson et al., 2006; Zhang et al., 2008; Griffin et al., 2013; Arndt et al., 2005, 2008).

A mineral system involves the concentration of commodities (metals) through a number of different processes to the point at which the concentration is sufficient to consider exploitation (Fig. 3). When the commodities can be extracted economically, the product of a mineral system is an ore deposit (it must be born in mind that although all mineral systems form mineral deposits or occurrences, most do not form ore deposits, sensu stricto). Processes that concentrate commodities operate at scales ranging from global to microscopic and can involve: crustal differentiation; magmatic processes including melting, fractionation, crystallisation, magma mixing, immiscibility and magmatic-hydrothermal fluid evolution; hydrothermal processes including leaching and depositional processes, such as boiling, fluid mixing, cooling and fluid-rock interaction; physical processes including density separation; biological processes; and post-depositional enrichment and upgrading.

As shown in Table 1, for economic extraction, many commodities require extreme enrichment relative to bulk Earth or average continental crust. With few exceptions (e.g. Fe, Al) the formation of an ore deposit requires enrichment factors generally in excess of 100 and in some cases in excess of 1000 (Au, Pt and W) relative to continental crust, the host of virtually all ore deposits. Enrichment factors relative to bulk Earth are extremely variable, with ore grades for some commodities such as Ni, Fe and Pt lower or approximately equal to bulk Earth concentrations as these commodities are strongly concentrated in the core. Other commodities such as W, Ta, U and Pb have enrichment factors in excess of 10,000 relative to bulk Earth as the elements are highly concentrated in continental crust.

For many mineral systems, the first process of concentration was the initial chemical differentiation of the earth (orange box in lower left of Fig. 3). Based upon concentration estimates of the core, mantle and crust established by McDonough (2003); Palme and O'Neill (2004) and Rudnick and Gao (2003) (Table 1), many commodities of economic interest (e.g. Sn, W, Ta, Pb, U and REEs) are highly enriched in the crust relative to bulk earth, whereas other commodities are only weakly enriched in the crust (e.g. Zn), and others are not significantly fractionated (e.g. Ag). Some elements are fractionated into the mantle and core, either moderately (e.g. Cu) or strongly (e.g. Ni, PGE). A small number of commodities (e.g. Au and Mo) have a complex distribution, being enriched in the core and crust relative to the mantle.

Following the initial chemical differentiation of the earth, geodynamic processes (flesh-coloured box in Fig. 3) further fractionate commodities (blue arrow) within the crust and upper mantle. These processes, which operate at the cratonic- to province-scale, generally occur along active tectonic margins, and can include processes such as mantle metasomatism, the formation of a mafic underplate and many others. The combination of the initial global chemical differentiation and subsequent "tectonic" enrichment associated with magmatic and metasomatic (hydrothermal) processes, along with surficial processes (e.g. weathering and sedimentation) in some cases, produces regions

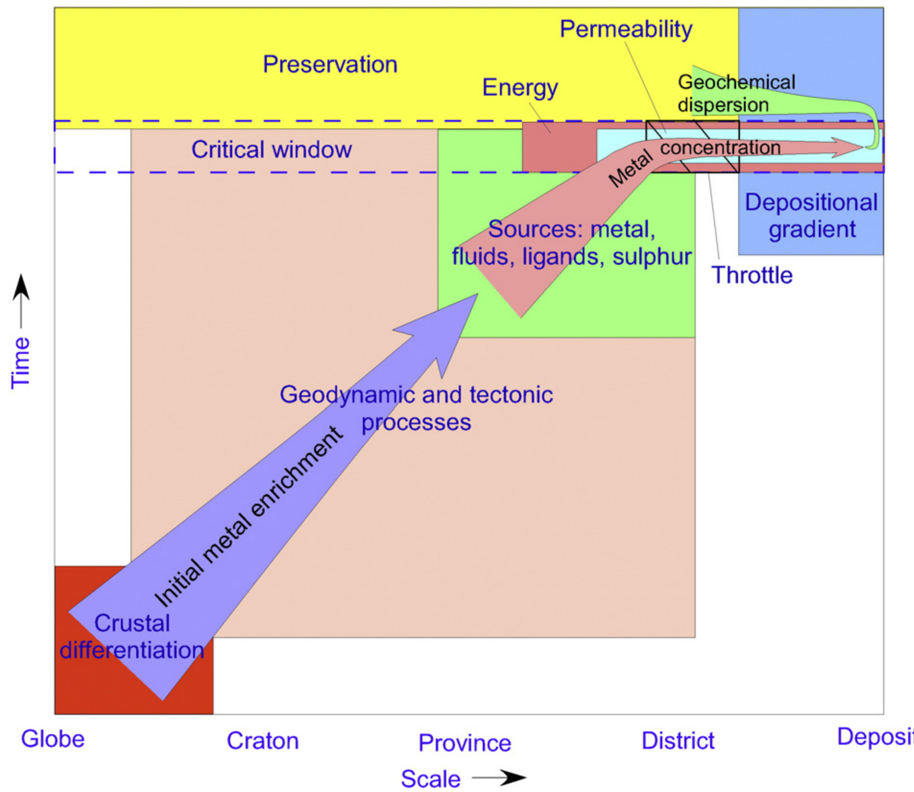

Fig. 3. Diagram showing important features and processes in mineral systems. The diagram illustrates the progressive concentration (arrows) of ore metals from the initial differentiation of Earth through ore deposit formation and post ore dispersion. It also highlights important processes and system features that enable metal concentration. It is designed to show how metal concentration occurs both in time and in space, including the importance of a critical window and/or trigger in the formation of ore deposits.

in the upper mantle and crust that act as sources of commodities (green box in Fig. 3) that are concentrated into mineral deposits by hydrothermal or magmatic processes (pink arrow). It is important to note that, in many cases, source regions form millions to hundreds of millions of years prior to mineralisation, and can also be spatially distant from the depositional site; the development of these source regions is one of many factors that influence the "fertility" of metallogenic provinces (cf., McCuaig and Hronsky, 2014).

In addition to producing geochemically fertile metallogenic provinces, tectonic processes initiate crustal- and province-scale architecture that is critical for the movement and production of later ore forming fluids. In many cases, this architecture is developed early in the tectonic history of a mineral province, in some cases tens to hundreds of millions of years prior to mineralisation, with fluid flow facilitated by reactivation of pre-existing structural architecture. In contrast, the time involved in the formation of the mineral deposit may be relatively short, in many cases on the order of 1 My or less for systems associated with magmatism (Jaques et al., 2002; Chiaradia et al., 2014).

The final concentration of commodities by hydrothermal or magmatic processes generally occurs during a critical window of time, during which hydrothermal fluids or magmas are driven along structures or other zones of permeability (e.g. aquifers formed by sedimentary, diagenetic or hydrothermal alteration processes). The presence of such fluid/magma pathways at the time of mineralisation is one of the most fundamental keys to the formation of a significant mineral system (McCuaig et al., 2010). Drivers of fluid production and flow are quite diverse, although in many cases they appear to have a tectonic trigger. These drivers can include temperature gradients induced by magmatism (which produce hydrothermal convection: Cathles, 1983), unroofing via erosion and tectonic processes (Bissig et al., 2002), stress changes (Blewett et al., 2010), magma buoyancy, the release of magmatic-hydrothermal fluids during magma crystallisation (Burnham and Ohmoto, 1980), metamorphic devolatilisation during contact or regional metamorphism (Groves, 1993), formation and expulsion of basinal brines (Leach et al., 2005), hydraulic pressure gradients caused by contractional or extensional deformation and other processes. In some cases, production of fluid is slow relative to flow (e.g. metamorphic devolatilisation), requiring a separate trigger for flow of stored fluids. In this context, seismic pumping is a key driver for vertical and lateral fluid flow along fault systems (Sibson et al., 1975).

Fluid-flow pathways form prior to, or during, fluid movement. These pathways can include extensional structures that developed as host basins or volcanic terranes formed and reactivated during subsequent tectonism, permeable zones formed during basin deposition or diagenesis, and structures formed during the event that triggered fluid flow. This trigger can be local, in which case it is generally evident in the local geological history, or distal, in which case evidence can be very subtle or cryptic (e.g. a change in apparent polar wander paths). The formation of a mineral system requires the concurrent availability of a driver, a metal (including S) source and a fluid pathway during a discrete critical window (upper part of Fig. 3). These are triggered, in many (most) cases, by specific tectonic events or switches and may be relatively short-lived.

In addition to these mineral systems requirements, formation of significant mineral and ore deposits requires two other features: a throttle (cf., McCuaig et al., 2010) that concentrates fluid flow, and a gradient, which can be either chemical or physical or both, that causes efficient deposition or segregation and trapping of commodities of economic interest. In most cases, chemical processes cause mineral deposition. For hydrothermal mineral systems, common causes of mineralisation include rapid changes in temperature, pH or redox caused by water–rock reaction or fluid mixing. In other hydrothermal systems, physical

Table 1
Variations in the abundance of selected ore elements between the core, mantle and crust, and within the crust.

Element	Solid earth	Core	Silicate earth	Mantle	Oceanic crust	Continental crust	Upper continental crust	Middle continental crust	Lower continental crust	Ore	Ore / Solid earth	Ore / Continental crust
Siderophile												
Fe (%)	32.0	85.5	6.26	6.30	7.03	5.22	3.92	4.68	6.66	35	1.1	6.7
Ni (ppm)	18,200	52,000	1960	1860	144	59	47	33.5	88	3500	0.2	59
Mo (ppm)	1.7	5	0.05	0.039		0.8	1.1	0.6	0.6	550	320	690
Au (ppm)	0.16	0.5	0.00088	0.00088		0.0013	0.0015	0.00066	0.0016	1.5	9.4	1200
Pt (ppm)	1.9	5.7	0.007	0.0066		0.0015	0.0005	0.00085	0.0027	2.0	1.1	1300
Lithophile												
Al (%)	1.59	*0.0*	2.35	2.38	8.28	8.42	8.15	7.94	8.94	50	31	5.9
Ca (%)	1.71	*0.0*	2.53	2.61	8.46	4.58	2.57	3.75	6.85			
K (%)	0.016	*0.0*	0.024	0.026	0.17	1.50	2.32	1.91	0.51			
Mg (%)	15.4	*0.0*	22.8	22.17	4.96	2.81	1.50	2.16	4.37			
Na (%)	0.18	*0.0*	0.27	0.259	1.78	2.28	2.43	2.51	1.97			
P (%)	0.0715	0.2	0.009	0.0086	0.05	0.06	0.07	0.07	0.04			
Si (%)	16.1	6.0	21	21.22	23.0	28.3	31.1	29.7	25.0			
O (%)	29.7	*0.0*	44	46.5		46.1	47.3	46.5	44.8			
Ta (ppm)	0.025	*0.0*	0.037	0.04		0.7	0.9	0.6	0.6	200	8000	290
U (ppm)	0.015	*0.004*	0.02	0.0218	0.12	1.3	2.7	1.3	0.2	300	20,000	230
Th (ppm)	0.055		0.08	0.0834		5.6	10.5	6.5	1.2			
Th/U	3.7		4.0	3.8		4.3	3.9	5.0	6.0			
Ba (ppm)	4.5		6.6	6.75	49	456	628	532	259			
W (ppm)	0.17	0.47	0.029	0.016		1.0	1.9	0.6	0.6	2700	16,000	2700
Cl (ppm)	76	200	17	60		244	294	182	250			
La (ppm)	0.44		0.65	0.686	5.5	20	31	24	8			
Lu (ppm)	0.046		0.068	0.0711	0.4	0.3	0.31	0.4	0.25			
Cr (ppm)	4700	9000	2625	2520	320	135	92	76	215	230,000	49	1700
Chalcophile												
Ag (ppm)	0.05	0.15	0.008	0.004		0.056	0.053	0.048	0.065	31	620	550
Cu (ppm)	60	125	30	20		27	28	26	26	3100	52	110
Pb (ppm)	0.23	0.4	0.15	0.185		11	17	15.2	4	8700	38,000	790
S (ppm)	6350	19,000	250	200	900	404	621	249	345			
Sn (ppm)	0.25	0.5	0.13	0.138		1.7	2.1	1.3	1.7	1300	5200	760
Zn (ppm)	40	*8.6*	55	53.5	78	72	67	69.5	78	25,000	630	350
Reference	1	1	1	2	3	4	4	4	4	5		

References and notes.
1. McDonough (2003); values in italics estimated from mass balance calculations.
2. Palme and O'Neill (2004).
3. Wedepohl and Hartmann (1994).
4. Rudnick and Gao (2003).
5. Based upon grades of relatively low grade deposits.

changes, such as depressurisation, can induce chemical and physical reactions (e.g. boiling and volatile loss) that cause mineralisation. In some orthomagmatic systems, chemical changes such as magma mixing or contamination by wall rocks can produce immiscible melts or cause crystallisation of minerals containing the commodities of interest. In these cases, final enrichment to form mineral deposits results from density-driven physical processes, and physical traps (e.g. constrictions) that control the location of mineralisation (Barnes et al., 2016-in this issue). These factors are also important in mineral systems formed during weathering and erosion.

In many hydrothermal mineral systems, the mineralising process produces zones of anomalous mineralogy, geochemistry and physical rock properties that extend well beyond the economic limits of ore (i.e. the footprint). These zones not only involve enrichment of the commodities of interest, but also enrichment and/or depletion of related elements. In most cases, the mineralising processes change the abundance or composition of minerals present. These zones can extend vertically above and below or laterally away from mineral deposits, and are integral parts of mineral systems. In addition, the chemical reactions that produce these zones can significantly change the physical properties of the affected rocks. This most commonly involves changes to the magnetic or electrical properties of the affected rock, although it can also change density and rock rheology. Hypogene geochemical dispersion and physical property changes are useful in exploration as they increase the footprint of individual mineral deposits or ore bodies (Hansen et al., 1966; Rose et al., 1979).

The last aspects of a mineral system, which are less commonly considered, are post-depositional processes, which can have major effects on preservation and the economic viability of mineral deposits. Certain types of mineral systems, such as porphyry Cu or epithermal Au–Ag deposits, which form in the linked porphyry-epithermal mineral system, have a lower probability of preservation, as they form in topographic elevated geological environments that are easily eroded (Groves et al., 2005). These deposits are mostly Cenozoic in age, with older examples, although known, much less abundant (Seedorf et al., 2005). Similarly, erosion of the crater facies of diamondiferous kimberlite pipes to leave small feeder dykes can render potentially economic deposits too small for economic extraction. Black smokers forming along mid-ocean ridges are also unlikely to survive, as these deposits are oxidised and as the oceanic crust that hosts these deposits is almost inevitably subducted and lost (Hannington et al., 2005).

In other cases, the geodynamic setting of mineralisation acts to preserve mineral deposits. For example, deposits formed in graben systems can subsequently be shielded from erosion (e.g. the Paraburdoo iron ore deposits: Hagemann et al., 2016-in this issue). In other systems, the deposition of post-depositional volcanic or sedimentary rocks can also enhance preservation. Post-depositional faults can down-drop deposits or zones, shielding them from erosion, as at the Kalamazoo porphyry Cu and the United Verde Extension deposits in Arizona (Lowell and Guilbert, 1970; Anderson, 1968). Understanding post ore movement of faults is critical for near deposit exploration, as demonstrated by these latter two examples. At the broad scale, the main driver of preservation is isostasy: topographic anomalies produced by orogenesis or arc formation, and mineral deposits within these anomalies will be removed through the interplay between by erosion and isostatic rebound, until peneplanation of the Earth's surface to near sea-level is approached.

An important indicator of the likely preservation, in some cases, is regional metamorphic grade, which reflects the depth of burial of a metallogenic province and, therefore, the amount of the rock column that has been removed by erosion. As high level deposits, such as porphyry Cu or epithermal deposits, are unlikely to be buried and are highly

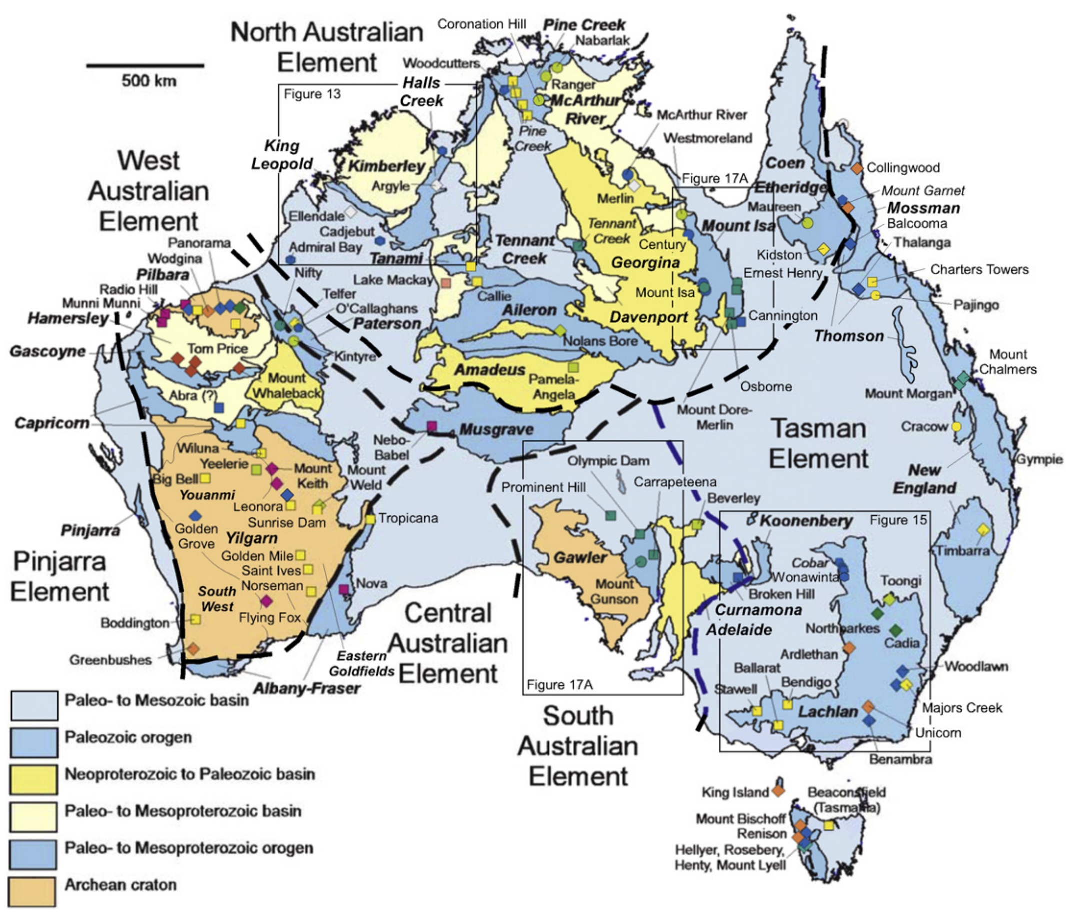

Fig. 4. Tectonic elements of Australia, showing the locations of major mineral deposits (updated from Huston et al., 2012).

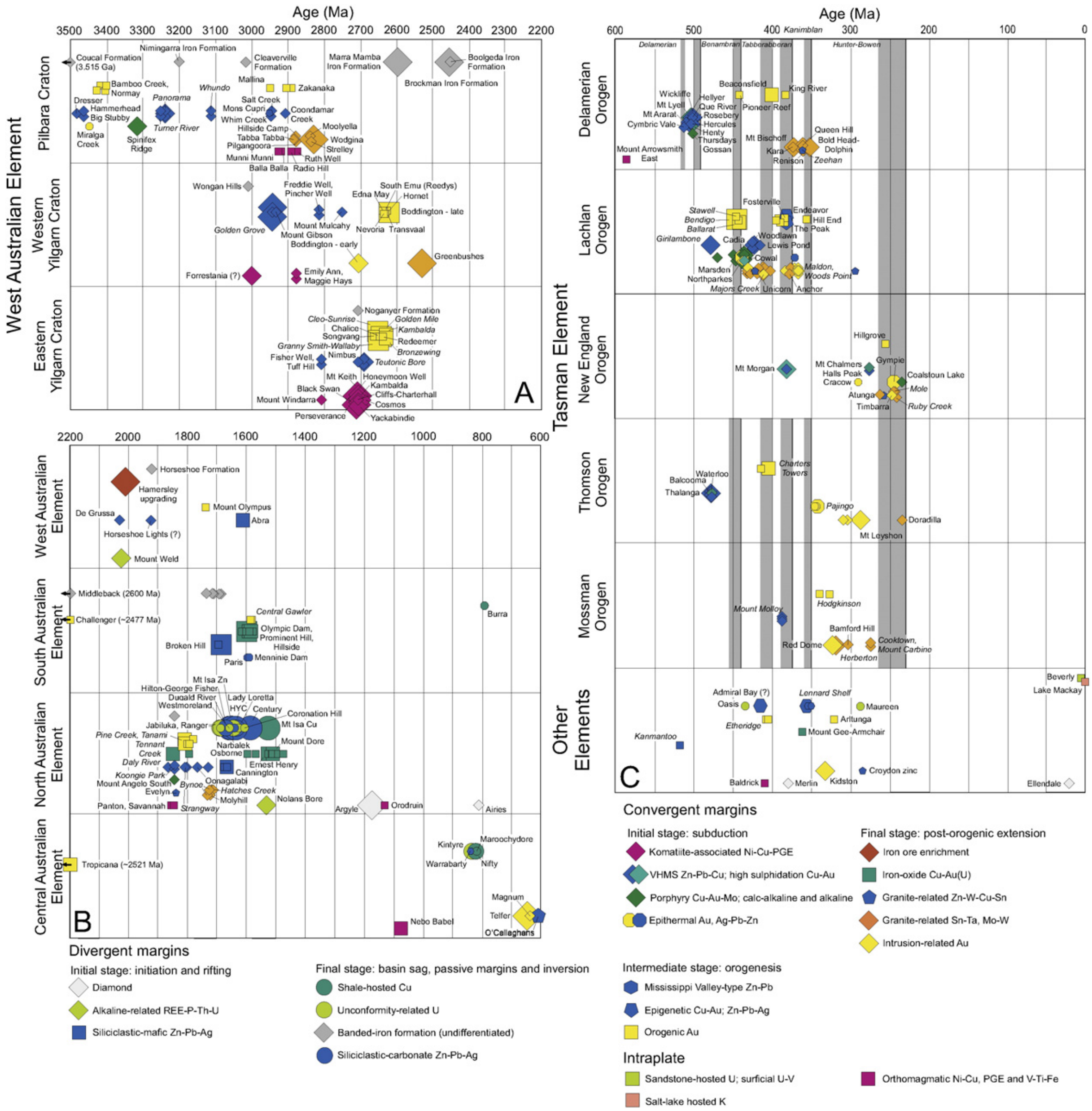

Fig. 5. Distribution of Australian mineral deposits through time: (A) 3500–2200 Ma, (B) 2200–600 Ma, and (C) 600–0 Ma (updated from Huston et al., 2012). In (C) the temporal boundaries of tectonic cycles are indicated by heavy lines and deformation events are indicated by grey fields. The mineral deposit age dataset (with references) is available as a supplementary item.

susceptible to erosion, except in highly unusual situations, these deposits will be restricted to very low-metamorphic-grade provinces. In contrast, other deposit types, such as orogenic gold deposits, tend to form at somewhat deeper levels in the crust and are preferentially concentrated in low- to medium-grade-metamorphic terranes (Groves et al., 2005). In contrast, other deposit types, such as some that form in basins (e.g. volcanic-hosted massive sulphide or sediment-hosted massive sulphide deposits) can be deeply buried and are preserved in even in high-metamorphic-grade terranes. However, it must be stressed that if mineralisation post-dated metamorphism, regional metamorphic grade does not indicate the amount of erosion post mineralisation. A major tectonic control on preservation is lithospheric architecture: thick, stable lithosphere is more likely to preserve deposits (Groves et al., 2005). Moreover, deposits that form late in the tectonic cycle are more likely to be preserved than deposits that form early in the tectonic cycle (McCuaig and Hronsky, 2014).

Post-depositional processes, such as metamorphism and supergene processes, can enhance grade and recoveries, turning mineral deposits into ore deposits. Either metamorphism coarsens ore minerals to allow greater recovery of commodities, or supergene processes enrich certain commodities (e.g. Cu, Pb, Fe, Mn, Ag, Au and Ni) during weathering. In some cases, refractive hypogene ore can be weathered to produce oxidised ore amenable to processing. Weathering processes can also enhance geochemical dispersion, increasing the detectable footprint of mineral deposits (Rose et al., 1979).

7. Linking mineral systems to tectonic and geodynamic processes: tectono-metallogenic systems

There is reasonable evidence to suggest that metallogenesis changes as tectonic processes evolve. Discrete age 'pulses' of mineralisation preserved in the regional rock record may relate to episodes in terrane evolution where rapid geodynamic changes have occurred — tectonic mode switches (Collins, 2002b; Lister and Forster, 2009). Collins and Richards (2008), in attempting to explain the timing and occurrence of S-type magmatism in the Lachlan Orogen, proposed that the Phanerozoic Tasman Element in eastern Australia (Fig. 4) grew along a convergent margin during a series of tectonic episodes (or cycles) related to episodes of slab rollback following contraction. These cycles involved two to three stages: (1) arc formation and crustal extension (back-arc basin) associated with roll-back of the plate being subducted, (2) crustal shortening associated with arc/terrane accretion and/or flattening of the subducting slab, and (3) post-collisional extension and magmatism. If an arc-back-arc system forms in response to post-collisional extension, a second tectonic cycle is initiated. Compilations of mineral deposit age data suggest that the formation of different types of mineral deposits is linked to the evolution of tectonic cycles within the Tasman Element, which, in turn, reflects changing geodynamic environments (Fig. 5C: Champion et al., 2009; Huston et al., 2012). For example, volcanic-hosted massive sulphide and many porphyry Cu deposits form early during tectonic cycles, whereas orogenic gold and certain granite-related deposits form late during individual cycles (Section 9). In addition to this temporal differentiation of deposits, the location of deposit types that form at or about the same time can vary in space. For example, during the early state of convergent cycles porphyry Cu deposits form along arcs (Sillitoe, 1972), whereas VHMS deposits form inboard in back-arc basins (Hannington et al., 2005).

Although not as clearly established as in the Tasman Orogen, the metallogeny of other mineral provinces in Australia also reflects changing geodynamic environments that may mirror changes in the tectonic cycle(s) (Fig. 5). For example, in the Eastern Goldfields Superterrane in the Yilgarn Province, komatiitic Ni–Cu–PGE and volcanic-hosted massive sulphide Cu-Zn deposits form early (Barnes et al., 2016-in this issue; Huston et al., 2012; Fig. 5A), whereas orogenic gold deposits form late (Groves et al., 1998; Hagemann and Cassidy, 2000; Czarnota et al., 2010; Fig. 5A; Section 9.3.1). This relationship is similar to that in the Tasman Element and can also be seen in individual deposits. For example, at the Mount Gibson deposit in the Murchison Domain, Youanmi Terrane, VHMS mineralisation has been overprinted by a later orogenic gold system (Yeats and Groves, 1998). Moreover, in the Saddleback Greenstone belt in the Western Gneiss Terrane, original low-grade porphyry Cu-Au mineralisation at Boddington was overprinted by an intrusion-related gold system to form one of Australia's largest gold deposits (McCuaig et al., 2001; Hagemann et al., 2007).

Although the linkage of certain mineral systems to tectonic or geodynamic environments has long been postulated (e.g. Sillitoe, 1972; Hutchinson, 1973; Meyer, 1981, 1988; Sawkins, 1984; Barley and Groves, 1992; Kerrich et al., 2005), changes of metallogenesis during the evolution of tectonic cycles has not been as extensively explored. In the following discussion we propose a classification that links mineral systems not only to tectonic settings, but to tectonic evolution. We integrate mineral systems in Australia and elsewhere with tectonic cycles, including those along plate margins and those within plates, to define “tectono-metallogenic systems”. The concept of tectono-metallogenic systems not only provides a classification system, but it can also be used to predict the likely location of mineral systems based on geodynamic or tectonic data or the presence of other tectono-metallogenically-related mineral systems. Moreover, the concept can provide additional constraints on ancient tectonic environments based upon mineral deposit/system assemblages.

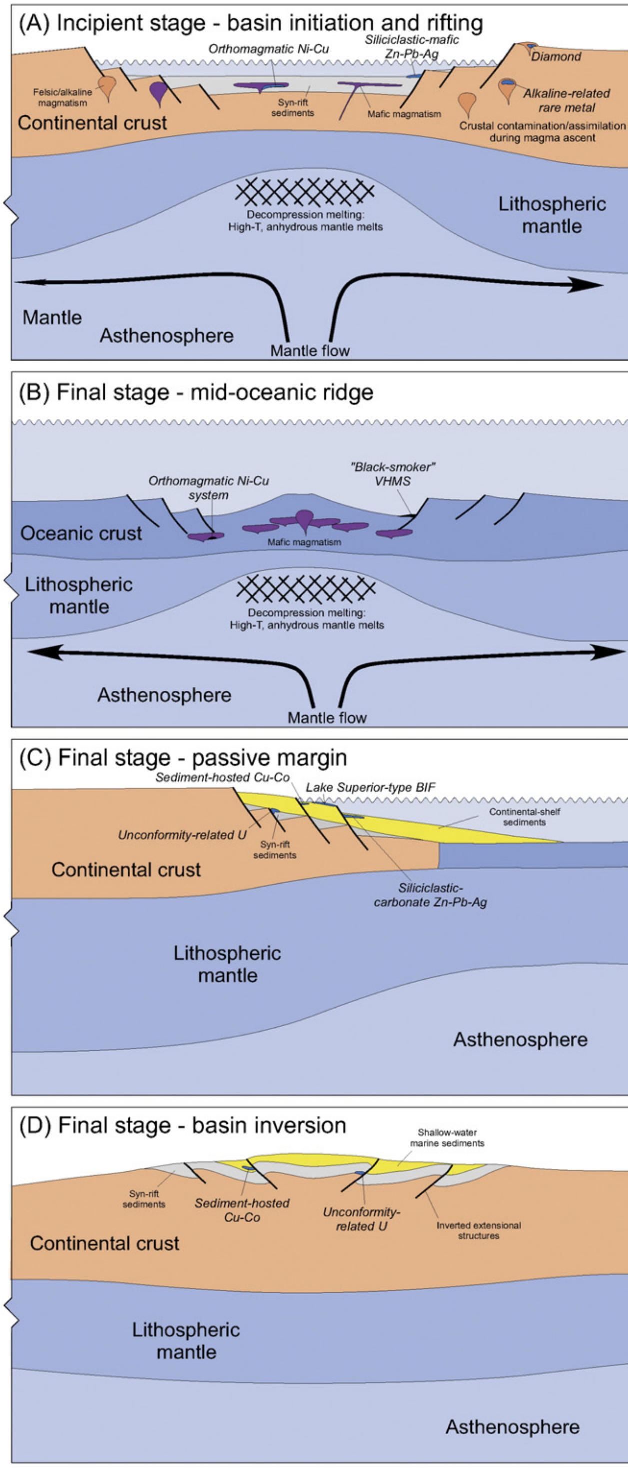

Fig. 6. Schematic diagrams showing stages in the evolution of a divergent margin and associated mineral deposits: (A) incipient stage — basin initiation and rifting, (B) final stage — mid-oceanic ridge, (C) final stage — passive margin, and (D) final stage — basin inversion. This diagram is simplified, with complexities discussed in the text.

8. The tectono-metallogenic system related to divergent margins

Divergent plate margins can form in oceanic and/or continental settings (Fig. 6A). The most extensive tectonic systems on modern Earth are the mid-ocean ridges (Fig. 6B). These divergent oceanic systems, which include the Mid-Atlantic Ridge and the East Pacific Rise, among others, circle the globe and are the main regions where new oceanic crust is added. This new crust is rarely preserved, however, as it is generally lost through subduction at convergent margins (except where anomalously thick, such as intersection of ocean ridges and plumes). The oldest preserved seafloor crust has an age of only 180 Ma (Müller et al., 2008), and ophiolites (which may only represent crust formed in back-arc basins and not ocean basin crust) are uncommon prior to Rodinia-time and unknown prior to 1900 Ma (Dann, 1991; Moores, 2002; Peltonen et al., 2003). Because oceanic crust is preserved very rarely, if at all, prior to 1000 Ma, the following discussion concentrates on divergent margins in continental settings that are more commonly preserved in the geological record. These include both the remnants of successful rifting – passive margins (Fig. 6C) – as well as failed rifts.

The best modern example of an incipient divergent margin system straddles eastern Africa and Arabia and involves the East African Rift, the Red Sea and the Gulf of Aden, forming a triple junction, where three plates are diverging from each other at rates of up to 2.7 cm/yr (McClusky et al., 2010). Although incipient divergent systems are not as well mineralised as mid-ocean ridges, they can contain active and extinct carbonatite volcanos (e.g. Ol Doinyo Lengai in northern Tanzania: Hay, 1989), and Zn deposits (e.g. the Atlantis II Deeps in the Red Sea: Hackett and Bischoff, 1973). Although the Ol Doinyo Lengai system is not known to contain significant mineral deposits, older carbonatites and related alkaline magmatic rocks are major sources of REEs and Cu (Section 8.1.2). Moreover, the Atlantis II Deeps was the site of the first discovery, in the 1960s, of active mineralisation on the seafloor. This deposit is estimated to contain a resource of 94 Mt grading 0.5% Cu, 2.0% Zn and 39 g/t Ag and 0.5 g/t Au (Herzig et al., 2002).

In non-convergent tectonic systems, there are several possible triggers for intracratonic rifting and the initiation of divergence. One mechanism is thermal insulation of the mantle by large continental landmasses, triggering mantle upwelling, doming of the Earth's surface and rifting (Anderson, 1982). A second trigger is the ascent of mantle plumes from the base of the mantle, again doming the Earth's surface, triggering rifting and, in some cases continent break-up (e.g. White and McKenzie, 1989; Heaman et al., 1992). Both of these triggers involve thermal anomalies within the convecting mantle and the development of tensional stresses within plates, and the former mechanism could be one trigger for supercontinent break-up. Storey (1995) suggested that mantle plumes may trigger the break-up of smaller blocks, whereas other mechanisms are required for the break-up of large continental land masses.

Here we do not consider the recently described setting of "hyper-extended margins" (Whitmarsh et al., 2001), in which continental break-up occurs over a cold mantle. Although this setting is known to contain petroleum plays (Unternehr et al., 2010), owing to the relatively recent demonstration of this tectonic setting, its metallogenesis is not yet understood.

Many incipient divergent margins fail, and do not proceed to form oceanic basins. In most cases one of the arms of triple junction rift systems fail, forming an aulacogen (Burke and Dewey, 1973; Dewey and Burke, 1973), which can be preserved in the geological record. Changes in plate motion caused, for example, by interaction with other plates, can also cause intracratonic rifts to fail. Examples of failed rifts around the world include Gulf of Suez (Steckler et al., 1988), the ~1100 Ma Mid-Continent Rift (Dickas, 1986), the Cenozoic Rio Grande Rift (Chapin, 1979), the Rhine Grabben (Ziegler and Dèzes, 2005), the Baikal Rift (Hutchinson et al., 1998), and the West Antarctic Rift (Behrendt, 1999). An Australian example in is the Paleozoic Canning Basin (Kennard et al., 1994), which marks the western end of the Larapinta Seaway, a Paleozoic seaway that extended across northern Australia to the Tasman Element (Bradshaw, 1993; Haines and Wingate, 2007: Fig. 4).

In the following discussion, the evolution of divergent margins is discussed in two stages, the incipient stage (Fig. 6A), which involved the initiation of rifting and basin development, and the final stage, which involved the formation of mid-ocean ridges (Fig. 6B), passive margins (Fig. 6C) and the thermal subsidence phase (Fig. 1A) of basin development. Rift basins can also form in convergent systems, where slab roll-back places the over-riding plate into extension, forming back-arc basins (Section 9).

8.1. Incipient stage — basin initiation and rifting

Initiation of divergent margins within continental land masses produces rift basins generally characterised initially by continental and lacustrine sediments, followed in many cases by marine sediments (Fig. 6A). The basin fill is dominated by siliciclastic sediments and, because thinning of the lithosphere causes decompression melting in the upper mantle, these basins also can contain extensive tholeiitic mafic or bimodal magmatic rocks — lavas, volcaniclastic rocks and shallow intrusive bodies (White and McKenzie, 1989). The development of carbonate-dominated sedimentary facies is limited, generally, to rift margins and shoulders (Bosence, 2005). Importantly, the incipient stages of divergence can also produce extensive evaporitic facies (Bosence, 2005) that can be tapped to produce saline basinal or hydrothermal fluids. Moreover, high heat flow associated with magmatism can drive moderate to high temperature hydrothermal fluid flow, and the magmatic events themselves can emplace Ni–Cu- and/or PGE-enriched intrusions (e.g. the Duluth complex in the Mid-Continent Rift: Dickas, 1986) or small carbonatitic and related alkaline bodies (Burke et al., 2003) into the evolving rift succession. Intrusions within rifts can be zoned in time and space, with early-emplaced alkaline bodies, which form from low-degree partial melts of the mantle, located on rift flanks, and more voluminous tholeiitic bodies, which crystallise from high-degree partial melts of the mantle, occupying the central parts of the rift as it evolves.

8.1.1. Siliciclastic–mafic Zn–Pb–Ag–Au–Cu deposits

Base metal deposits hosted by siliciclastic-dominated successions (i.e. the shale-hosted deposits of Gustafson and Williams, 1981, or the clastic-dominated deposits of Leach et al., 2005, 2010) are the largest global source of Zn and Pb, and significant sources of Ag and, in some cases, Au. This broad group of deposits was divided into flysch-hosted and epicratonic/platform-hosted types by Morganti (1981). Most of the flysch-hosted deposits of Morganti (1981) are also associated with mafic volcanism. Australian examples of this deposit type include Broken Hill and Cannington. Global examples include Sullivan in Canada (Lydon, 2007), and the active Atlantis II deeps deposit (Hackett and Bischoff, 1973).

The lithological make-up of the initiating rift basin (Fig. 6A) has a significant influence on the characteristics of siliciclastic–mafic Zn–Pb deposits (Huston et al., 2006). In many districts around the world (e.g. Broken Hill, Australia (Willis et al., 1983); Sullivan, Canada (Lydon, 2007)), the stratigraphic position that hosts the orebody also corresponds to the position at which point mafic magmatism ceases. As most workers (Gustafson and Williams, 1981; Leach et al., 2005, 2010) consider these deposits to have a syngenetic or diagenetic timing with respect to the host succession, this suggests a close genetic link between this mafic magmatism and ore formation: magmatism provided heat and/or metals, either directly (Crawford and Maas, 2009) or by leaching (Huston et al., 2006), to the evolving mineral system. The high heat flow and geothermal gradients associated with rifting caused the convection of relatively high temperature fluids which were reduced as a consequence of reaction with reduced, Fe^{2+}-rich basin fill (e.g. mafic volcanic rocks and immature turbiditic rocks). These fluids not only transport Zn, Pb and Ag, but can also transport Cu and Au (e.g. Cooke et al., 2000). Sediment-hosted Zn–Pb deposits in siliciclastic–mafic host successions commonly contain low but economically recoverable concentrations of Au. The Broken Hill deposit had an average of grade of 0.47 g/t (Woodall, 1990) and produced the most gold in New South Wales until the Au-rich porphyry Cu deposits in the Macquarie arc (Section 9.4.1) were discovered and began production.

Another characteristic of many continental rifts around the world is the presence of evaporites towards the base of the basin fill (Warren, 2010; Fig. 6A). Interaction of circulating hydrothermal fluids with these evaporites would increase the salinity of the fluids, with two important consequences. First, higher salinity fluids can transport higher concentrations of base metals, making them more potent ore fluids. Second, a significant increase in salinity also increases ore fluid density, making the formation of brine pools possible when the fluids reach the seafloor (Sato, 1972). This behaviour, which is observed in the modern Atlantis II Deep deposit (Anschutz et al., 1998), can be very effective in trapping metals.

As discussed above, many of the features of Zn–Pb deposits hosted by siliciclastic–mafic successions are the direct product of the basins in which they form. Active mafic magmatism, combined with the presence of evaporites at depth in the succession, produces saline and reduced, relatively high-temperature ore fluids capable of transporting, Zn, Pb, Ag and Au (Huston et al., 2006). Structures and aquifers formed during basin evolution focus the ore-forming fluids, allowing metal sourcing and determining the locus of ore deposition.

8.1.1.1. Australian examples. Important Zn–Pb deposits hosted by such basins in Australia include Paleoproterozoic (1710–1670 Ma) deposits at Cannington, Pegmont and near Einasleigh in Queensland, and at Broken Hill (Solomon and Groves, 2000; Huston et al., 2006). Cambrian (~520 Ma) deposits of this type are present in the Kanmantoo rift in South Australia (Belperio et al., 1997).

8.1.2. Deposits associated with carbonatites and alkaline (ultra-potassic) igneous rocks

Most carbonatites and all kimberlites are of intraplate origin and located on Precambrian cratons (Mitchell, 1986; Woolley and Bailey, 2012). The majority of carbonatites, especially young carbonatites, are associated with crustal doming and/or extension in rift systems or near-rift settings (Woolley, 1989; Tappe et al., 2006; Fig. 6A) but some are associated with extension in orogenic or collisional settings (Woolley and Kjarsgaard, 2008). Similarly, many, though not all, kimberlites and related alkaline rocks are also associated with rifting, including that which broke up Gondwana (Jelsma et al., 2009). Others, such as those in North America, may be the result of partial melting of continental lithosphere induced by mantle plume activity producing hotspots (e.g. Heaman and Kjarsgaard, 2000) or by fluid subduction (e.g. Currie and Beaumont, 2013).

Large scale lithospheric structures play an important role in controlling the location of both kimberlites and carbonatites, at the regional scale, within cratons. Repeated episodes of magmatism focussed at craton margins and intracratonic domain boundaries typically results in refertilisation of the sub-continental lithospheric mantle, and these weakened boundaries are the sites of repeated cycles of extension, rifting, subduction and accretion (Begg et al., 2009).

Australian kimberlites, lamproites and ultramafic lamprophyres, and carbonatites are typically located at major lithospheric structures that can be identified in upward continued potential field data at continental, regional and local scales (Jaques and Milligan, 2004). Some of these structures have been active over extended periods of time with multiple intrusion events. Some alkaline suites may be related to extensional events such as the breakup of Rodinia, whereas others exploited major structures at the margin of cratons, and, in some cases, exploited older structures associated with earlier tectonism and magmatism (Jaques and Milligan, 2004).

The geodynamic settings of most Australian carbonatites, kimberlites, lamproites and associated alkaline rocks is uncertain but likely involves far-field stresses associated with the re-arrangement of tectonic plates. The trigger for eruption of kimberlites and related alkaline rocks appears to have been thermal perturbations and/or stress changes at the base of the thick sub-continental lithosphere induced by the break-up (and possibly assembly) of supercontinents that generated small volume partial melts of the mantle (Jelsma et al., 2009). Modelling suggests that the shape of the cratonic keel, the proximity to the craton boundary and stress gradients at the lithosphere-asthenosphere boundary are all important in determining the location and extent of partial melting with the (kimberlite) melt migrating to trans-lithospheric discontinuities (O'Neill et al., 2005). Isotopic and trace element data indicate that kimberlites, lamproites and lamprophyres represent small degrees of partial melting of asthenospheric mantle or, in the case of lamproites which have more radiogenic Sr, unradiogenic Nd and more enriched trace element signatures, of sub-continental lithospheric mantle that has undergone long-term geochemical enrichment (Smith, 1983; Fraser et al., 1985; Jaques et al., 1989; Mitchell, 1995). Trace element and isotopic data suggest that many carbonatites may also represent small volume melts from either asthenospheric mantle and/or refertilised sub-continental lithospheric mantle (e.g. Nelson et al., 1988; Graham et al., 2004; Tappe et al., 2006).

Alkaline rocks, especially kimberlites, lamproites and lamprophyres, and carbonatites typically have high abundances of Sr, Zr, Nb, Ba, LREE, Ta, Th and U, and commonly other 'incompatible elements' as a result of their fractionation during partial melting of their mantle sources. They host a range of deposits, including those of diamond, rare-earth element (REE) and other rare metals (e.g. Nb, Ta, Zr), which are of increasing importance as sources of high technology metals.

8.1.2.1. Australian examples. The world-class Argyle diamond deposit, hosted in the Argyle AK1 lamproite (Jaques et al., 1986), and smaller deposits at Ellendale (Ahmat, 2012) are hosted in olivine lamproite and lie at the margins of the Kimberley Craton. Several Australian carbonatites have economic or potentially economic concentrations of REE, U, P, Nb, Ta and other 'rare metals' (e.g., Mount Weld, Cummings Range). High-grade hydrothermal mineralisation can be present locally (Downes et al., 2014), but the mineralisation is typically hosted in regolith developed primarily by supergene enrichment (Hoatson et al., 2011). Weathering with pronounced vertical and lateral groundwater flow resulted in leaching and dissolution with enrichment in resistate minerals and formation of secondary REE-rich phosphates and aluminophosphates under conditions of high fluid/rock ratios, long fluid residence tomes, and a range of pH conditions (Lottermoser, 1990). Some felsic alkaline suites are strongly enriched in Y, Zr, Nb and LREE by extended crystal fractionation and other processes, leading to potentially economic concentrations (e.g. the Toongi deposit, and the large, low-grade Hastings-Brockman deposit: Chalmers, 1990; Ramsden et al., 1994; Skirrow et al., 2013; Alkane. Resources, 2014; Jaireth et al., 2014).

8.1.3. Other deposit types

As discussed above, because similar geodynamic ore-controlling processes can occur in several tectonic settings, some types of ore deposits occur in different tectonic settings. As an example, Zn–Pb deposits hosted by siliciclastic–mafic successions also occur in back-arc basins, for example, the Selwyn Basin in Canada (Goodfellow, 2007a). It is possible that basin initiation could also form other types of deposits that are more typical of other tectonic environments, including unconformity-related U and shale-hosted Cu, which are more prevalent as basins evolve (see next section), or orthomagmatic Ni–Cu–PGE, which commonly form during large mafic magmatic events associated with mantle plumes (Pirajno and Hoatson, 2012), and which can initiate divergence (Section 10). Finally, mid-ocean ridges (Fig. 6B) are the sites where black smoker deposits, the modern analogues of volcanic-hosted massive sulphide (VHMS) deposits, were discovered. Due to the low probability of preservation of oceanic crust, however, most known ancient VHMS deposits are associated with convergent margins (Franklin et al., 2005).

8.2. Final stage — basin thermal subsidence, formation of passive margins and basin inversion

After the initial rift phase, cooling of the underlying lithospheric mantle leads to continued subsidence, the thermal subsidence phase of basin evolution (Fig. 1A). At this time the sedimentary fill generally becomes dominated by shallow water marine sediments (Fig. 6D). If extension continued to the point of oceanic crust production, passive margins formed (Fig. 6C), which are associated with wide ranging paleo-water depths. In and of themselves, thermal subsidence basins and passive margins are generally atectonic and lack active magmatism.

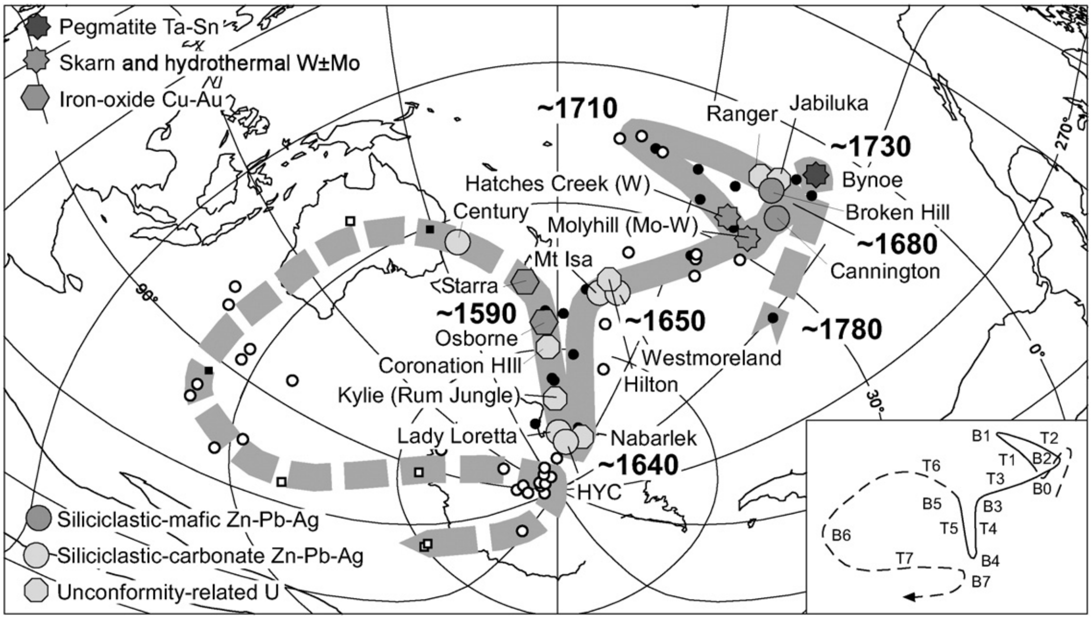

Fig. 7. Timing of mineralisation relative to the late Paleoproterozoic apparent polar wander path for the North Australian Element (modified after Idnurm, 2000). Symbols in the inset number translations (T1–T7) and bends (B0–B6).

Out-of-area tectonic events, however, which are commonly subtle and have very weak local expression, can trigger mineralisation. This is best illustrated in northeast Australia, where distal tectonic events, as recorded by changes in the apparent polar wander path (Idnurm, 2000: Fig. 7), appear to have triggered Zn–Pb–Ag and, possibly, U mineralising events (Section 8.2.1). Alternatively, the sedimentation may be terminated by tectonic events and the basin inverted (Fig. 6D). Mineral systems may be present in many phases of basin development and termination, but some classes appear to develop better in different settings.

8.2.1. Siliciclastic–carbonate Zn–Pb–Ag deposits

A number of large Zn–Pb–Ag deposits in Australia, North America and elsewhere are hosted by epicratonic and platformal basins (Plumb and Derrick, 1975; Morganti, 1981). These basins are characterised generally by relatively shallow-water sedimentary successions that contain siliciclastic rocks along with large amounts of carbonate-bearing units, but lack coeval mafic magmatism; hence the term siliciclastic–carbonate Zn–Pb–Ag deposits. In addition to deposits in the North Australian Zinc Belt (HYC, Mount Isa, Hilton-George Fisher, Century and Dugald River: see blow), other deposits of this type around the world include Howards Pass in Canada (Goodfellow, 2007a), Red Dog in Alaska (Kelley et al., 2004) and deposits in the Gamsberg district in South Africa (Rozendaal, 1986; Ryan et al., 1986).

As with Zn–Pb deposits hosted by siliciclastic–mafic successions, the make-up of the host basin influences the characteristics of siliciclastic–carbonate deposits. These basins lack mafic magmatism at the time of mineralisation, and are dominated by rock packages that are either oxidised or redox neutral, such as carbonates and redbeds (Plumb and Derrick, 1975; Morganti, 1981; Southgate et al., 2000; Gibson et al., 2012). Because the basin lacks reductants and magmatic heat sources, the fluids are relatively low temperature (<200 °C) and oxidised (Cooke et al., 2000; Huston et al., 2006).

It is likely that the metal sources for these deposits are felsic and mafic volcanic rocks underlying the host basin. Cooke et al. (1998) demonstrated that intermediate to mafic rocks from the Settlement Creek Volcanics in the basin underlying the host basin to the HYC deposit in the Northern Territory have lost over 90% of Zn and Pb during K-feldspar-hematite alteration. Moreover, paleomagnetic data reported by Cooke et al. (1998) indicate that this alteration event was coeval with the mineralising event at the HYC siliciclastic–carbonate Zn–Pb deposit.

The ore deposits are generally hosted by sulphide-rich black shale/siltstone, a rock type that is fairly restricted in host basins (Southgate et al., 2000). Environments represented by black shale were a source of H_2S from biogenic reduction of sulphate (Machel, 2001) required for base metal sulphide deposition.

8.2.1.1. Australian examples. In Australia, these deposits constitute the North Australian Zinc Belt (or Carpentaria Zinc Belt of Sweet et al., 1993), which stretches from Dugald River in the southeast to HYC in the northwest and includes many of the largest Zn–Pb deposits in the world – Mount Isa, Hilton-George Fisher, Century and HYC – as well as a number of smaller, but still significant deposits such as Dugald River, Lady Loretta and the recently discovered Myrtle and Teena deposits. An unusual characteristic of the North Australian Zinc Belt is the low abundance of small Zn prospects and deposits (L Wyborn, pers. comm.) that characterise many of the districts dominated by siliciclastic–mafic, volcanic-hosted or Mississippi Valley-type Zn–Pb deposits. This suggests that fluid flow in siliciclastic–carbonate systems lasted longer, was more tightly throttled, and/or that depositional processes were more efficient in precipitating and retaining metals at the site of deposition. The two depositional environments proposed for these deposits, early diagenetic mineralisation (e.g. Williams, 1978) or syngenetic precipitation within a brine pool (Large et al., 1998), would both be very effective at precipitating and retaining Zn–Pb sulphide minerals.

An interesting feature of the North Australian Zinc Belt is the temporal association of major deposits with bends in the apparent polar wander path, as noted initially by Loutit et al. (1994) and documented in detail by (Idnurm, 2000: Fig. 7). This relationship, which also applies to U deposits, suggests that these deposits formed in response to far-field tectonic events or global plate reorganisations. The age of the HYC deposit (~1640 Ma: Sun et al., 1994, 1996) corresponds to the age of the Liebig Orogeny (1640–1635 Ma: Scrimgeour et al., 2005) some 1800 km to the south. Geological relationships, seismic data and paleomagnetic data from this deposit (Rawlings et al., 2004; Ireland et al., 2004; Symons, 2007) suggest that the depositional site was tectonically active at the time of mineralisation. Based upon a seismic traverse crossing the ore-related Emu Fault 40 km to the north of the HYC

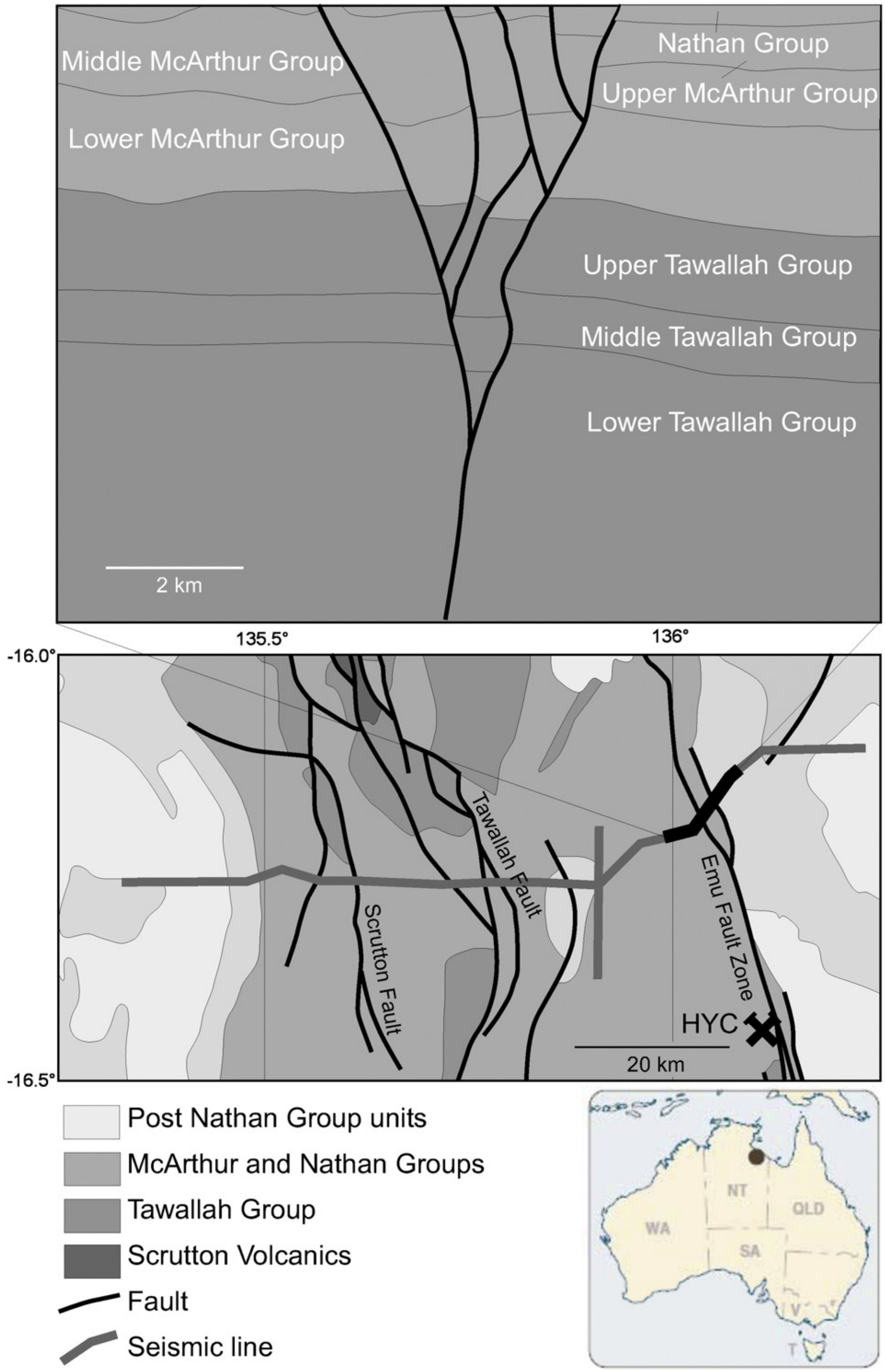

Fig. 8. Geologic environment during formation of the HYC deposit (after Rawlings et al., 2004).

deposit (Rawlings et al., 2004: Fig. 8), this fault is part of a flower structure, and changes in stratigraphic thickness within the structure suggest that it formed at the same time as sedimentation. The interpretation of syn-ore tectonic activity is supported by the presence of sedimentary breccias containing clasts of ore (Ireland et al., 2004). Such breccias are unusual as the character of the host rocks (carbonate-bearing, carbonaceous mudstone) suggests quiescent conditions, and are best interpreted as local responses to syn-depositional tectonic activity. This interpretation is further supported by paleomagnetic data (Symons, 2007), which suggests that paleomagnetic poles associated with mineralisation post-dated folding. These data suggest that the HYC deposit is genetically associated with syn-sedimentary deformation along the Emu Fault that occurred contemporaneously with the far-field Liebig Orogeny. Hence, at least some siliciclastic–carbonate Zn–Pb deposits appear to be associated with local deformation driven by major far-field tectonic events.

Finally, evidence from the Proterozoic Australian Zn–Pb deposits suggests that evolution in the tectonic environment can produce different kinds of mineral systems. The Curnamona Province in South Australia and western New South Wales, which hosts the Broken Hill siliciclastic–mafic deposit, is now considered to be a possible dismembered segment of the North Australian Superbasin System (Gibson et al., 2012). This deposit, and the coeval Cannington deposit in north Queensland, are thought to have formed early in the North Australian Superbasin System at ~1680 Ma (Sun et al., 1994; Page et al., 2005a,b) during early rift formation. The younger (1665–1575 Ma) siliciclastic–carbonate deposits in this basin system formed later as the basin system evolved from rift to thermal subsidence (Huston et al., 2006; Neumann et al., 2006; Gibson et al., 2012). A similar temporal relationship is seen from the Cambrian (~521 Ma: Goodfellow, 2007a) Anvil Range siliciclastic–mafic district to the Silurian (~436 Ma: Goodfellow, 2007a) Howards Pass siliciclastic–carbonate deposit in Canada.

8.2.2. Lake Superior-type banded iron formation

Lake Superior-type banded iron formation (BIF) is most extensively developed in Proterozoic passive-margin sedimentary basins (Gross, 1980, 1993), where it is interpreted to have formed through the upwelling of reduced, Fe-rich bottom waters onto passive margins, where the iron is deposited (Beukes and Gutzmer, 2008). These basins contain some of the largest known iron accumulations in the world, including extensive deposits in the Hamersley (Thorne et al., 2008, 2014; Angerer et al., 2014) and Transvaal (Beukes and Gutzmer, 2008) basins in Australia and South Africa, respectively. Lake Superior-type BIF deposits are characterised by their undeformed nature, lateral continuity and large size, and association with cratonic margin shelf successions, including platform carbonates, carbonaceous shale, and quartz-rich sandstones (Beukes and Gutzmer, 2008). This contrasts with Algoma-type BIF deposits (Section 9.2.2) that are associated with volcanic-dominated successions, are restricted in extent and volume, and have variable metamorphic and deformational characteristics (Gross, 1980, 1993). The third type of BIF, Rapitan-type, is temporally and spatially restricted, associated with transgression during interglacial periods related to Neoproterozoic "Snowball Earth" conditions (Klein and Beukes, 1993).

Lake Superior-type iron formation has a restricted temporal distribution, with virtually all deposits forming during the Archean and Paleoproterozoic. This distribution is the consequence largely of the evolution of the atmosphere and hydrosphere (e.g. Bekker et al., 2010), and not directly related to tectonic processes.

Although important accumulations of iron, iron formations of all types are only the precursor (about 30–35% Fe; Klein, 2005) to enriched (in many cases, >60% Fe; Hagemann et al., 2016-in this issue) iron ore deposits. The enrichment processes are often part of the intermediate stage orogenesis at convergent tectonic margins (Section 9.4.4), but can also be related to early basinal or late supergene processes.

8.2.3. Unconformity-related uranium deposits

Unconformity-related U deposits generally form at a redox boundary near the unconformity between a thick, sandstone unit and the underlying metamorphosed basement lithologies. The largest and best known unconformity-related deposits are associated with the Athabasca Basin in Canada and the Kombolgie Subgroup of the McArthur Basin in northern Australia. They are structurally-hosted either in the basement or in the unconformably overlying sandstone unit. Most known unconformity-related U deposits are hosted by Mesoproterozoic basins, and are spatially related to Paleoproterozoic orogens (Kyser and Cuney, 2008). These include broadly coeval (2.1–1.8 Ga) orogens in northern Australia and northern Canada, which formed during Nuna assembly. Orogenesis was followed by a period of continental readjustment and relative tectonic quiescence of around 500 million years. During this time several large intracratonic basins formed and evolved, including the Athabasca Basin in Canada and the Kombolgie basin in Australia.

8.2.3.1. Australian examples. In the Kombolgie Basin, far-field tectonic events are thought to have produced hydraulic gradients across the basins causing episodic fluid flow between ~1740 Ma and 1600 Ma (Polito et al., 2004, 2005; Orth et al., 2014). Based on higher precision (better

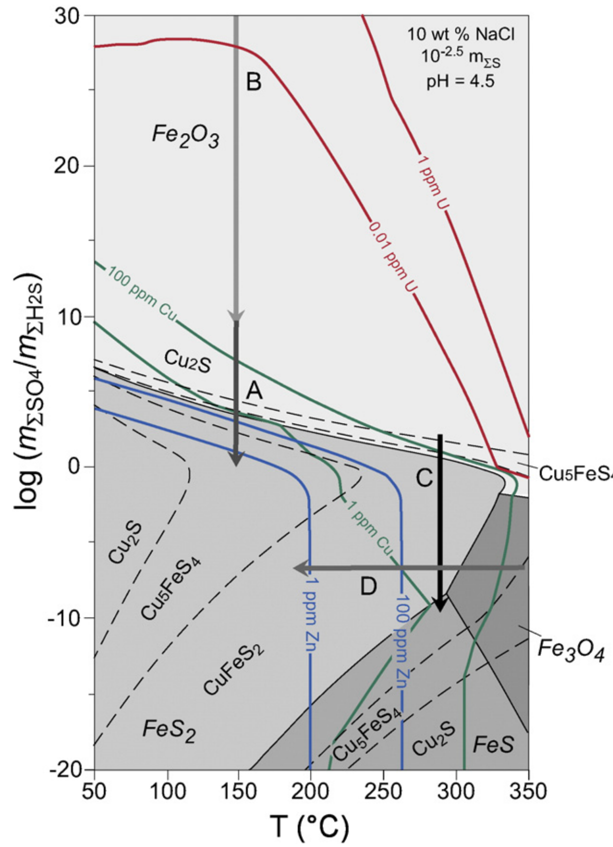

Fig. 9. log ($m_{\Sigma SO4}/m_{\Sigma H2S}$) diagram showing the solubility of Zn, Cu and U and the stability of Fe–S–O and Cu–Fe–S minerals. Shading indicates Fe–S–O mineral stability fields; dashed black lines indicate Cu–Fe–S mineral stability fields: coloured solid lines indicate solubilities of Cu, Zn and U. The diagram was calculated assuming a fluid salinity of 10 wt.% NaCl, a total concentration of sulphur species of $10^{-2.5}$ *m*, and a fluid pH of 4.5. Arrows indicate fluid evolution paths discussed in text (modified from Huston et al., 2006).

than 50 Myr) geochronology data, these events have produced deposits at Ranger (~1688 Ma: Skirrow et al., 2016-in this issue), Jabiluka (~1674 Ma: Polito et al., 2005), Nabarlek (~1642 Ma; Polito et al., 2004), Kylie (~1627 Ma: Von Pechmann, 1986) and Coronation Hill (~1607 Ma: Orth et al., 2014). Other data with poorer precision (>50 Myr: Maas, 1989) support the 1688–1607 Ma age range.

Although U mineralisation can occur above the unconformity in Canada, in Australia it typically occurs in the basement below the unconformity, commonly forming along major faults that transect the unconformity into the overlying basin (Mernagh et al., 1994, 1998). An unusual feature of the Australian deposits, that sets them apart from the Canadian deposits, is the presence of anomalous to ore-grade platinum group elements (Pt and Pd), although these zones can be separate from U-rich zones. As an example, a (pre-JORC) resource of 6.69 Mt of 6.42 g/t Au, 0.3 g/t Pt and 1.01 g/t Pd was defined at the Coronation Hill deposit (Ahmad et al., 2009) separate to U zones (0.344 Mt at 0.537% U_3O_8: McKay, 1990). Australian unconformity-related U deposits are associated with a sericite-chlorite ± kaolinite ± hematite alteration assemblage, and the composition of the host rock can vary from graphitic and chloritic schist, phyllite and shale, with some deposits hosted by carbonate rock (Mernagh et al., 1994, 1998).

Current models for the formation of unconformity-related U deposits can be divided into two types. One type involves the basement as the source of U and basins as the source of the fluids (Cuney, 2005) and the other involves the overlying basin as a source for both U and fluid (Kyser et al., 2000). In the first model U is sourced from the breakdown of monazite along fault zones as basinal brines interact with the basement. In the second model, U precipitates when the oxidised basinal brine carrying U reacts with a reduced basement lithology or mixes with reduced fluids. In this model, U is sourced from the breakdown of U-bearing detrital phases by basinal fluids in deep paleoaquifers (Kyser, 2007; Polito et al., 2011).

Outside of the Kombolgie Basin, other unconformity-related mineral deposits in Australia include the Kintyre and Westmoreland deposits (Jackson and Andrew, 1990; Mernagh and Wygralak, 2011). Although the Westmoreland deposits formed during the same mineralising events as the Kombolgie deposits (~1655 Ma: Polito et al., 2005), the Kintyre deposit, which is located near the unconformity between the Neoproterozoic Yeneena Basement and the underlying Paleoproterozoic Rudall Complex, is significantly younger at 837 ± 33 Ma (Cross et al., 2011).

8.2.4. Shale-hosted Cu ± Co ± Ag deposits

Globally, shale-hosted Cu ± Co ± Ag deposits, which include major deposits of the Kupferschiefer in central Europe and the Zambian Copper Belt in south-central Africa (Hitzman et al., 2005; Selley et al., 2005), are the largest producers of Co and major producers of Cu and Ag.

8.2.4.1. Australian examples. Although not large by world standards, Australia contains a number of shale-hosted Cu deposits in the Centralian and North Australian superbasins (Solomon and Groves, 2000). The first mining rush in Australia, during the mid-1840s, was to develop Cu deposits, including the Burra deposit, in the Adelaide Rift System. This basin, which is the southernmost extension of the Centralian Superbasin, initiated at prior to 800 Ma (probably ~830 Ma, associated with the Gairdner dike swarm: Wingate et al., 1998) as the supercontinent Rodinia broke up (Li et al., 2008). Copper deposits in the Adelaide Rift System are present at several stratigraphic levels, with the only well-dated stratigraphic position at ~797 Ma (Burra: Preiss et al., 2009).

Larger shale-hosted deposits, which include the Nifty and Maroochydore deposits, are present in the similar-aged Yeneena Basin in Western Australia, which is the western-most extension of the Centralian Superbasin. The age of the Nifty deposit is reasonably well constrained at 822 ± 23 Ma (age updated from Huston et al., 2007a), an age which overlaps with that of the nearby Kintyre unconformity-related U deposit. Moreover, Smith (1996) estimated a Pb-Pb model age of ~840 Ma for the Warrabarty Zn–Pb deposit, also hosted by the Yeneena Basin. These data suggest contemporaneous or near contemporaneous Cu, U and Zn–Pb mineralisation, possibly as part of a large regional mineral system. This mineral system is interpreted to have been active either late during Yeneena Basin formation, or during the first phase of basin inversion (Huston et al., 2010a; e.g., Fig. 6D). This age, and the limited ages from the Adelaide Rift System are similar to the oldest ages reported (~825 Ma: Barra-Pantoja, 2005) from the Zambian Copper Belt.

In the North Australian Superbasin, the age and origin of shale-hosted Cu deposits is not as well constrained. These deposits include the Mount Isa Cu orebodies, which are thought to have formed syntectonically at ~1523 Ma (Perkins et al., 1999), ~130 Myr after deposition of the host succession, and deposits in the Gunpowder district (Richardson and Moy, 1998), which are undated. Although the hosts to most of these deposits are dolomitic shale (e.g. the Urquart Shale at Mount Isa), some are hosted by sandstone (e.g. Mammoth and Esperanza in the Gunpowder district: Solomon and Groves, 2000; Richardson and Moy, 1998; Hutton et al., 2012).

8.2.4.2. Comparison with other global examples. As discussed by Hitzman et al. (2005), many shale-hosted Cu ± Co ± Ag deposits have peripheral Pb–Zn halos that are part of a deposit- to regional-scale metal zonation that can be related to the oxidation state of the host rocks. This has been well documented in the Kupferschiefer deposits in Poland (Oszczepalski, 1999), where the host rocks progressively become oxidised with stratigraphic depth, with a highly oxidised (Fe^{3+}-dominated; Fe_2O_3/TOC[1] > 15), but barren Rote Faule underlying a Cu–Co mineralised transitional zone (Fe_2O_3/TOC > 2), which, in turn, underlies a Zn–Pb–Cu-mineralised reduced zone (Fe_2O_3/TOC < 0.8). This zonation is interpreted to be the consequence of the influx of highly oxidised, diagenetic brines from below, with mineralisation caused by progressive reduction of the ore fluid (and oxidation of the host succession) as it interacted with initially organic-carbon-rich shale and dolomite (Oszczepalski, 1999). Uranium, Au and PGEs also form part of this metal zonation. Oszczepalski (1999) indicated that Au and PGEs are enriched in the oxidised zone and highly enriched (to 100 ppm Au and 14 ppm Pt) in the transition zone, but low in the reduced zone. Piestrzyński (1990) indicated that U is most enriched in the oxidised zone (Weißliegendes sandstone; to 677 ppm), but also in the reduced zone (to 360 ppm). Although not as well documented as in the Kupferschiefer, shale-hosted Cu deposits in the Yeneena Basin are also zoned. At Nifty, Anderson et al. (2001) documented a zone of Zn–Pb enrichment that stratigraphically overlies the Cu-rich ore, the pyrite marker chert. They also documented the presence of uranite and pitchblende in the Cu-rich ores.

Although most workers infer that shale hosted Cu ± Co ± Ag deposits form from oxidised and saline, low to moderate temperature ore fluids (e.g. Hitzman et al., 2005), there is surprisingly little data on fluid characteristics from this deposit type. Limited fluid inclusion (Annels, 1989; Tonn et al., 1987; Breit and Meurnier, 1990) data from the Zambian Copper Belt, the Kupferschiefer and the Paradise Basin in the western United States suggest low temperature (70–170 °C), moderate to high salinity (7–23 wt.% NaCl eq) brines. At Mount Isa, most fluid inclusion homogenisation temperatures for the silica-chalcopyrite stage of ore formation were 140–180 °C, but it was not possible to estimate the pressure correction (Heinrich et al., 1989). Although Heinrich et al. (1989) originally proposed a reduced ore fluid, subsequent work (Heinrich et al., 1993; Waring et al., 1998) proposed a relatively oxidised ($\Sigma SO_4 \geq \Sigma H_2S$), high temperature (300 °C) ore fluid.

8.2.5. Chemical links between basin-hosted U, Cu ± Co ± Ag and Zn–Pb–Ag deposits?

As discussed above, the metallogeny of U, Cu ± Co ± Ag and Zn–Pb–Ag is linked in a number of basins around the world. This linkage is seen both at the deposit and basin scale. In many deposits, these metals are spatially zoned (e.g. Kupferschiefer) and in some basins, deposits with these diverse metal assemblages appear to closely coincide in time. Although in detail these deposits and basins are in many ways different, in most cases the ore fluids are thought to be oxidised, saline and sulphide poor, with production of sulphide at the site of mineralisation being a critical process in the mineral systems.

Brown (1971), Rose (1976) and Kirkham (1989) proposed that ore deposition in shale hosted Cu ± Co ± Ag deposits was caused by reduction of initially oxidised fluids, noting that this process can account for the Cu–Fe–S mineralogical (chalcocite → bornite → chalcopyrite → pyrite) and metal zonation (Cu → Zn–Pb) observed at many deposits. Fig. 9 illustrates this process at low temperatures (100–150 °C). As the ore fluid is reduced (arrow A), Cu solubility decreases and initially chalcocite and then bornite are deposited with hematite, followed by the deposition of chalcopyrite and pyrite. Zinc (and Pb) deposition only occurs when sufficient H_2S becomes available (i.e. when $\Sigma H_2S \rightarrow \Sigma SO_4$). If the initial fluid conditions are sufficiently oxidised, U precipitates at the very earliest stage of the reduction processes (arrow B).

If temperatures are sufficiently high (e.g. ~300 °C, as suggested for the Mount Isa Cu orebodies: Heinrich et al., 1989; Waring et al., 1998) and the fluids are only moderately oxidised (Waring et al., 1998), the ore fluids cannot transport significant U, and Zn (and Pb) will not

[1] TOC = total organic content.

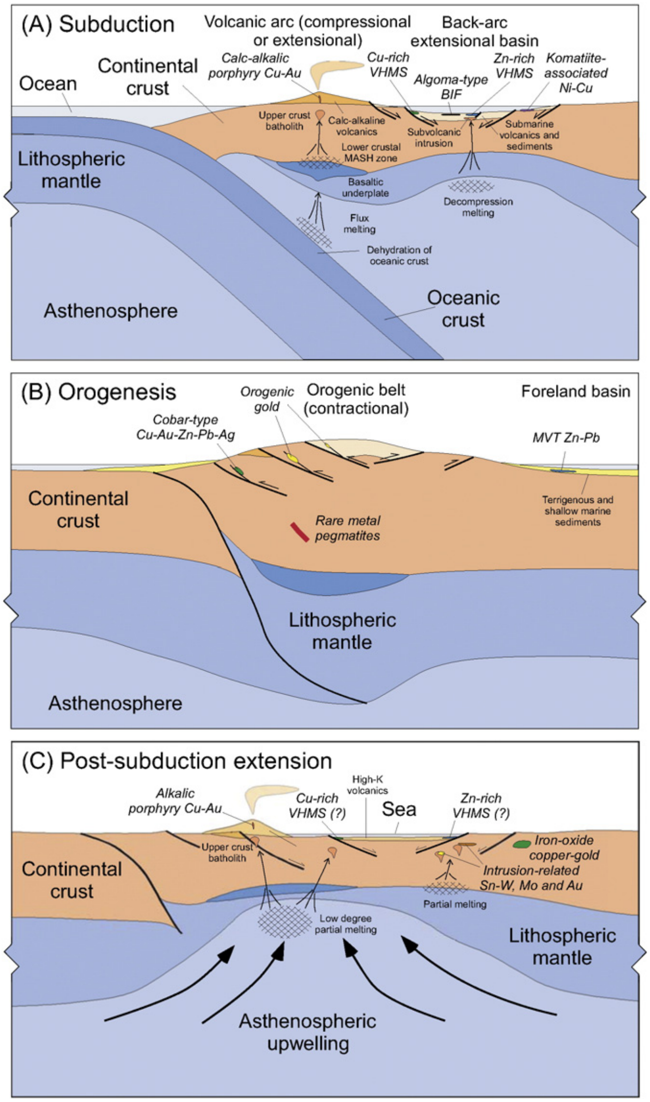

Fig. 10. Schematic diagram showing stages in the evolution of a convergent margin and associated mineral deposits: (A) subduction, (B) orogenesis, and (C) post-subduction extension. This diagram builds upon an original diagram from Richards (2009).

precipitate as the fluid evolution parallels Zn solubility contours (arrow C in Fig. 9). This evolution path will, in fact, dissolve Zn due to the very high Zn solubilities (>500 ppm), which is consistent with observations at Mount Isa that the Cu orebodies replace the Zn–Pb orebodies (Waring et al., 1998).

The presence of coeval U, Cu ± Co ± Ag and Zn–Pb deposits in some metallogenic provinces suggests that these depositional processes and paths might apply at a larger scale, with redox gradients within basins producing a range of deposits from oxidised fluids that are the natural products of relatively oxidised rock assemblages that characterise many thermal subsidence basins or passive margins. Importantly, these relationships suggest that in these oxidised basins, the presence of a deposit characterised by one metal assemblage may indicate the presence of geologically-related metal assemblages, either as part of deposit-scale zonation or as separate deposits along redox gradients. As this basin-hosted metallogenic system required oxidised ore fluids that form in oxidised basins, it is restricted temporally to after the oxidation of the atmosphere in the Paleoproterozoic, which, for the first time, produced oxidised rock packages and surficial fluids (Leach et al., 2010). Another factor that may be important is the amalgamation and breakup of the supercontinent Nuna (and, later, Rodinia), which produced extensive passive margins and shallow seas in which this metallogenic system developed (Leach et al., 2010).

9. The tectono-metallogenic system related to convergent margins

As discussed by numerous authors, and shown in Fig. 10, tectonic processes that occur along convergent margins are not restricted to the formation of arcs and related basins, but include orogenesis caused by the accretion of arcs, oceanic plateaux, seamounts, microcontinents and and/or other exotic terranes, or by changes in the vector or dip of subduction. Post-subduction extension commonly follows and can be associated with underplate delamination (e.g., Richards, 2009) or be the consequence of an evolving tectonic system (see Section 9.5). Although convergent margins are dominated by convergence between two plates, and generally involve thickening of the over-riding plate to form magmatic arcs, this tectonic system can involve (significant) local extension, for example the formation of back-arc basins. Moreover, if subduction and subsequent accretion is oblique, significant zones of transpression or transtension can develop along convergent margins.

In the following discussion, the evolution of convergent margins is divided into three stages. The first stage is subduction, which is accompanied by the formation of volcanic arcs and back-arc basins in the over-riding plate (Fig. 10A). Importantly, this stage can involve both thickening and thinning of the over-riding plate. It is followed by an intermediate stage of orogenesis, during which the over-riding plate goes fully into contraction, due to either a change in the geometry of subduction or the accretion of an exotic block (Fig. 10B). The final stage is post-subduction extension (Fig. 10C), which can be accompanied by post-orogenic collapse to compensate for gravitational instability, re-initiation of subduction, delamination of underplated materials formed during the earlier stages of convergence, and other processes (Section 9.5).

These convergent stages produce a large range of ore deposit types, including porphyry Cu-Mo-Au and related epithermal and skarn deposits, orogenic gold deposits, Mississippi Valley-type (MVT) deposits, granite-related Sn–W–Mo deposits, and, possibly, iron oxide–copper–gold (IOCG) deposits. In addition, ancient volcanic-hosted massive sulphide (VHMS) deposits appear to be restricted to convergent margins, as subduction and seafloor oxidation (after the Great Oxidation Event) appear to remove deposits formed along divergent mid-ocean ridges (Huston et al., 2010b). Although some of these types of deposit seem to be associated with specific stages of the convergent tectono-metallogenic system (e.g. orogenic gold and MVT deposits with orogenesis), others span this system. For example, although porphyry Cu ± Au ± Mo and related deposits are mostly associated with subduction, new data suggest that an important sub-type of these deposits – those associated with alkaline magmatism — can be associated with post-subduction extension (Müller and Groves, 1993; Crawford et al., 2007; Richards, 2009). In addition to the temporal patterns noted above, in some cases there is a distinct spatial distribution pattern within specific stages, as discussed below. Finally, the time scale over which convergent tectono-metallogenic systems evolve can be very rapid, commonly less than 50 Myr (Section 9.5), and in some cases, existing geochronology may be too coarse to resolve stages, particularly in older systems.

9.1. Secular variations in convergence and implications for mineral systems

The tectono-metallogenic systems approach integrates mineral systems with tectonic systems and relates their evolution to geodynamic processes. However, it is likely that both tectonic processes and underlying geodynamic drivers have changed through Earth's history. Although some secular trends are evident within the rock record, such as the restriction of komatiites largely to the Archean and Paleoproterozoic (Arndt et al., 2008), the underlying causes of these changes are controversial. Although a detailed discussion is beyond the scope of this paper, we provide below a brief summary of how geodynamic processes may have changed over time, and how these changes have affected both tectonic and metallogenic systems.

The Archean was a period of fundamental changes in geodynamic processes on Earth, which changed from a magma ocean during the Hadean to a planet with some features of modern-style plate tectonics (Fig. 11). The fundamental underlying process governing these changes is the thermal evolution of the mantle. Most models suggest the mantle was 150–250 °C hotter in the mid- to late Archean, relative to present day, and cooled from the late Archean–early Proterozoic onwards (e.g. Herzberg et al., 2010; van Hunen and Moyen, 2012). Secular changes in global geodynamics can be directly, or indirectly, linked to this cooling process (e.g. van Hunen and van der Berg, 2008) and changes to the manner in how heat is lost through time (e.g. Stern, 2007). The Archean is considered to have been characterised by the transition from an unstable stagnant-lid tectonic regime (Brown, 2014; Johnson et al., 2014), with intense mantle plume activity and/or mantle overturn events (Griffin et al., 2014), to (some form of) a plate tectonic regime (e.g. Pease et al., 2008). The transition seems rather gradational resulting in the coexistence of both tectonic regimes during this time period, and possibly for some time thereafter (e.g. Stern, 2008).

Other processes may have changed/stopped more abruptly towards the end of the Archean, such as the formation of continental crust and subcontinental lithospheric mantle (SCLM). Isotopic data suggest that the Archean was a period of rapid continental crustal growth, with over 60% of continental crust formed by its end (Belousova et al., 2010; Dhuime et al., 2012). Most (70%) of the SCLM is suggested to have formed between 3.3 and 2.7 Ma (Griffin et al., 2014 and the references therein; Fig. 11). Compared to SCLM that was formed later, Archean SCLM is highly depleted and thus buoyant and rigid (Boyd, 1998), both attributes which facilitated its preservation through time. The formation of these lithospheric blocks has profound consequences for geodynamics, because they may assemble to form larger cratons, facilitate far-field stress transfer, and focus deformation (and convergence) along their margins (e.g. Begg et al., 2010).

A number of secular trends are evident in the geological record that can be used to constrain tectonic regimes (Fig. 11). Although the interpretation of some early geological features as indicators for plate tectonics (e.g. the ~3.8 Ga Isua greenstone belt in Greenland as an accretionary complex: Komiya et al., 1999) remains controversial, there is increasing consensus that from the Neoarchean onwards (<2.8 Ga) plate tectonics became the dominant tectonic regime (e.g. Cawood et al., 2006; Condie and Kröner, 2008). Arguments in support of this interpretation include the (widespread) appearance of passive margins (Bradley, 2008), arc volcanic rocks (e.g. boninites, shoshonites) and paired metamorphic belts (Brown, 2014). Isotopic and mineral chemistry on sulphide and silicate inclusions in diamond indicate a major compositional change occurred at 3 Ga after which eclogitic inclusions became prevalent (Shirey and Richardson, 2011). The style of plate tectonics, in particular the style of Precambrian subduction, however, is debated, and current views are strongly informed by theoretical considerations and numerical modelling (Korenaga, 2013; Gerya, 2014).

Undoubtedly, a hotter mantle had significant consequences on convergence and subduction zone dynamics in the Archean. For example, melting would have started at greater depth and occurred at higher temperatures (van Thienen et al., 2004), leading to thicker, more buoyant oceanic crust and SCLM. Moreover, the mantle would have had a lower viscosity, providing less support for a subducting slab, which would have had lower strength and coherency (van Hunen and Moyen, 2012). The latter two characteristics are particularly important, as thermo-mechanical models suggest that they contribute to frequent slab break-off in the Archean (van Hunen and Van der Berg, 2008). This effectively removes the slab pull, and would lead to a much more short lived, episodic style of subduction (or "proto-subduction"; Moyen and van Hunen, 2012). Moyen and van Hunen (2012) suggested that this proto-subduction evolved into more sustainable subduction settings as a function of mantle cooling.

Using a variety of parameters assumed to be different in the Archean (e.g. radiogenic heat production, melt-related weakening, thickness of the oceanic crust, and oceanic plate velocity), Sizova et al. (2010) modelled the initiation and style of subduction, identifying three tectonic regimes: (1) "no-subduction", where, due to extremely weak plates, horizontal movements are accommodated by internal strain; (2) "pre-subduction" regime, where plates can deform internally due

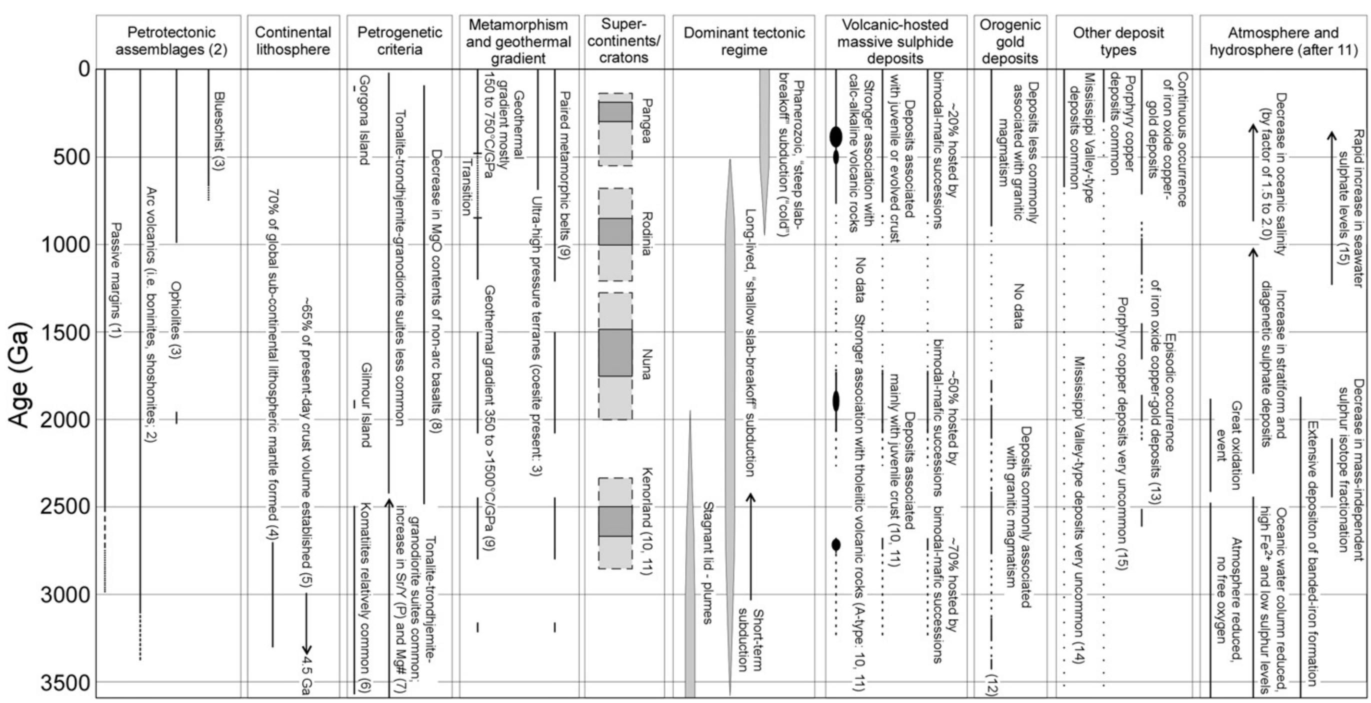

Fig. 11. Overview of secular trends of the geological record and mineral deposits, partly compiled from literature. Numbers in brackets correspond to references: 1 — Bradley (2008); 2 — Condie and Kröner (2008); 3 — Stern (2007); 4 — Griffin et al. (2014); 5 — Dhuime et al. (2012); 6 — Arndt et al. (2008); 7 — Smithies and Champion (2000); 8 — Herzberg et al. (2010); 9 — Brown (2014); 10 — Huston et al. (2015); 11 — Huston et al. (2010b); 12 — Goldfarb et al. (2001); 13 — Groves et al. (2010); 14 — Leach et al. (2010); 15 — Richards and Mumin (2013).

to melt-related weakening, and convergence causes shallow underthrusting of the continental plate by the oceanic plate without formation of a mantle wedge; and (3) “modern subduction”, where stable one-sided subduction has been possible. The transition between the regimes is controlled by mantle temperature, with the first transition taking place at mantle temperatures 200–250 °C above present day values, and the transition to modern subduction at temperatures of 175–160 °C above present day values, during Meso- to NeoArchean time (Sizova et al., 2010).

The Neoproterozoic–Paleozoic transition is another time in Earth history where significant changes are preserved in the geological/rock record. These include the first widespread occurrence of ophiolites and blueschists (e.g. Stern, 2007), ultra-high pressure (UHP) terranes, and “cold” apparent geothermal gradients in subduction zone settings (Brown, 2014). This period is interpreted to mark the transition from a shallow- to deep slab break-off, and hence deeper subduction, which creates a colder environment in the subduction zone (Maruyama and Liou, 1998; Brown, 2014). Thermomechanical numerical modelling by Sizova et al. (2014) suggests that the transition occurs at ambient mantle temperatures 80–100 °C warmer than present, during the late Proterozoic.

These transitions in geodynamic drivers can be related to changes in the characteristics of mineral systems, for example the volcanic-hosted massive sulphide (VHMS) mineral system. Prior to ~2.8 Ga, these deposits, though present, are reasonably uncommon, with the most significant deposits being those in the Golden Grove district in the Youanmi Terrane of the Yilgarn Craton (Fig. 4). This district is thought by some to have formed in an intracratonic setting, where crustal attenuation may be attributed to mantle plume activity (Ivanic et al., 2012). Alternatively, the setting could be a back-arc basin. Volcanic-hosted massive sulphide deposits in the Eastern Pilbara Granite-Greenstone Terrane where they are thought to have formed in a geodynamic setting analogous to an oceanic plateau (Huston et al., 2007b; Van Kranendonk et al., 2014). In general, these deposits are not considered to have formed in a subduction setting.

The first major peak in VHMS abundance, between 2.75 and 2.65 Ga (Huston et al., 2010b) corresponds to the time when “proto-subduction” is interpreted to have been prevalent. These deposits and those that formed the second major peak in VHMS abundance at 1.90–1.70 Ga are hosted by volcanic successions with higher abundances of mafic volcanic rocks relative to the third peak, which formed at 0.75–0.00 Ga (Fig. 11). Moreover, the first and second peaks are most closely associated with volcanic successions with tholeiitic affinities (FIIIa and FIIIb types of Lesher et al., 1986), whereas the third peak is associated with calc-alkaline volcanic successions (FII type; Hart et al., 2004).

The changes between the second and third peaks of VHMS abundance correspond to the transition between shallow and deep slab breakoff. Continuation of subducting slabs at depth has two significant consequences. First, the presence of a coherent, deep and cold slab would favour slab pull and rollback, as rollback is most likely driven by the negative buoyancy of the subducting plate (Schellart, 2008). Second, subduction of crust would encourage fertilisation of the mantle wedge, enhancing the production of calc-alkaline magmas. Within this context the change in the geochemical affinity of associated rocks is remarkable: while Neoproterozoic and Phanerozoic VHMS deposits are affiliated with calc-alkaline rocks, which indicate proximity to a subduction setting, tholeiitic volcanic rocks associated with Paleoproterozoic and Archean systems may indicate a greater geographic spread between back-arc basins, which host the deposits, and the associated arcs.

In summary, this section warrants a note of caution when using an overly uniformitarian (i.e. modern/Phanerozoic plate tectonic) view as a reference frame for mineral systems, because other tectonic regimes (stagnant lid, plume-related, mantle overturn) have been present on Earth. Moreover, the changes in plate tectonic style and subduction zone configuration since its suggested onset in the Archean until at least the Phanerozoic, are still debated. Integration and consideration of these secular changes (and related uncertainties) into the mineral systems assessment, however, provides a powerful tool for exploration.

9.2. Initial stage — subduction and formation of arcs and back-arc basins

On modern Earth, two broad types of convergent margins are known: (1) East Pacific-type margins, and (2) West Pacific-type margins (Stern, 2002). Although both of these convergent margins contain magmatic arcs and associated fore-arc basins, West Pacific-type margins have extensive back-arc basins, whereas in East Pacific-type margins, such as the Andean cordillera, back-arc basins are not as well developed. This difference is related to the interaction between the converging plates, including the relative rates at which the over-riding plate advances, and the subducting plate sinks (and so retreats oceanward). The rate of sinking is related to the age of the subducting slab (i.e. young, hot crust versus old, cold crust: Stern, 2002). If the rate of advance is greater than the rate of retreat, the over-riding plate is entirely in contraction, and back-arc basins, which are a product of extension of the over-riding plate, do not form. This style of convergence is termed advancing accretionary orogenesis and results in East Pacific-type margins (Cawood and Buchan, 2007). If the rate of advance is less than the rate of slab sinking and retreat, the over-riding plate is partly, or wholly, in extension, and back-arc basins form, producing West Pacific-type margins (Cawood and Buchan, 2007). Importantly, the style of convergence has a significant influence not only on the types of mineral systems that develop, but on the architecture available to and used by these systems, as the convergent margins evolve from subduction to orogenesis, as discussed below. Systems can switch rapidly from one style to the other (e.g. Collins, 2002a) and more than one style of convergence may exist along convergent margins, suggesting that mineral systems can evolve in a complex manner in space and in time along individual convergent margins. This is well illustrated by the Lachlan tectonic cycle in the Tasmanides as discussed in Section 9.5.

9.2.1. Calc-alkaline porphyry Cu ± Au ± Mo deposits

Fig. 10A schematically illustrates the distribution of mineral deposits formed during the first stage of the convergent tectono-metallogenic system. Porphyry Cu ± Au ± Mo and related epithermal deposits associated with calc-alkaline magmatism form in magmatic arcs that develop over the most thickened part of the over-riding plate and appear to have formed in overall contractional settings (Sillitoe, 1998; Cooke et al., 2005). Importantly, these deposits form in both advancing (e.g. Andean margin) and retreating (e.g. Philippines, Papua-New Guinea) accretionary convergent systems. The magmas responsible for subduction-related porphyry-epithermal mineral systems were derived from the melting of a subduction-enriched mantle wedge, with ascent into the upper crust driven by buoyancy (Richards, 2003: Fig. 10A). According to Sillitoe (1998) and Cooke et al. (2005), the overall contraction setting of this environment promotes the formation of porphyry Cu ± Au ± Mo deposits because (1) contraction impedes magma ascent through the upper crust and, thus, impedes volcanism, (2) the resultant magma chambers in these settings are larger, (3) as eruption is impeded, fractionation in these magma chambers is promoted, resulting in the generation of large volumes of magmatic-hydrothermal fluids, (4) contraction restricts the number of apophyses that form on the magma chamber, providing the potential for more efficient fluid focusing, and (5) rapid uplift and erosion promotes efficient extraction and transport of magmatic-hydrothermal fluids due to abrupt decompression. Formation of porphyry Cu ± Au ± Mo deposits seems to be triggered by perturbations in the overall contractional stress field that reflect changes in the geometry of the convergent system (e.g. changes in slab dip or vector or relative motion). Although details of the relationships and characteristics of porphyry Cu ± Au ± Mo and related deposits are beyond the scope of this contribution, this information can be found in Titley and Beane (1981), Seedorf et al. (2005), Simmons et al. (2005) and numerous other publications.

9.2.1.1. Australian examples. Compared to many other countries (e.g. Chile, Peru, USA, Indonesia and Philippines), Australia is poorly endowed with porphyry Cu ± Au ± Mo and related deposits, probably because Australia does not contain the young (Cenozoic) magmatic arc rocks that preferentially host these deposits. Rather, the limited Australian porphyry Cu ± Au ± Mo and related deposits are much older, ranging in age from Paleoarchean to Permo-Triassic. At ~3325 Ma (Huston et al., 2007b), the Spinifex Ridge (Coppins Gap) Cu–Mo deposit is the oldest known porphyry deposit in the world. This deposit is thought by some workers (e.g. Smithies et al., 2005; Van Kranendonk et al., 2014) to have formed in an oceanic plateau environment, although others (Barley, 1997) infer a subduction-related origin. The first phase (~2740 Ma: Stein et al., 2001) of mineralisation at the giant Boddington Au–Cu deposit is interpreted to have formed as a porphyry-style deposit (McCuaig et al., 2001) within the Saddington oceanic arc (Qui et al., 1997; Korsch et al., 2011). Small porphyry-style deposits, with probable ages of ~1840 Ma, are known in the Halls Creek Orogen (Hassan, 2000). In Western Victoria, the Cambrian Stavely Arc is interpreted to have formed in a continental margin setting along the east Gondwanaland margin (Cayley et al., 2011a) and contains a number of porphyry-style Cu prospects (Willocks et al., 1999). This system extends north, beneath Murray Basin cover, into western NSW as the Mount Wright Arc (Greenfield et al., 2011).

Economically the most significant porphyry Cu–Au deposits in Australia are those in the Ordovician to Silurian Macquarie Arc (Cadia and Northparkes districts). Although most of these deposits are interpreted to have formed during post-subduction alkaline magmatism (Crawford et al., 2007; Section 9.4.1), a small number of older deposits (e.g. the Marsden and the ~450 Ma Copper Hill deposits: Perkins et al., 1995) are associated with calc-alkaline volcanic rocks (Crawford et al., 2007) and formed during subduction.

9.2.2. Volcanic-hosted massive sulphide deposits and Algoma-type banded iron formation

Although porphyry Cu ± Au ± Mo and related deposits form in both advancing and retreating accretionary systems, modern VHMS deposits that form along convergent margins[2] seem to be restricted to retreating accretionary systems as extension of the over-riding plate forms back-arc basins and/or rifted arcs that are essential to the formation of these submarine systems (Fig. 10A). Volcanic-hosted massive sulphide deposits are some of the most widespread on earth. They formed through much or Earth's history and are known to exist on all continents except Antarctica (Franklin et al., 1981, 2005).

Most ancient VHMS deposits are interpreted to have formed during volcanism and sedimentation associated with the formation of back-arc basins or rifted arcs, and magmatism caused by decompression melting of the lower crust and upper mantle (Fig. 10A). The deposits form both as exhalative bodies at, or as replacement bodies just below, the seafloor. Although not an exclusive relationship, these deposits are commonly associated with felsic volcanic rocks, even in mafic-dominated volcanic successions (Franklin et al., 1981, 2005; Hart et al., 2004). The deposits can be associated with changes in volcanic facies or with syn-depositional extensional structures (Doyle and Allen, 2003). In many cases, there is a temporal and spatial relationship with subvolcanic intrusions and with regional alteration zones (Galley, 1993, 2003).

In Archean and, possibly, Paleoproterozoic terranes, VHMS deposits seem to be associated with juvenile, attenuated continental crust, as indicated by Nd and Pb isotope characteristics (Huston et al., 2014; Champion and Huston, 2016-in this issue), but this relationship breaks down in younger terranes. Moreover, older VHMS deposits are strongly associated with high temperature (A-type), felsic volcanic rocks with tholeiitic affinities (the FIII types of Lesher et al., 1986). This relationship breaks down for younger terranes (Piercey et al., 2001; Hart et al., 2004), however, suggesting that the tectonic setting that hosts VHMS deposits may have changed over time. The changes in the geochemistry of VHMS-associated volcanic rocks suggest that these changes with time could relate to a general cooling of the upper mantle, longer stability of arc systems, the increasing importance of evolved continental crust as a substrate to the arc systems, and/or to an increasing importance of (peri-)continental rifted arc and back-arc (as opposed to oceanic back-arcs) environments as preferential loci for the formation of VHMS deposits (e.g. Lentz, 1998; Piercey et al., 2001; Piercey, 2011).

In the last twenty years a new type of VHMS deposits has been recognised - Cu-Au-rich deposits associated with advanced argillic alteration assemblages, or “high sulphidation” deposits (Sillitoe et al., 1996; Hannington et al., 1999; Mercier-Langevin et al., 2011). In Canada, these deposits (e.g. those in the Bousquet district: Dubé et al., 2007, 2014; Mercier-Langevin et al., 2007a,b) have become attractive exploration targets. In Australia, the Mount Lyell (Huston and Kamprad, 2001) and Henty (Callaghan, 2001) deposits in western Tasmania are examples of these deposit types, which are thought to be hybrid between the porphyry-epithermal deposits that form in arcs, and the more typical Zn-rich deposits that form in back-arcs. Some workers (e.g. Large et al., 1996; Hannington et al., 1999; Huston et al., 2011; Mercier-Langevin et al., 2011; Dubé et al., 2014) have suggested that “high sulphidation” Cu–Au-rich VHMS deposits may have a significant magmatic-hydrothermal component. The Mount Morgan Au–Cu deposit (Messenger et al., 1998) may also be an example of a hybrid deposit. Geochemically, hybrid Archean deposits are associated with type FI and FII felsic volcanic rocks of Lesher et al. (1986), suggestive of a calc-alkaline affinity. Spatially, these deposits may form in the transition from arc to back-arc basin (Mercier-Langevin et al., 2007a,b; Huston et al., 2011).

9.2.2.1. Australian examples. Volcanic-hosted massive sulphide deposits/events occur through much of the geologic history of Australia, from ~3480 Ma (Dresser deposit: Van Kranendonk et al., 2008) to ~275 Ma (Mount Chalmers: Crouch, 1999). The earliest significant deposits are those in the ~3240 Ma Panorama district in the East Pilbara Granite-Greenstone Terrane (Brauhart et al., 1998, 2001). Other important districts include the ~2950 Ma Golden Grove district in the Youanmi Terrane (Ashley et al., 1988; Sharpe and Gemmell, 2002), the ~505–500 Ma Mount Read district in western Tasmania (Seymour et al., 2006; Mortensen et al., 2015), the ~480 Ma Mount Windsor and Balcooma districts in north Queensland (Berry et al., 1992; Huston et al., 1992), the ~475 Ma Girilambone district in northern New South Wales (Huston et al., 2016-in this issue) and ~420 Ma deposits associated with extensional zones in southeastern New South Wales and north-eastern Victoria (Champion et al., 2009).

9.2.2.2. Links to Algoma-type banded iron formation. It was recognised shortly after syngenetic models for VHMS were developed that, in many districts, the deposits were associated with Algoma-type BIFs at or near the stratigraphic position of mineralisation (Stanton, 1960). In contrast to Lake Superior-type BIFs, which form on passive margins (Section 8.2.2), Algoma-type deposits are much smaller, are associated with volcanic- and turbidite-dominated successions, and are thought to be the products of exhalation related to volcanism (Gross, 1980). Although developed predominantly during Archean and Paleoproterozoic, the association of VHMS deposits with Algoma-type BIFs is also present in some younger districts, such as the Ordovician Bathurst district in New Brunswick and the Hokuroku district in Japan (Goodfellow, 2007b; Kalogeropoulos and Scott, 1983).

9.2.3. Orthomagmatic komatiite-associated Ni–Cu–PGE deposits

The other significant type of deposit that formed along continental margins is the Ni–Cu–PGE type associated with komatiitic volcanic and related shallow intrusive rocks. These deposits are typically

[2] Volcanic-hosted massive sulphide deposits also form along mid-ocean ridges, but these are unlikely to be preserved in the geological record as discussed earlier.

associated with mafic–ultramafic greenstone sequences that are controlled by the interaction of mantle plumes with the thinned margins of continental lithosphere (Mole et al., 2014). Komatiites are found in most Archean granite-greenstone terranes, but nickel-sulphide mineralisation associated with these rocks has a very restricted distribution. Although the vast majority of Archean komatiite-associated Ni–Cu–PGE deposits are present in the Yilgarn Craton in Western Australia, other Archean provinces contain small deposits, including the greenstone belts in the Abitibi Subprovince of Canada, and the Zimbabwe Craton (Naldrett, 1981).

9.2.3.1. Australian examples. The global endowment of nickel-sulphides in Archean komatiites is overwhelmingly dominated by the Kalgoorlie Terrane of the Eastern Goldfields Superterrane in the Yilgarn Craton of Western Australia. Barnes and Fiorentini (2012) questioned whether the Kalgoorlie Terrane komatiites possess any exceptional attributes that could explain this bias through an exhaustive compilation of geochemical data from this terrane and a number of other terranes containing komatiite-hosted nickel-sulphide mineralisation worldwide, including the southeastern Youanmi Terrane of the Yilgarn Craton, the eastern terranes of the Eastern Goldfields Superterrane (Kurnalpi, Burtville and Yamarna terranes), and the Abitibi Subprovince in Canada. The presence of adcumulate dunites, formed by high magma fluxes in central conduits, is the common feature between all of these mineralised terranes. Coupled with evidence for higher degrees of contamination in the Kalgoorlie Terrane, and the availability of accessible crustal S sources (Bekker et al., 2009; Fiorentini et al., 2012), it appears that magma flux, rather than primitive magma composition or source (cf. Zhang et al., 2008), was the critical factor, and that craton-scale deep lithospheric structure is the ultimate control on rates of magma transfer between mantle source and crustal emplacement site (cf. Begg et al., 2010). The Kalgoorlie Terrane komatiites were emplaced at exceptionally high rates, giving rise to well-developed long-lived magma conduits, either lava tubes or subvolcanic channelised sills, which are interpreted to be the essential condition for forming large deposits (Barnes and Fiorentini, 2012).

Begg et al. (2010) discussed the role of lithospheric setting on the localisation and clustering of nickel-sulphide mineralisation associated with mafic and ultramafic magmas. These authors based their arguments on observations derived from large tomographic data sets and on whole-rock Sm-Nd analyses of granitoid rocks hosting variably endowed greenstone belts, which were used as proxies of the timing and nature of crustal differentiation processes (Champion and Cassidy, 2007). More recently, other authors put forward the idea that these komatiite-hosted nickel-sulphide deposits, which are restricted to Archean and Paleoproterozoic terranes, are associated with isotopically more evolved crust (Barnes and Fiorentini, 2012; Huston et al., 2014; Champion and Huston, 2016-in this issue) in contrast to the juvenile crust that characterises VHMS-rich domains. Finally, Mole et al. (2012, 2013, 2014) illustrated that the evolving nature of the lithospheric architecture has played a key role over the entire Archean aeon in the genesis of komatiite-hosted nickel-sulphide camps. These authors utilised radiogenic isotopes as proxies for the age and thickness of the lithosphere, emphasising how craton margins are preferred loci for the setting of mineralised komatiite camps. Unlike whole-rock Sm–Nd isotope systematics, which allow imaging of lithospheric architecture roughly at the time of felsic magmatism, the application of in-situ Lu-Hf techniques were utilised as a "paleo-geophysical" tool to image cryptic craton boundaries through time (Mole et al., 2014).

9.3. Intermediate stage — orogenesis

The incipient stage of the convergent margin tectono-metallogenic system is typically ended by either accretion of an exotic block (large or small: Fig. 10B) or a shallowing of subduction (Collins and Richards, 2008), which places the over-riding plate into contraction. This results in orogenesis and metamorphism, and the development of three types of mineral deposits/systems that characterise this phase: orogenic gold deposits, orogenic base metal (Cu–Au–Zn–Pb–Ag) deposits, and Mississippi Valley (MVT) deposits.

9.3.1. Orogenic gold deposits

Orogenic gold systems typically form late in the evolution of convergent margin settings, during the main contractional orogenic stage and largely in fore-arc to back-arc settings in an accretionary margin (e.g. Kerrich and Wyman, 1990; Goldfarb et al., 2001, 2005; Groves and Bierlein, 2007; Bierlein et al., 2009). They are the result of stabilisation and cratonisation of an accretionary assembly, and are preserved in most Archean to Tertiary accretionary settings (Groves et al., 1998; Kerrich et al., 2000; Goldfarb et al., 2001; Wyman et al., 2003; Bierlein et al., 2009). The excellent preservation is largely due to the large crustal depth range (from 3 km down to at least 15 km) over which these systems form. Orogenic gold systems represent orogeny-wide release of fluids and magma, from previously metasomatised sub-continental lithospheric mantle or fertile lower crust to the near-surface environment as a response to: (1) thermal relaxation following accretionary suturing, and (2) metamorphic dehydration reactions, and (3) changes in the far-field stresses due to anomalous plate geometries and their interaction with continental margins or previously accreted terrains (Goldfarb et al., 1991; Kerrich and Wyman, 1994; Wyman et al., 1999; Groves and Bierlein, 2007; Bierlein et al., 2009; Willman et al., 2010). Metasomatised sub-continental lithospheric mantle, which can form during the earlier subduction stage, is now also recognised as a potential source of gold (Bierlein et al., 2006; Richards, 2009).

9.3.1.1. Orogenic gold deposits in the Yilgarn Craton. There is now evidence that the relative timing of gold mineralisation in the Yilgarn Craton is closely linked in time and space to a group of mantle-derived, felsic–mafic intrusions (Cassidy et al., 2002; Champion and Cassidy, 2007; Czarnota et al., 2010), and the contraction of post-volcanic siliciclastic basins (e.g. 'late basins': Krapež, 2007; Krapež et al., 2008a,b). Deposition of the 'late basins' directly precedes major transpressional deformation (Czarnota et al., 2010) and basin inversion. In the Yilgarn Craton, the timing of 'late basin' inversion, correlates broadly with a switch from mantle-derived (i.e. 'mafic' group granites) to crustal-derived (i.e. 'low-Ca' granites) felsic magmatism and overlaps with the range of robust age data for gold mineralisation, i.e. 2655–2650 Ma at several gold camps (Cassidy et al., 2002; Champion and Cassidy, 2007, 2008; Czarnota et al., 2010).

Orogenic gold systems are, in most cases, spatially linked to large-scale (>100 km in length), transcrustal deformation zones (Eisenlohr et al., 1989; Robert, 1989; Neumayr and Hagemann, 2002). In the Eastern Goldfields Superterrane, they are located within, or along the edges, of greenstone belts, and commonly mark the boundary between volcano-plutonic and metasedimentary subprovinces or terrains (Hagemann and Cassidy, 2000). In the western Victorian goldfields these deposits are located in the hanging wall of major transcrustal shear zones (Willman et al., 2010; Cayley et al., 2011b). These fault systems have a long-lasting history, with some of them, for example the Moyston Fault (Cayley and Taylor, 2001), possibly formed as early as the constructional stage of the accretionary margin setting (Miller et al., 2005), with subsequent multiple reactivation during orogenic and late-to post-orogenic events. In some cases the location of these structures is marked by gradients in radiogenic isotopes, such as Nd and Pb (Champion and Huston, 2016-in this issue).

Systematic isotopic mapping of the Yilgarn Craton, using Sm–Nd and Lu–Hf isotope systems (Champion and Cassidy, 2008; Mole et al., 2014; Champion and Huston, 2016-in this issue) suggests that although some orogenic gold districts are located along convergent margins, others can occur well inboard of the former margin. The central Murchison Domain is the second area (after the Eastern Goldfields Superterrane) with a high concentration of gold deposits in the Yilgarn Craton, and represents a zone of attenuated, juvenile crust hosting several large, layered, mafic intrusion complexes (Ivanic et al., 2012), and also the largest VHMS district in the Yilgarn (Golden Grove). Similar prospective juvenile "corridors" may be present within other Archean cratons, and can be identified by systematic isotopic mapping.

9.3.1.2. Other Australian examples. Although the Yilgarn Craton has historically been the largest orogenic gold producer in Australia, a number of other orogenic gold provinces are known in Australia, the most important being the ~445 Ma Victorian goldfield (Phillips et al., 2003, 2012), the ~1810–1795 Ma Tanami (Cross et al., 2005; Huston et al., 2007c; Cross, 2009) and Pine Creek (Ahmad et al., 2009) gold provinces, and the ~410 Ma Charters Towers district (Kreuzer, 2005). These gold provinces share many of the characteristics documented in the Eastern Goldfields Superterrane: the association with orogenesis, the large depth range of mineralisation, and the association with large-scale transcrustal deformation zones with long tectonic histories; however, there are also important differences. An interesting difference in Australia is that Archean (Eastern Goldfields) and Proterozoic (Tanami and Pine Creek) orogenic gold events accompanied coeval magmatism, yet despite an extensive geochronological database, such a relationship has not been established for ~445 Ma orogenic gold deposits in the Victorian goldfields (Phillips et al., 2003; Champion et al., 2009).

9.3.1.3. Metamorphic overprinting of orogenic gold deposits. There are several gold-rich deposits hosted in amphibolite facies metamorphic rocks that have experienced a complex history of deformation and metamorphism. These include VHMS and porphyry systems that were coeval with volcanism or plutonism during the initial stages of arc and back-arc basins, respectively. These were subsequently overprinted during synorogenic deformation by associated hydrothermal fluids that may have introduced new metals. Yeats et al. (1996) showed that there is a substantial age difference between VHMS style alteration and hypozonal orogenic gold mineralisation in the Mount Gibson gold deposits, which are located in the Murchison Domain (Youanmi Terrane) of the Yilgarn Craton. Recently, some gold deposits located in amphibolite to lower granulite facies Archean greenstone belts, such as Hemlo (Ontario, Canada), Challenger (South Australia), and Griffins Find (Western Australia), have been re-interpreted as metamorphosed (orogenic?) gold systems. Textural, petrogenetic and oxygen isotope evidence was used to propose a pre-metamorphic gold mineralisation event, followed by regional metamorphism including the modification of the originally (orogenic?) gold style mineralisation (Tomkins and Mavrogenes, 2002; Tomkins and Grundy, 2009; Tomkins et al., 2004; Hagemann et al., 2011).

9.3.2. Orogenic base metal deposits

Mineralisation in the Cobar area, which is located in north-central New South Wales, can be divided into two events. The first event involved formation of VHMS (e.g. Tritton) deposits within the Girilambone Group, which most likely formed as a back-arc basin to the early stage Macquarie Arc (Huston et al., 2016-in this issue). The Cobar Superbasin formed subsequently, and the second event, which involved epigenetic Cu–Au–Zn–Pb–Ag deposits, occurred during basin inversion. This event was characterised by metamorphism and local remobilisation of early VHMS mineralisation, and formation of orogenic base metal ("Cobar Style") mineralisation (e.g. CSA, New Cobar, New Occidental and Peak), quartz-vein hosted gold deposits (e.g. Gilgunnia Goldfield) and Mississippi Valley-Type deposits (e.g. Wonawinta) (Solomon and Groves, 2000).

Orogenic base metal mineralisation consists of syn-tectonic, remobilised structurally controlled deposits dominated by Cu–Au mineralisation (Lawrie and Hinman, 1998; Stegman, 2001). There is also a regional concentric metal zoning of deposits along the eastern edge of the Cobar trough, however, centred on a cluster of Au-rich deposits immediately south of the township of Cobar. Deposits become

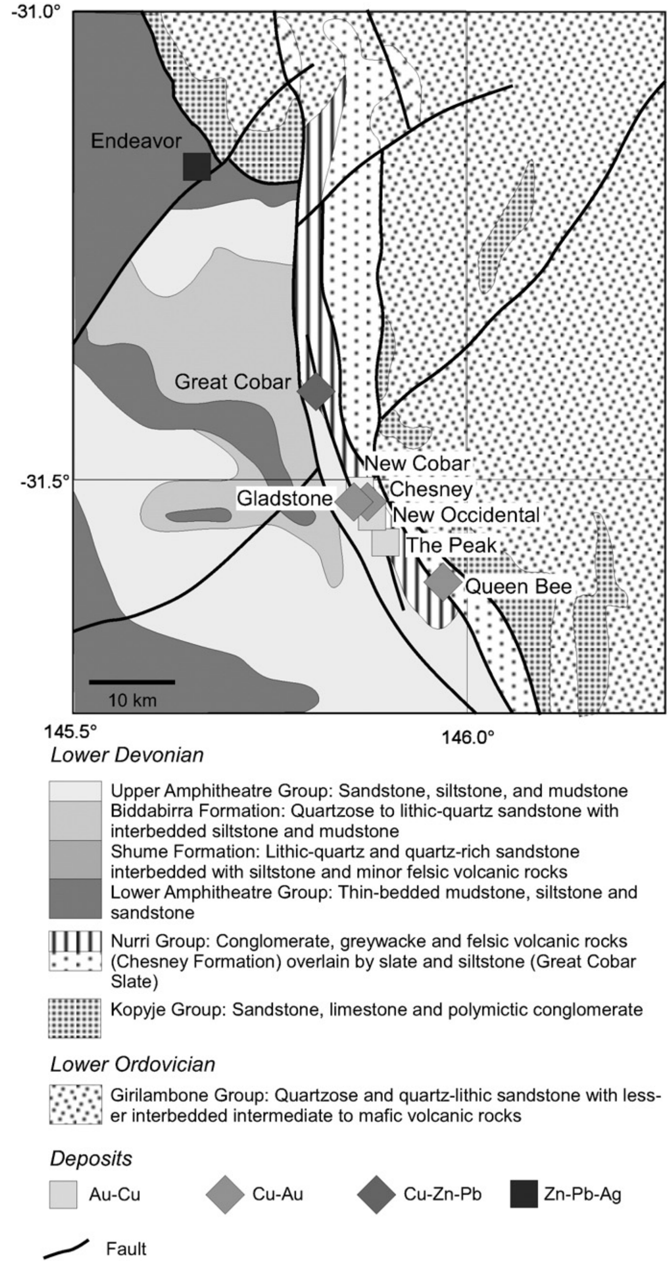

Fig. 12. Geology of the Cobar Basin showing the location and zonation of major deposits (modified after Stegman, 2001 and Murphy, 2007).

progressively more Au-poor and Cu-rich to the north and south and then ultimately become Pb–Zn–Ag-rich (Fig. 12: Stegman, 2001). Possible international examples of orogenic base metal deposits include Tsumeb in northern Namibia (Lombaard et al., 1986), vein-type Pb–Zn–Ag deposits (e.g. Woodcutters) in the Pine Creek region, Australia (Ahmad, 1998) and Pb–Zn–Ag deposits in the Coeur d'Alene mining district in northern Idaho, USA (Leach et al., 1988).

9.3.3. Mississippi Valley-type deposits

Globally, Mississippi Valley-type Zn–Pb deposits are temporally associated with orogenesis, although mineralisation can be well removed (100s of km) from the orogenic front. These deposits form when oxidised, low temperature basinal brines are gravitationally or tectonically driven away from an orogenic front into carbonate-bearing platformal successions (Garven, 1995; Solomon and Groves, 2000). Base metals are deposited when these fluids encounter a source of reduced sulphur (Fig. 7: Leach et al., 2005). Most MVT deposits are geologically young, with almost all deposits of Phanerozoic age (Leach et al., 2010). This may be because foreland basins, which these deposits can be associated with, have relatively poor preservation potential (Section 3.2) or because extensive carbonate reefs, the host rock for this type of deposit, are largely restricted to the Phanerozoic (Leach et al., 2010).

9.3.3.1. Australian examples. The most significant MVT deposits in Australia (cf., Solomon and Groves, 2000) are hosted by the Ordovician to Cretaceous Canning Basin (Lennard Shelf deposits and Admiral Bay deposit: Fig. 13) in Western Australia, with smaller deposits hosted in the Bonaparte Basin (WA: Sorby Hills, Fig. 13), the Cobar Superbasin (NSW: Wontawinta, Fig. 12) and the Eldon Group (Tasmania: Oceana). With the exception of the Lennard Shelf deposits, which have a sphalerite Rb–Sr age of ~357 Ma (Christensen et al., 1995a,b), the age of these deposits are poorly constrained. Based on relationships to cement chronology, McCracken et al. (1996) inferred an age for the Admiral Bay deposit of between 425 Ma and 410 Ma. On a national scale, the Lennard Shelf age is slightly younger than the Brewer Movement of the Alice Springs Orogeny in central Australia but similar to the Kanimblan Orogeny in eastern Australia (Fig. 14). Although the age of the Admiral Bay deposit is poorly constrained, it corresponds to the Bindian Orogeny in eastern Australia (Fig. 14: Huston et al., 2012). These data are consistent with the concept (e.g. Leach et al., 2001, 2005) that MVT deposits are the far-field responses to orogenesis, which drives low-temperature and H_2S-poor basinal fluids into platformal successions. Zinc and Pb are then deposited when H_2S, either derived from a separate fluid or produced by organic or inorganic sulphate reduction, is added to the ore fluid (Anderson, 1975; Beales, 1975).

9.4. Final stage — post-subduction extension

Following orogenesis associated with terrane accretion and/or a shallowing of subduction, convergent margins commonly go into extension, due to post-collisional relaxation related to orogenic collapse, the re-initiation of subduction (Collins and Richards, 2008; Richards, 2009), or other processes (Section 9.5). This change in tectonics also produces a change in metallogeny, with alkaline porphyry Cu–Au and related epithermal, intrusion-related W–Sn–Mo–Au (including porphyry Mo), skarn Zn–Pb–Ag and IOCG deposits becoming important in the extensional phase. Although these deposits all form during post-subduction extension, it appears that they may be in response to different drivers.

9.4.1. Alkaline porphyry Cu–Au deposits

As indicated by Richards (1995, 2009) and Müller and Groves (1993), porphyry Cu–Au deposits associated with intermediate alkaline volcanic rocks are now thought to have formed during post-subduction extension, with 440–435 Ma deposits in the Macquarie "Arc" being the best Australian examples (Crawford et al., 2007; Cooke et al., 2007:

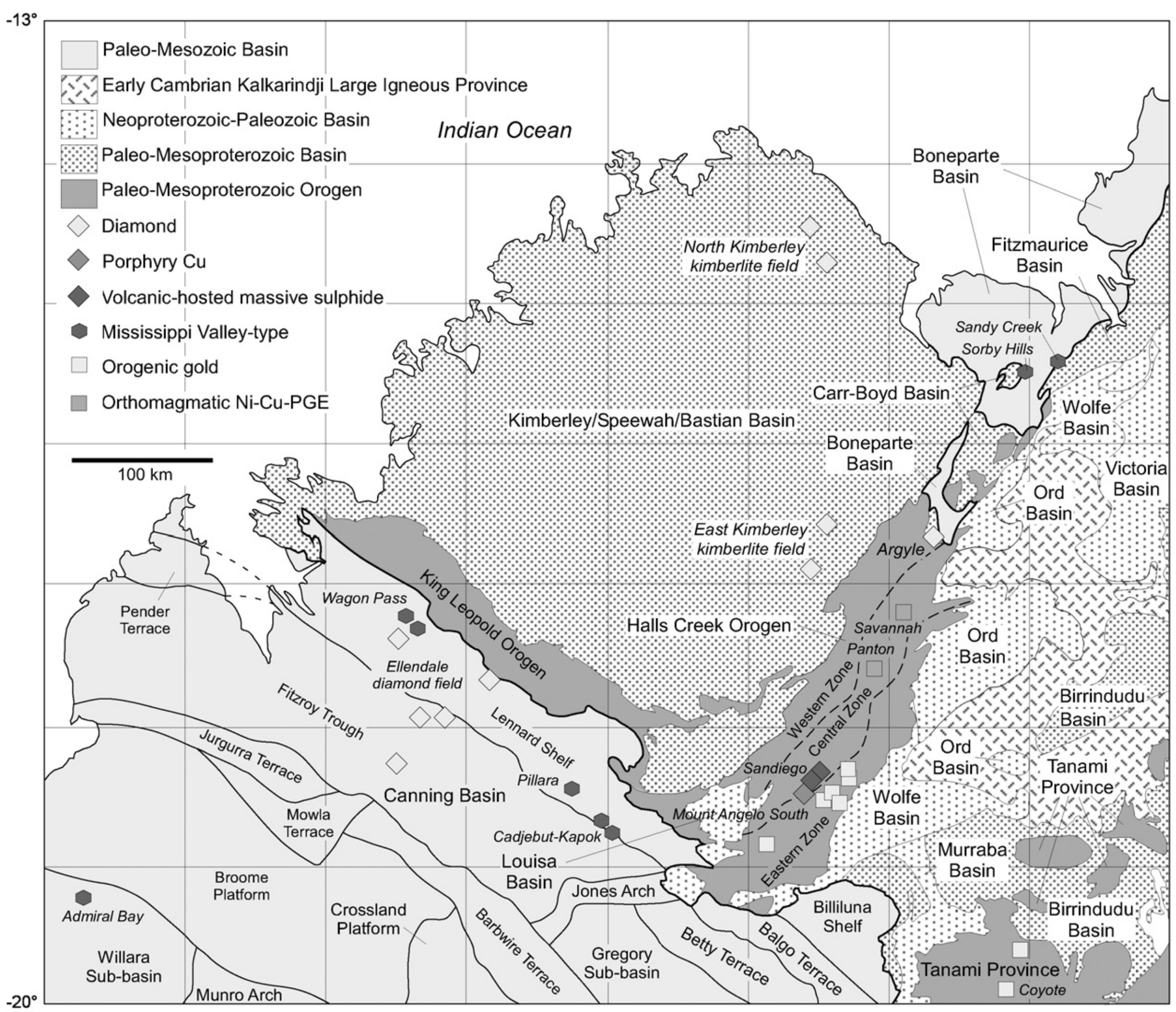

Fig. 13. Geology of the Kimberley Province and surrounding basins showing the location of major Mississippi Valley-type deposits (modified after Ahmad and Scrimgeour, 2013 and Geological Survey of Western Australia GeoVIEW.WA (www.dmp.wa.gov.au/7113.aspx; accessed June 2015)).

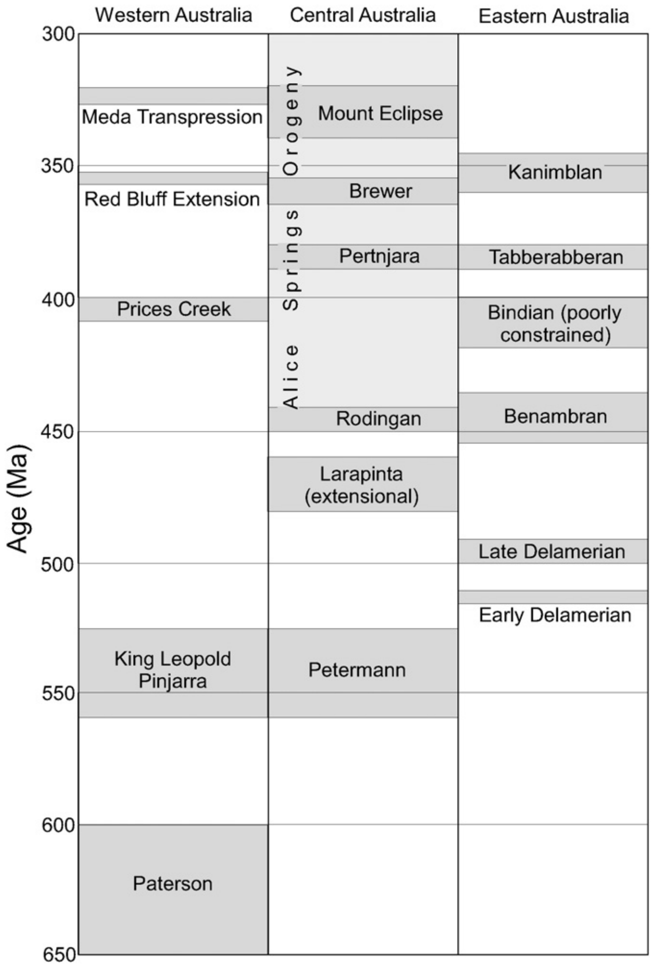

Fig. 14. Correlation of major late Neoproterozoic and Phanerozoic orogenies across Australia (modified after Huston et al., 2012).

Fig. 15). Other examples around the world include those in the Yulong belt in northern China (Hou et al., 2003) and deposits in the North American Cordillera, particularly in British Columbia (Barrie, 1993). The alkaline melts are thought to have been produced by the remelting of subduction-modified mantle triggered by delamination of mantle lithosphere, asthenosphere upwelling and/or crustal extension (Richards, 2009). These post-collisional deposits can overprint the earlier-formed magmatic arcs: the alkaline post-subduction phases of the Macquarie arc are built upon older (470–450 Ma) calc-alkaline volcanic rocks that were deposited as the arc grew (Crawford et al., 2007). As discussed above, these older volcanic rocks contain ~450 Ma and older deposits associated with calc-alkaline magmatism.

Porphyry Cu–Au deposits in the Cadia valley (Fig. 15) are the largest resource of Au and Cu in eastern Australia, with global pre-mining resources of 7.5 Mt Cu and 1150 t Au. Although overall these deposits are low grade (2800 Mt grading 0.27% Cu and 0.41 g/t Au: Wood, 2012), local zones, such as Ridgeway (78 Mt grading 0.67% Cu and 2.0 g/t Au: Cooke et al., 2007), can be high grade. The Cadia deposits and the temporally- and genetically-related Northparkes porphyry and epithermal deposits formed late during the evolution of the Macquarie Arc, according to Crawford et al. (2007) as the consequence of post-subduction extension (Section 9.5).

9.4.2. Intrusion-related W–Sn–Mo–Au deposits

One of the central tenets of intrusion-related mineralisation has been the recognition of the role played by the variations in granite properties, such as oxidation state (Burnham and Ohmoto, 1980; Ishihara, 1981). Blevin et al. (1996) and Blevin (2004), and others (Thompson et al., 1999) have documented the relationship between the degree of compositional evolution (commonly expressed as degree of fractionation) and the oxidation state of the associated intrusives and commodity-types in granite-related mineralisation systems (e.g. Sn, W, Mo, Cu, Cu–Au). These parameters relate to the behaviour of ore metals within the magma, that is whether they are compatible or incompatible, and whether or not concentrations of such metals are highest in compositionally-unevolved or compositionally-evolved magma compositions. As noted by Blevin et al. (1996), these and other metallogenically important parameters (e.g. S, Cl and volatile contents of the magma) are all, to at least some degree, a function of the source components of the magma, and so are all indirectly related. These simple but powerful relationships (Fig. 16) can be utilised predictively; for example, Champion and Mackenzie (1994) demonstrated the very strong correlation between Sn occurrences in the north Queensland region and reduced, strongly fractionated granites of both varying ages and, importantly, of varying granite types (I- or S-type). The latter relationship holds for all tin deposits in eastern Australia (Blevin et al., 1996), although the largest of these deposits, Renison in western Tasmania, is associated with S-type granites. The important implication of this is that Sn (±W) deposits, accordingly, can form in any geodynamic environment that favours generation of both reduced and strongly differentiated granite magmas — empirically this appears to be mostly within extensional continental environments, such as back-arc settings, post-collisional/post-tectonic settings, or even continental rift settings. It is not entirely clear why these magmas are produced in such environments. Part of the answer may reflect higher crustal geotherms in these regions, which would promote higher temperature crustal melting, promoting lower viscosity melts with potential for solidus-lowering volatiles such as F and B. Crustal melting is clearly an important factor and some protoliths are more fertile, e.g., metasediments—one of the reasons why larger tin deposits are associated with S-type granites.

Similar implications, as for tectonic environments and Sn-mineralisation, hold for W, although oxidation state is less critical (Blevin et al., 1996). As noted previously, most magmatic arc environments are characterised by more oxidised and primitive magmas making them unfavourable for such Sn and W mineralisation, though inherently more favourable for porphyry Cu–Au and Cu–Mo deposits (Fig. 16). The association of Mo deposits with oxidised magmas covering a broad range of compositions, however, means that such deposits will not be confined to arc environments, but can occur in back-arc, post-subduction and rift environments (e.g. Climax-type porphyry molybdenum deposits: Luddington and Plumlee, 2009).

Work in the last decade has resulted in the identification of a class of Au-deposits associated with felsic intrusions, consolidating a variety of previously recognised intrusion-related mineralisation under the one mineral system, known as intrusion-related gold (IRG: Thompson et al., 1999). Such systems are often characterised by a distinctive Au-Mo-W-Bi association (e.g. Blevin, 2004). Styles included within this mineral system are broad; they range from deposits proximal to granites (greisen, disseminated gold and skarns) to those more distal, and more controversial, deposits (breccias, and vein systems), where the relationship with granites is equivocal. Although the models for IRGs are largely influenced by North American deposits, a number of well-studied Australia examples exist, including Timbarra (Mustard, 2001), Kidston (Baker and Andrew, 1991), and Red Dome (Blevin, 2004).

Most IRGs are clearly associated with evolved felsic intrusives (Fig. 16): this is particularly true for Australian deposits (Timbarra: Mustard, 2001), but less so for North American deposits (Thompson et al., 1999; Lang et al., 2000). Most controversy concerns the relative oxidation state of the intrusives. Early IRG models emphasised the mildly to moderately reduced nature of the granites, based on aeromagnetic signatures, whole rock Fe_2O_3/FeO ratios, magnetic susceptibility measurements, and the lack of modal primary magnetite (Thompson et al., 1999). This, however, contrasts with reported data for Australian IRGs

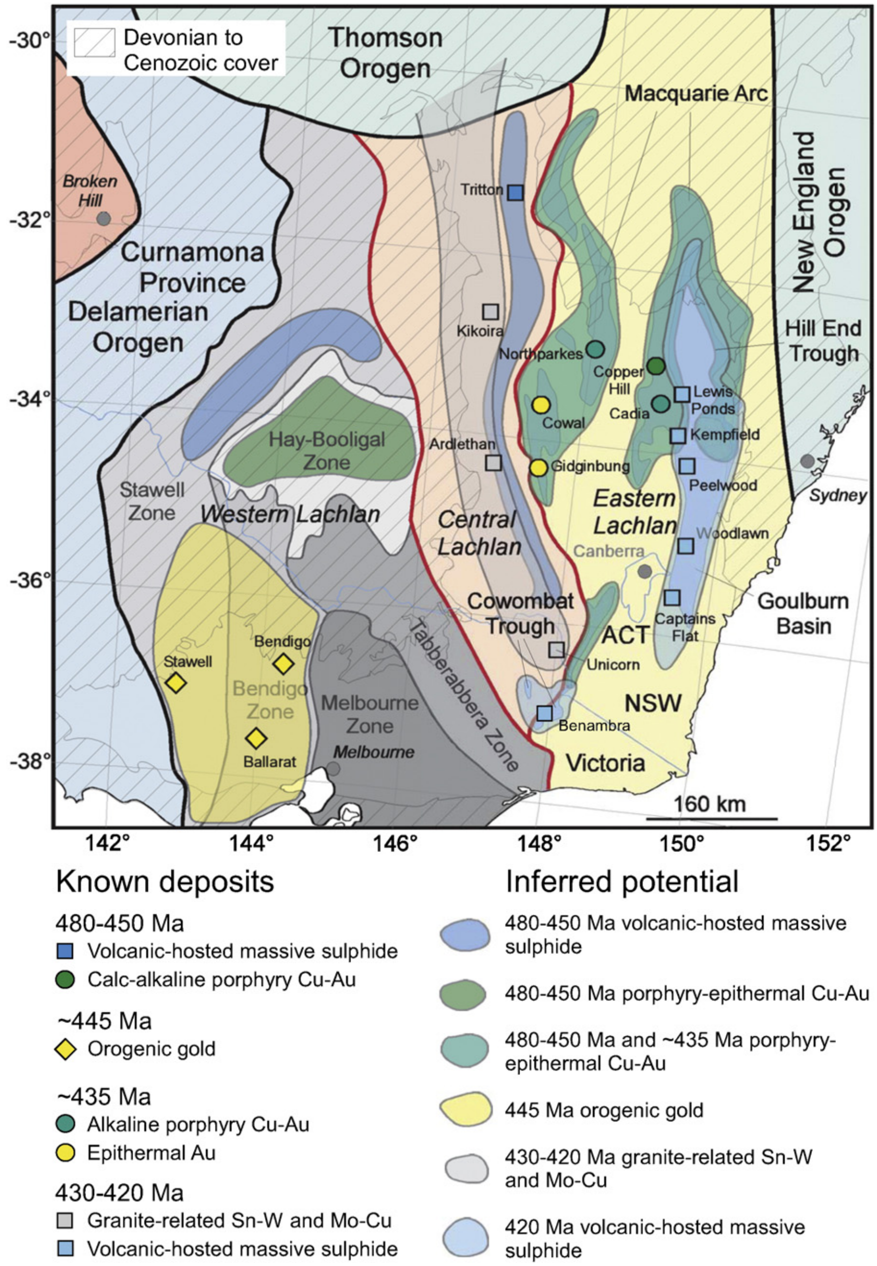

Fig. 15. Known and inferred mineral potential associated with the Benambran event in southeastern Australia. Geological provinces from Glen (2013); overlay of Devonian to Cenozoic cover is from Geoscience Australia databases.

which are commonly weakly- to moderately-oxidised, largely just above the FMQ buffer (Blevin, 2004). The presence of titanite (if primary) in at least some North American granites associated with IRGs (e.g. Lang et al., 2000) is also more consistent with a weakly oxidised (though magnetite-free) nature. Accordingly, we have followed the redox state interpretation of (Blevin, 2004; Fig. 16), an interpretation now also largely supported by Baker et al. (2005). Unfortunately, the characteristics of granites associated with IRGs (moderately- to strongly-evolved compositions that are weakly magnetic) are shared by many intrusives, in a variety of tectonic settings (arc to back-arc settings, post-orogenic settings). Often such deposits have a spatial association with areas of tin mineralisation (north Queensland, New England and Tasmania) in eastern Australia, suggesting continental extensional settings are important.

In summary, intrusion-related W–Sn–Mo–Au deposits may occur in a variety of tectonic environments largely characterised by extension. In contrast to porphyry-style mineralisation, they are commonly not found within environments directly-related to arcs. Regions of intrusion-related W–Sn–Mo–Au mineralisation are separated spatially and/or temporally from regions of porphyry-style mineralisation. This was also recognised by Thompson et al. (1999, their Fig. 5) who suggested porphyry style mineralisation was associated with arc environments versus back-arc or non-arc crustal environments for intrusion-related W–Sn–Mo–Au mineralisation. This antipathetic association suggests regions with potential for intrusion-related W–Sn–Mo–Au mineralisation may be recognisable by evolved isotopic signatures. As discussed by Champion and Huston (2016-in this issue), however, felsic magmatic rocks associated with such mineralisation, have a demonstrably wide range of isotopic signatures (from juvenile to evolved), and metallogenic terranes with such mineralisation cannot easily be identified from isotopic data. This is not unexpected given that geodynamic environments evolve with time: arcs associated with porphyry mineralisation evolve into backarcs or post-collisional settings associated with intrusion-related W–Sn–Mo–Au mineralisation.

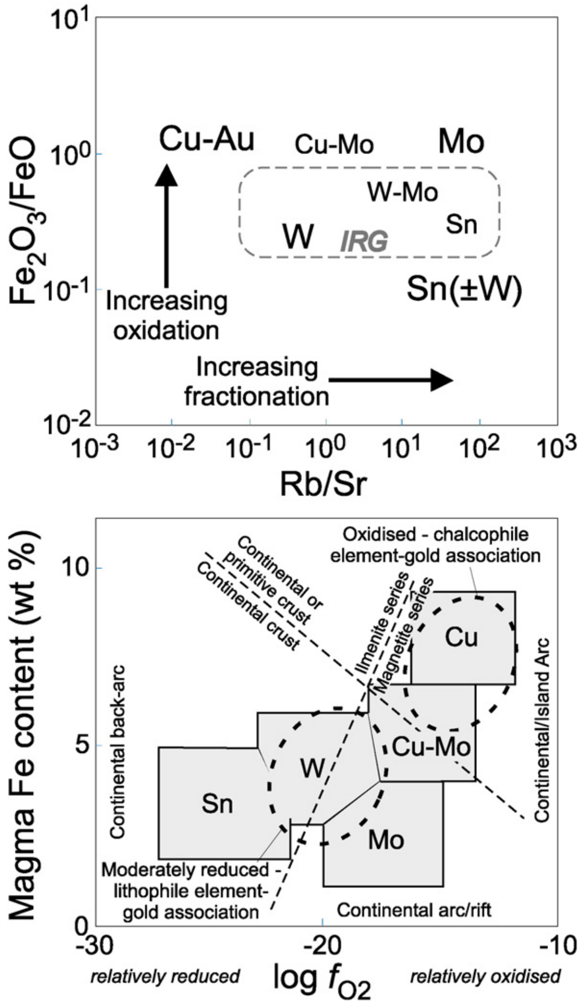

Fig. 16. (A). Rb/Sr ratio versus Fe_2O_3/FeO ratio plot of Blevin et al. (1996). The plot illustrates the relationship between the degree of oxidation and compositional evolution of the magma (based on whole-rock compositions) and the dominant commodities in related mineralisation. Intrusion related gold deposit (IRGD) field from Blevin (2004). (B). Plot of oxygen fugacity versus amount of total Fe in the magma overlaid on tectonic fields. Plot modified after Thompson et al. (1999) and Lang et al. (2000).

More importantly these switches highlight the role played by tectonic setting. Finally, it should be emphasised that the best indicators for regions with potential intrusion-related W–Sn–Mo–Au mineralisation are either the granites themselves (as recognised by Blevin and Chappell, 1992), and/or known mineralisation. For example, regions with potential for intrusion-related gold mineralisation may be recognised by the presence of tin mineralisation.

9.4.3. Iron oxide–copper–gold deposits

The geodynamic settings of IOCG deposits (Hitzman et al., 1992) are less well understood than many other mineral deposit types, partly due to their relative youth in terms of recognition as a deposit class and also due to their relative antiquity in regard to the age of formation of many of the largest deposits. Nevertheless, a working hypothesis based mainly on the Australian Proterozoic examples is that IOCG deposits formed during post-subduction extension within distal back-arc to intracontinental settings, as described below.

Iron oxide–copper–gold deposits are a diverse family characterised by the following features (Hitzman et al., 1992; Williams et al., 2005; Groves et al., 2010; Barton, 2014): (1) Cu with or without Au, as the major economic metals; (2) hydrothermal epigenetic ore styles and strong structural controls, (3) abundant magnetite and/or hematite, (4) Fe oxides with Fe/Ti greater than those in most igneous rocks, and (5) no clear proximal associations with igneous intrusions as, for example, displayed by porphyry and skarn ore deposits. Additionally, IOCG deposits occur in crustal settings with very extensive and commonly pervasive alkali metasomatism, and many are enriched in a distinctive, geochemically diverse suite of minor elements including U, REE, F, P, Mo, Ag, Ba, Co, Ni and As (Williams et al., 2005). At the Olympic Dam deposit, U is part of the economic ore suite. Although not closely related to igneous intrusions, many IOCG deposits display a broad space-time association with batholithic granitoids (Williams et al., 2005). Iron oxide-copper-gold deposits formed from the Archean (e.g. Carajás district, Brazil, ~2.7 Ga; Moreto et al., 2013), through the Proterozoic (e.g. Olympic Dam deposit, Gawler Craton, Australia), to at least as recently as the Mesozoic (e.g. Candelaria deposit, Chile, ~115 Ma; Mathur et al., 2002).

The geodynamic settings of IOCG deposits have been widely debated (e.g. Williams et al., 2005; Groves et al., 2010). An intracontinental anorogenic setting has often been cited, with mantle plume-driven melting of subcontinental lithospheric mantle (e.g. Groves et al., 2010). Hayward and Skirrow (2010) reviewed tectonic and geodynamic models for the Gawler Craton in southern Australia, and proposed a distal continental retro-arc environment where much earlier subduction-related processes (possibly at ~1850 Ma) led to metasomatism of the upper mantle. Melts derived from this enriched mantle, driven by a mantle plume or alternatively by removal of lithospheric mantle (Skirrow, 2010), resulted in extensive crustal melting and production of high-temperature A- and I-type magmas associated with K-rich mafic melts between ~1595 Ma and ~1575 Ma. These igneous rocks are temporally and spatially linked to IOCG deposits in the Olympic IOCG Province. In contrast to anorogenic models, however, this magmatism and associated IOCG hydrothermal systems are proposed to have developed during or after a major contractional tectonic event that switched to extension (Skirrow, 2010; Hayward and Skirrow, 2010).

Crustal domain boundary zones initiated during earlier orogenic events are believed to form part of the crustal-scale magma and fluid pathways for major IOCG systems (e.g. beneath the Olympic Dam deposit; Lyons and Goleby, 2005; Heinson et al., 2006). Groves et al. (2010) extended this concept to other major IOCG deposits globally.

There is growing consensus that IOCG deposits can be the product of mixing of two distinct fluids: (1) an oxidised fluid (e.g. evolved meteoric/ground waters), and (2) deep-sourced high-temperature brines (magmatic-hydrothermal fluids and/or fluids reacted with metamorphic rocks). In many IOCG systems, there is also evidence of gas-rich fluids during ore formation (e.g. CO_2-bearing: Williams et al., 2005, and the references therein). The sources of Cu, Au, S, Cl and CO_2 may be either coeval magmas (felsic and/or mafic) or sedimentary and igneous rocks that were leached by the ore fluids, as marked by the presence of Na–Ca regional alteration zones (Oreskes and Einaudi, 1992; Johnson and McCulloch, 1995; Haynes et al., 1995; Williams et al., 2005; Oliver et al., 2004; Skirrow et al., 2007). Uranium and REE were most likely leached from granitoid or felsic volcanic rocks (Hitzman and Valenta, 2005).

9.4.3.1. Australian examples. There are two major IOCG provinces in Australia of global significance: the Olympic IOCG Province (Fig. 17A) near the eastern margin of the Gawler Craton in South Australia, and the Cloncurry province (Fig. 17B) in the eastern Mount Isa Province of northwest Queensland. In addition, there are several other metallogenic provinces that contain smaller IOCG deposits, including the Tennant Creek district (Northern Territory) and the Curnamona Province (South Australia and New South Wales). The Olympic IOCG Province is defined by the distribution of known early Mesoproterozoic IOCG ± U mineralisation and alteration, and encompasses three known districts (from north to south): Mount Woods Inlier which hosts the Prominent Hill deposit; Olympic Dam district hosting the Olympic Dam, Carrapateena and Wirrda Well deposits; and the historic Moonta-Wallaroo Cu–Au mining district. The Olympic Dam deposit is currently the world's fourth largest Cu resource, fifth largest Au resource and

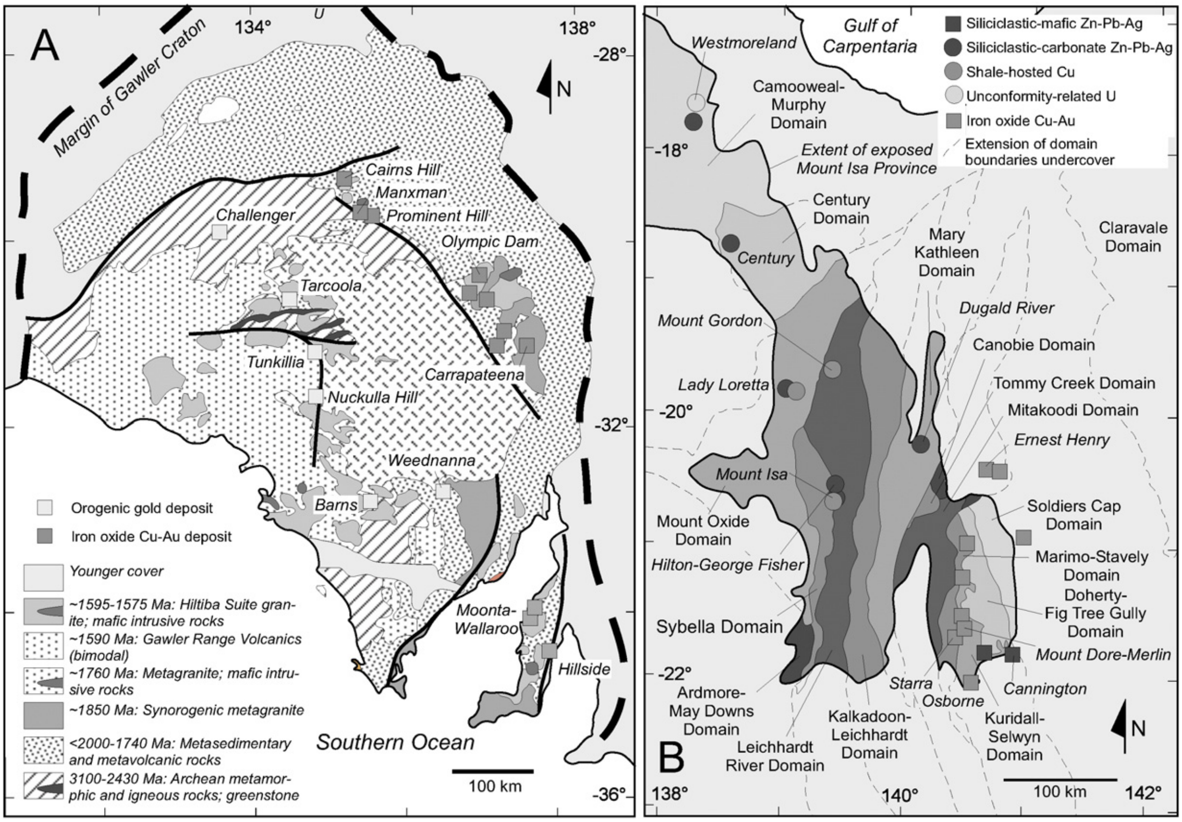

Fig. 17. Geology and mineral deposits of major IOCG provinces of Australia: (A) Olympic province, and (B) Cloncurry province (modified after Hayward and Skirrow, 2010 and Geological Survey of Queensland, 2011).

the world's largest U resource by far, with all resources contained in a single deposit of less than 25 km^2 (BHP Billiton, 2012).

The Cloncurry district in the eastern Mount Isa Province hosts the large Ernest Henry IOCG deposit as well as several smaller IOCG and affiliated deposits including the Osborne, Eloise, Mount Elliot, Mount Dore and Monakoff deposits (Williams et al., 2005). IOCG mineralisation developed in the Mesoproterozoic, during two periods, around 1590 Ma and 1500–1530 Ma (Williams et al., 2005 and the references therein). The breccia-hosted Ernest Henry deposit differs from the generally hematite-rich IOCG deposits in the Olympic IOCG Province in that the major iron oxide is magnetite, with associated hydrothermal K-feldspar, biotite and carbonate (Williams et al., 2005). Notwithstanding this difference, the magnetite-rich Ernest Henry deposit and hematite-rich IOCG deposits of the Olympic IOCG Province constitute a spectrum of deposit styles within the IOCG family.

9.4.4. Enriched BIF-hosted iron ore deposits

The enrichment of BIF to form enriched, low- and high-grade BIF-hosted iron ore deposits occurs during post-subduction extension following contractional deformation (Hagemann et al., 2016-in this issue). This type of iron ore system has produced the world's largest and highest grade iron ore districts (e.g. Carajas: Figueiredo e Silva et al., 2008; Hamersley: Thorne et al., 2008; Iron Quadrangle: Rosière et al., 2008) and deposits (e.g. Mount Whaleback: Brown et al., 2004). Hydrothermal fluid flow and mineralisation are controlled by km-scale normal and strike slip fault systems (Southern Batter Fault at Tom Price and Carajás Shear Zone at the Serra Norte deposits at Carajás, respectively), which allow large volumes of ascending and descending hydrothermal fluids to circulate during Archean or Proterozoic orogenic or early extensional events. Structures are also (passively) accessed via downward flowing supergene fluids during the Cenozoic.

At the depositional site, the transformation of BIF to low- and high-grade iron ore is controlled by: (1) structural permeability, (2) hypogene alteration caused by ascending deep fluids (largely magmatic or basinal brines), plus descending meteoric water, and (3) supergene enrichment via weathering processes. Hematite- and magnetite-based iron ores include two groups of ore assemblages: (1) a combination of microplaty hematite with little or no goethite, martite–goethite or specular hematite, and (2) magnetite, magnetite–martite, magnetite–specular hematite and magnetite–amphibole. Goethite ores, with variable amounts of hematite and magnetite, are mainly encountered in the weathering zone.

In most large deposits, up to three major hypogene stages and one supergene ore stage are present (Hagemann et al., 2016-in this issue): (1) silica leaching and formation of magnetite and local carbonate; (2) oxidation of magnetite to hematite (martitisation), further dissolution of quartz and formation of carbonate, (3) further martitisation, replacement of Fe silicates by hematite and new microplaty hematite formation and dissolution of carbonates; and (4) replacement of magnetite and any remaining carbonate by goethite, martitisation of magnetite, and formation of fibrous quartz and clay minerals.

Hypogene alteration of BIF and surrounding country rocks is characterised by Hagemann et al. (2016-in this issue) as: (1) changes in the oxide mineralogy and textures, (2) development of distinct vertical and lateral distal, intermediate and proximal alteration zones defined by distinct oxide-silicate-carbonate assemblages, and (3) mass loss reactions such as de-silicification and de-carbonatisation, which significantly increase the porosity of high-grade iron ore, or lead to volume reduction by textural collapse or layer-compaction. Whereas stage 1 carbonate may be diagenetic and sourced from BIF-contemporaneous seawater, carbonates in ore stages 2 and 3 are sourced from external fluids with respect to BIF. In the case of Lake Superior-type basin-related deposits, carbon is interpreted to be derived from underlying carbonate succession, whereas in the case of Algoma-type granite-greenstone belt deposits, carbonate is interpreted to be of magmatic origin, either scavenged from mafic rocks or sourced from granitic country rocks.

Hypogene and supergene fluids of varying composition are paramount for the upgrade of BIF to high-grade iron ore (Angerer et al., 2014; Hagemann et al., 2016-in this issue). Processes causing this upgrade require enormous amounts of: (1) warm, silica-undersaturated and alkaline fluids necessary to dissolve quartz in BIF, (2) oxidised fluids that cause the oxidation of magnetite to hematite, (3) alkalic fluids that form widespread metasomatic carbonate, (4) carbonate-undersaturated fluids that later dissolve the diagenetic and metasomatic carbonates, and (5) oxidised fluids to form hematite species in the hypogene- and supergene-enriched zone and hydroxides in the supergene zone.

The extensional tectonic setting includes two discrete end-member models for the formation of hypogene low- and high-grade BIF-hosted iron ore (Hagemann et al., 2016-in this issue). In deposits hosted by granite-greenstone belts, Algoma-type BIF is upgraded by early magmatic (± metamorphic) fluids and late meteoric water controlled by strike-slip fault zones. In rift-related basins and passive margins, Hamersely-type BIF is upgraded by basinal (± evaporitic) brines and late meteoric water focused along normal fault zones. One variation of the latter model is the metamorphosed rift basin model where BIF is significantly metamorphosed and deformed during distinct orogenic events (e.g. deposits in the Quadrilátero Ferrífero in Brazil and Simandou Range in Guinea). A second variation is the Rapitan-type model where early BIF mineralisation in a reduced, glacier-controlled environment and subsequent very low-grade metamorphism is responsible for the bulk of the iron mineralisation. Typically, it is during the orogenic event(s) that the upgrade of BIF to low- and high-grade hypogene iron takes place.

9.4.4.1. Australian examples. The Hamersley Basin, a sedimentary basin in the Pilbara Craton produces the vast majority of iron ore in Australia. This basin, which is though to have formed as a passive margin during the late Neoarchean to very earliest Paleoproterozoic, contains several banded iron formations that range in age from ~2600 Ma to ~2540 Ma (Barley et al., 1997; Trendall et al., 1998). The iron ore deposits, which are hosted by banded iron formation along the deformed southern margin of the Hamersley Basin (Barley et al., 1999), are thought to have been upgraded, as described above at ~2009 Ma, possibly during or after the Opthalmia Orogeny (Powell et al., 1999; Müller et al., 2005). This iron-ore province is one of two major global provinces, and it dwarfs all other iron-ore districts/provinces in Australia.

Outside of the Pilbara iron-ore province, other Australian iron-ore deposits are commonly magnetite-rich and include unenriched banded iron-formation (Tallering Peak and Mount Gibson in the Yilgarn Craton), enriched or hydrothermally altered iron formation (as described above: Koolyanobbing, Yilgarn Craton; Middleback Ranges (Iron Duke, Iron Knight and Iron Duchess), Gawler Craton), and hydrothermal (Savage River, western Tasmania) deposits (Clout, 2003). The other economically significant source of iron-ore are the channel iron deposits (Clout, 2003), that occur within paleochannels that have cut into a Paleogene paleosurface developed in the Hamersely Basin. Channel iron deposits are thought to have formed during the Miocene as the consequence of deep weathering, ferrugenisation and erosion of susceptible rocks (iron-ore depostis, iron formation and mafic volcanic rocks) followed by deposition into paleochannels (Morris and Ramanaidou, 2007).

9.5. The Lachlan tectono-metallogenic system

As an example of how spatial and temporal distributions of mineral deposits can be understood in the context of tectono-metallogenic systems, we consider the metallogenesis of southeastern Australia between 480 Ma and 410 Ma, which produced major orogenic gold and porphyry Cu–Au metallogenic provinces (Figs. 15 and 18). Previous data have suggested that alkaline porphyry Cu–Au mineralisation overlapped with the major phase of orogenic gold mineralisation in the Victorian Goldfields at 440–435 Ma (Perkins et al., 1995; Crawford et al., 2007; Bierlein et al., 2001) presenting an apparent conundrum of coeval, yet widely geographically separated, porphyry and orogenic mineral systems. Recognition that the 440–435 Ma porphyry deposits are post-collisional (Crawford et al., 2007) and documentation by Phillips et al. (2012) that most Victorian orogenic gold deposits formed during the Benambran Orogeny at ~445 Ma (VandenBerg et al., 2000), however, removes this conundrum. The new interpretations suggest that these deposits were not coeval, but that the post-subduction porphyry Cu–Au mineralisation post-dates the orogenic Au mineralisation by several million years. Importantly, this change in metallogenic patterns can be directly linked to a major tectonic event that perturbed

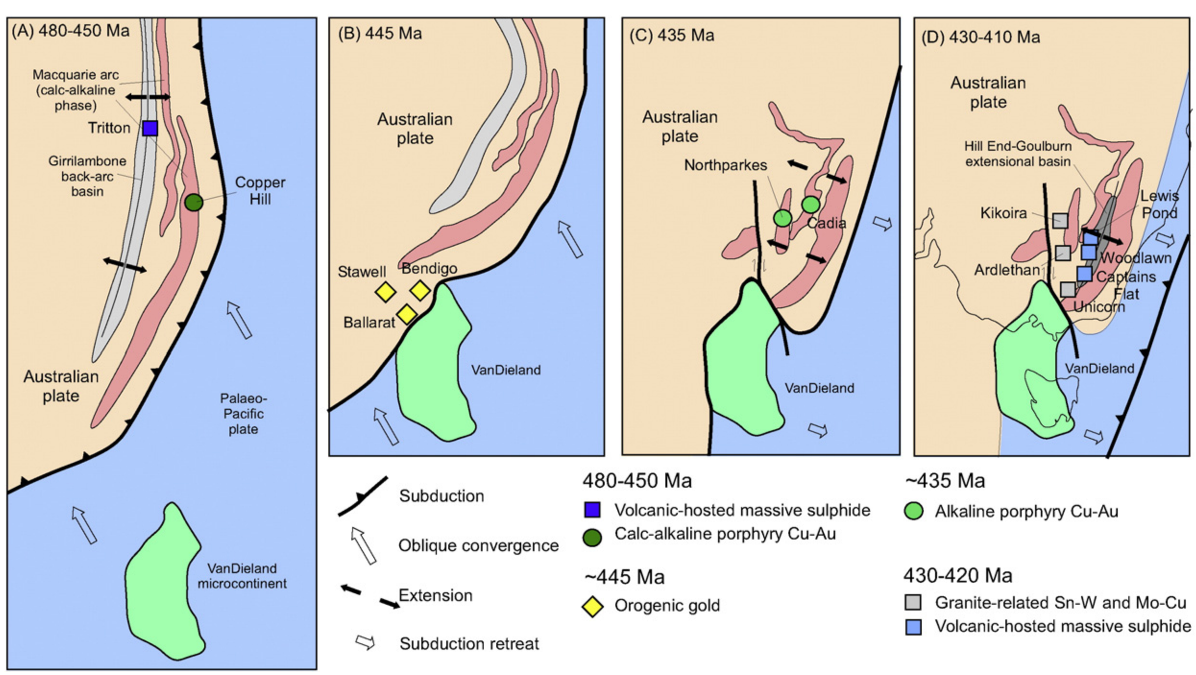

Fig. 18. Inferred evolution of the Lachlan tectono-metallogenic system (adapted from Moresi et al., 2014): (A) oblique subduction and approach of VanDieland (480–450 Ma), (B) accretion of VanDieland and the Benambran Orogeny (445 Ma), (C) formation of Lachlan orocline (~435 Ma), and (D) subduction retreat and re-establishment of oblique subduction (430–410 Ma). The locations of significant deposits are shown as they formed during the evolution of the Lachlan tectono-metallogenic system.

convergence along the eastern margin of Australia at the very end of the Ordovician, during the Benambran Orogeny.

At that time, the tectonic evolution of eastern Australia was complex, leading to a variety of tectonic models being proposed (see review by Champion et al., 2009). The coverage of southeastern Australia by high quality aeromagnetic data and new methods of processing such data, along with new geological data from modern regional geological mapping programmes and insights from geodynamic modelling (Moresi et al., 2014), however, have led Cayley (2012) to propose a tectonic model in which the accretion of the exotic VanDieland (or Taswegia: Gibson et al., 2011) continental fragment, which consists of most of Tasmania and the Selwyn Block (Cayley et al., 2002), into the Macquarie Arc subduction zone caused the formation of an orocline. The development of the Lachlan Orocline involved parts of the subduction zone adjacent to the site of microcontinent accretion rolling-back asymmetrically throughout the Silurian, eventually wrapping and translating parts of the Lachlan Orogen around the outer margins of VanDieland. A west-dipping subduction zone re-established farther outboard in the Early Devonian (Fig. 18). In the following discussion, the metallogeny of southeastern Australia in discussed in the context of this tectonic event, which is termed the Lachlan tectono-metallogenic event. The resulting metallogenic model is the used to infer zones of higher metallogenic potential that are the consequence of the Lachlan event (Fig. 15).

The evolution of this tectonic system explains the known spatial and temporal distribution of mineral deposits (systems) from 480 Ma to 410 Ma in eastern Australia. Prior to the impact of VanDieland at ~445 Ma, convergence along the eastern margin of mainland Australia was characterised by the presence of an arc-back-arc system on the over-riding Australian plate. This system developed in response to the oblique (from present-day southeast) subduction of a proto- Pacific plate that contained the microplate VanDieland. During this period of subduction, VHMS deposits formed in back-arc basins in north-central New South Wales (e.g. Tritton: Fig. 18A) and northern Queensland (e.g. Thalanga and Balcooma; not shown in Fig. 18A), and calc-alkalic porphyry Cu–Au deposits formed during the early stages of the Macquarie Arc (e.g. Copper Hill and Marsden: Fig. 18A). The impact of VanDieland (Fig. 18B) caused orogenesis, the Benambran Orogeny, in the immediate hinterland (e.g. central Victoria), with the associated development of an orogenic gold mineral system (Victorian goldfields: Squire and Miller, 2003) that did not extend far to the north. Collision also initiated extension directly to the north along strike, leading to the formation of an orocline as mainland Australia partly enveloped VanDieland (Fig. 14B–C). With the mainland Australian plate to the north in extension (Moresi et al., 2014: Fig. 18C) low degree partial melting of the underlying mantle produced alkaline magmatism and porphyry Cu-Au mineral systems in the Macquarie Arc (e.g. Cadia and Northparkes). Subduction continued to roll-back into the Late Silurian, causing extension and forming post-orogenic granite-related Sn and Mo deposits (430–420 Ma) and then VHMS deposits (~420 Ma: Fig. 18D), after which steady-state subduction was re-established.

The Lachlan tectono-metallogenic system evolved over a period of 60 million years, producing a range of deposit types, and a large proportion of eastern Australia's mineral wealth. Moreover, the development of a tectono-metallogenic system model for the Lachlan tectonic event can be used predictively to assess potential mineral system plays in southeastern Australia. For example, the geometry of the tectonic system suggests that 480–450 Ma porphyry Cu–Au and VHMS mineral systems could have been present along much of the Tasmanides, including areas covered by Murray Basin and other Cenozoic basins (Fig. 15), but that ~445 Ma orogenic gold systems would have only developed in and adjacent to the hinterland of VanDieland, and ~435 Ma alkalic porphyry Cu–Au systems would have developed only in extensional zones north of VanDieland. Given their association with post-subduction extension (Section 9.4.3), IOCG deposits could be present north of the VanDieland indentor.

10. Intraplate tectono-metallogenic system

Although most mineral systems appear to operate along or near plate margins either as a consequence of tectonic divergence or convergence, some mineral systems are present within the interiors of plates. The most important of these are orthomagmatic Ni–Cu–PGE systems that can form in intraplate settings as a consequence of mantle plume activity and the formation of large igneous provinces. Large igneous provinces (LIPs) represent significant reservoirs of energy and metals that can either drive or contribute to a variety of metallogenic systems. The main features were recently discussed in a comprehensive paper by Ernst and Jowitt (2013). In some cases, these events can trigger plate break-up and could be considered very early parts of the divergent tectono-metallogenic system (Section 8.1.3). Other mineral systems in intraplate settings are those that form within intracratonic basins, including some types of U systems and deposits associated with salt lakes, although, again, some of these systems may be the earliest indications of the initiation of a divergent system.

10.1. Orthomagmatic Ni–Cu–PGE systems associated with intracratonic large igneous provinces

Unlike many metal resources that are concentrated in the continental and oceanic crust and can be recycled and redistributed through typical crustal processes, Ni, PGEs and, to a lesser extent, Cu are strongly partitioned into the core (e.g. O'Neill et al., 1995: Table 1), residing primarily in a reservoir that is inaccessible to direct observation and sampling. Indeed, most of our understanding of the behaviour of these elements within the terrestrial system is derived from rare insights provided by iron meteorites (in essence fragmentary pieces of the cores of other planetary bodies), or through analysis of cm-scale mantle nodules and xenoliths entrained and brought to the surface by deeply sourced magmas, which are usually emplaced in intra-cratonic settings.

A number of times over the last ca. 3.5 billion years, however, major magmatic events, probably linked to the emplacement of mantle plumes (Campbell and Griffiths, 1992), have disturbed this segregated system, extracting metal from depth, and depositing large quantities of these elements in the crust as Ni–Cu–PGE sulphide deposits (Barnes et al., 1985; Naldrett, 1989; Maier et al., 2009). Although formed mostly in the deep crust (e.g. Li et al., 2007; Seat et al., 2007), post-mineralisation exhumation has, in some cases, brought these sulphide bodies into the near-surface environment, allowing exploitation. Many of these events are not related to tectonic processes linked to convergence or divergence, but are intracratonic and related to perturbations in mantle convection. However, in many cases, these deposits are located near the edges of cratonic blocks, possibly because this environment allowed easy passages of the magmas to the mid-crust (Begg et al., 2010).

The Earth's upper mantle melts when it is rapidly decompressed or in response to fluid flux. The latter occurs above dehydrating subducting oceanic slabs and is associated with construction of magmatic arcs (Hildreth, 1981; Tatsumi and Eggins, 1995). Mantle decompression occurs when the lithosphere is perturbed by the arrival of a plume from the deep mantle, or during continental stretching and breakup (Ernst et al., 2005; Pirajno and Hoatson, 2012). These triggers of partial melting are commonly connected and may lead to the production of enormous volumes of mafic magma that are preserved throughout the geological record as large igneous provinces (LIPs).

The large volumes of mafic magma, generated through partial melting of the mantle, ascend through the crust via interconnected networks of dykes and sills, and pipe-like structures known as chonoliths. These pathways are capable of moving melts vertically and over 2000 km laterally from the locus of mantle melting (Ernst and Baragar, 1992; Elliot and Fleming, 2004; Holt et al., 2014). The primary conduits are sub-vertical dykes, whereby mantle-derived magma is transported to the middle and upper crust to form sills, layered mafic intrusions and flood basalts (Ernst et al., 2005; Elliot and Fleming, 2004).

On the other hand, sills are capable of transporting magma laterally over 10s to 100s of kilometres at all crustal levels and are considered to be the fundamental building blocks of much larger magmatic bodies, which generally consist of several stacked sills (Cruden and McCaffrey, 2013; Menand, 2011). The leading edges of both dykes and sills can break down and propagate as pipe-like or cylindrical structures (Nicholson and Pollard, 1985; Schofield et al., 2013). Formation of these pipe-shaped magma conduits and their role and location within larger-scale melt transport networks are very poorly constrained. Although some of this magma eventually erupts on the surface, as much as 75% of this magma may remain as sills, layered mafic intrusions and chonoliths of dolerite, gabbro and peridotite (Ernst et al., 2005; Gudmundsson, 1983), which are generally localised in intra-cratonic settings.

Mafic magmas that leave the mantle and ascend through the crust are generally enriched in Ni, Cu and PGEs, which under favourable chemical and physical conditions become concentrated in the crust to form economic ore bodies. There is current consensus that such deposits form in high flux magma systems that develop by self organisation of flow within magma pathways (Naldrett, 2004). In such conduits, sulphide minerals typically accumulate in mafic and ultramafic igneous rocks. The most highly mineralised rocks form by high degrees of mantle melting, followed by transport, emplacement and crystallisation of the resulting magmas either within the crust (intrusion-hosted deposits: a strong but poorly understood spatial and genetic association exists between chonoliths and magmatic Ni–Cu–PGE mineralisation, best exemplified by the giant Noril'sk-Talnakh deposits in Siberia; cf. Naldrett, 2004), or onto the surface (komatiite lava-hosted deposits: Section 9.2.3). In all cases, Ni–Cu–PGE sulphide mineralisation forms if: (1) the magma reaches sulphur saturation, triggered either by contamination by sulphur-bearing crustal rocks and/or by sudden change in intensive physical parameters in the magma; (2) the resulting sulphide droplets can scavenge Ni–Cu–PGE from a large volume of magma; and (3) the sulphide droplets aggregate and then are transported and deposited to form a coherent ore body.

10.2. Deposits hosted by intracratonic basins

10.2.1. Sandstone-hosted uranium deposits

Sandstone-hosted U deposits are epigenetic deposits in which U minerals are present as disseminations and replacements primarily in fluvial, lacustrine and deltaic sandstones (Finch and Davis, 1985). Uranium is precipitated under reducing conditions, caused by one or more reducing agents within the sandstone, including carbonaceous material, sulphides, hydrocarbons and interbedded Fe^{2+}-rich volcanic rocks. Most basins that host these deposits formed on stable cratons that were closed to the sea, which produced local reducing conditions during and after sedimentation, forming chemical traps that were essential for U precipitation. There are five important prerequisites for the formation of sandstone-hosted U deposits, which include: (1) formation after the oxidation of the atmosphere and hydrosphere, (2) a leachable U source; (3) the presence of a stable, permeable aquifer and stable aquicludes; (4) a stable groundwater system to form an upstream oxidised zone; and (5) a reduction front, where the U is precipitated and concentrated.

A large number of sandstone-hosted uranium deposits occur in the western United States. The Powder River Basin in Wyoming, the Colorado Plateau and the Gulf Cost Plain in south Texas are major sandstone-hosted provinces. Other large sandstone-hosted deposits occur in Kazakhstan, Gabon and South Africa (Karoo Basin). The Callabonna Sub-basin in South Australia is the most richly endowed basin in Australia, but other deposits and prospects occur in the Ngalia and Amadeus basins in the Northern Territory, and the Canning, Carnarvon, and Western Eucla basins in Western Australia.

10.2.2. Surficial uranium and vanadium deposits

Surficial U deposits are young (Paleogene to Recent), near-surface concentrations in sediments or soils. Calcrete-hosted U deposits are the most significant of these deposits and form when U is leached from source rocks, and transported by oxidised surface water or shallow groundwater. Vanadium is extracted from mafic rocks or iron-rich (meta)sediments. Evaporation of the surface or groundwater triggers precipitation of carnotite (a hydrated U- and K-bearing vanadate). In the vicinity of playa lakes, precipitation of carnotite can occur due to mixing of valley groundwaters with more saline lake waters (Mernagh et al., 2013).

10.2.2.1. Australian examples. The main calcrete-hosted deposits in Australia include Yeelirrie, Lake Way, Centipede, Thatcher Soak, Lake Maitland and Napperby (all in Western Australia). Other calcrete-hosted deposits occur in Namibia, South Africa, Mauritania, Somalia, Argentina and China.

The Julia Creek V–Mo deposit represents a different style of deposit where the mineralisation has resulted from weathering of the calcareous oil shale between the surface and 15 m vertical depth. Lewis et al. (2010) suggest that V was accumulated biogenically and is associated with organic carbon. As a result of diagenesis and weathering V in the oxidised zone has been remobilised and is bound in poorly ordered iron oxide/hydroxide phases such as goethite. Molybdenum was accumulated into diagenetic sulphide minerals and is associated with sulphides in the fresh units and with goethite in the oxidised units.

10.2.3. Salt lake deposits

Globally, salt lakes are major sources of Li, potash, borates and other strategic mineral commodities (Mernagh et al., 2013). Salt lakes or 'salars' in the semi-arid to arid regions of Chile, Peru, Argentina and Bolivia contain most of the world's low-cost supply of Li, whereas salt lakes in Jordan, Israel and China are significant sources of potash. Three basic conditions are needed to form salt lakes (Eugster and Hardie, 1978). First, outflow must be absent or severely restricted to ensure hydrological closure. Second, evaporation must exceed inflow, and third, inflow must be sufficient to form a body of water (that may be ephemeral) at or very close to the surface. Therefore, favourable locations for the formation of salt lakes are arid basins in the rain-shadows of mountain ranges or highland areas, which provide the catchment for precipitation. In areas of lower relief, shallow basins may act as the focus of local discharge and evaporation from regionally extensive groundwater systems.

The mineralising fluids are meteoric, surficial, and/or groundwater. Groundwater may be derived from a local or regional meteoric system, interstitial water from sediments, or deep basinal or hydrothermal fluids. Hydrothermal discharge (e.g. from geothermal springs), though often small in volume, can be significant in terms of solute contribution. Most of the surface and groundwater flow is gravity-driven, with fluids driven from regions of higher hydraulic head to regions of lower hydraulic head. Potassium, Li and B enrichment in salt lakes mainly results from two processes: (1) enrichment of fluids in K, Li, and B, prior to the evaporation process, due to leaching, and (2) evaporative concentration of the inflow water. In some cases, Li and/or B enrichment is also related to long-lasting, geothermal springs, associated with prolonged volcanic activity or a high underlying geothermal gradient.

Potassium may be sourced from a variety of lithologies but the most productive basins contain some surface or near-surface evaporites or salts. Economic K systems are characterised by ~10- to 100-fold enrichment in K relative to the values that can be expected from concentrated or diluted seawater with the same chloride concentrations. The fluids have variable salinity up to ~400,000 mg/L TDS (Total Dissolved Solids). Economic Li systems are characterised by ~100- to 1000-fold enrichment in Li content relative to the values that can be expected from concentrated or diluted seawater with the same chloride concentrations;

and by ~10- to 100-fold enrichment in B, respectively. An important factor in Li-enrichment is repeated cycles of evaporation of freshly introduced waters in the lake system. The Li and B enrichment reaches its maximum level at halite-saturation and the total salinity of economic lakes is >100,000 mg/L TDS (Mernagh et al., 2013).

11. Genetic and exploration implications

In this discussion, we have attempted to move beyond observation-based classifications of mineral deposits, rather placing deposits into a tectonic-based system, following Hutchinson (1973); Sawkins (1984) and Kerrich et al. (2000, 2005). Moreover, we have considered how mineral systems change as tectonic systems evolve, and how tectonic systems themselves have evolved through geological time. Although not addressed explicitly in the contribution, the evolution of the atmosphere and hydrosphere has also influenced the secular distribution and some mineral deposit classes as well as characteristics of other classes (Meyer, 1981; Groves et al., 2005; Goldfarb et al., 2010, and the references therein). This analysis suggests that there are fundamental differences between mineral systems that form in different tectonic systems and different stages of individual tectonic systems. These differences include the triggers and drivers of mineralising events and the chemical composition of the ore fluids and, thereby, depositional mechanisms. The tectonic system also determines the lithospheric architecture in which mineral systems form.

11.1. Links between tectonic and metallogenic systems

As many types of mineral deposits form as the consequence of geodynamic processes that occur in different tectonic systems, individual types of mineral deposits, in most cases, are *not* diagnostic of tectonic system. As an example, VHMS deposits are forming currently in both divergent and convergent systems (Hannington et al., 2005). Assemblages of deposit types produced within one metallogenic system, however, can be strongly indicative of the tectonic system in which they formed. For example, the presence of porphyry Cu, VHMS and orogenic gold deposits in close temporal and geographical proximity is indicative of an overall convergent tectonic system (cf. Mercier-Langevin et al., 2014). Associations of tectonically-related mineral deposits may provide additional constraints in determining the tectonic environment in which the host succession formed.

11.2. Tectonic system and the chemical characteristics of resident fluids

Many of the chemical characteristics of ore fluids appear to depend, to a significant degree, upon the stage of the tectono-metallogenic system in which they form. For example, as high heat flow and extensive (mafic) magmatism characterise all evolutionary stages of the convergent system and the rift stage of the divergent system, ore fluids produced in these environments (e.g. those associated with porphyry Cu, VHMS, siliciclastic–mafic Zn–Pb deposits, among others) tend to be high temperature (>200 °C) and relatively reduced (i.e. $\Sigma H_2S > \Sigma SO_4$) (Titley and Beane, 1981; Franklin et al., 1981; Huston et al., 2006). As a consequence, the major causes of mineral deposition in these environments are steep temperature decreases (arrow D in Fig. 9), water–rock reactions, depressurisation (boiling) and fluid mixing. Depending upon the reactive rock, water–rock reactions can increase the pH (with carbonates), desulphidise (with Fe-rich rocks) and reduce (with Fe- or reduced-C-rich rocks) the ore fluids, all mechanisms which can cause mineral deposition.

This contrasts with the latter stages of the divergent tectono-metallogenic system, which is amagmatic and, therefore, characterised by low heat flow. Moreover, the relatively oxidised character of many basins deposited during this stage (i.e. thermal subsidence or passive margin basins) results generally in the production of oxidised, H_2S-poor, low temperature (<200 °C) fluids (e.g. those associated with shale-hosted Cu–Co, unconformity-related U and siliciclastic–carbonate Zn–Pb deposits, among others: Hitzman et al., 2005; Jaireth et al., 2016-in this issue; Huston et al., 2006). As shown in Fig. 9 (arrows A–B), the most effective way of depositing metals from such fluids is the provision of H_2S, either through sulphate reduction (Fig. 9, arrow A) or interaction with a local H_2S-rich source (e.g. sour gas: Huston et al., 2006).

An important consequence of different characteristics of fluids that form within tectono-metallogenic systems is that the characteristics of depositional sites differ, and can be predicted once the tectono-metallogenic system is known or can be inferred.

11.3. The architecture of tectono-metallogenic systems

As shown in Fig. 3, mineral deposits are the result of nested processes that can occur over periods of millions to hundreds of millions of years. This not only involves multiple geochemical enrichment events to produce enriched source regions, but also the development of crustal architecture that is later used by mineralising fluids.

For example, subduction is not only directly responsible for the formation of many porphyry Cu belts, as the melts that form these deposits are subduction-related, but devolatilisation of the subducting slab also enriches the mantle in volatiles, and slab roll-back can cause extension of the over-riding plate (Richards, 2009). Both of these latter processes can have important ramifications to later metallogenic systems. For example, Hayward and Skirrow (2010) suggested that the formation of IOCG deposits may have involved the melting of mantle that was fertilised via devolatilisation of subducted slabs during a prior period of subduction. Moreover, it is likely that extension related to subduction produces the structural architecture utilised by syn-volcanic mineral systems (e.g. VHMS), and later by orogenic systems as it is reactivated (e.g. orogenic gold) (see also Miller et al., 2010; McCuaig and Hronsky, 2014; Mercier-Langevin et al., 2014). Hence, chemical and structural architecture may develop well before a metallogenic event can create favourable conditions for the development of fluids and melts critical to metallogenesis, and can also channel and compartmentalise fluid flow.

11.4. Tectonic perturbations and giant ore deposits?

For most commodities, economically extractable resources are concentrated in a relatively small proportion of deposits, and these deposits are where most of the economic value lies. As exploration moves into more remote and/or covered terranes, these giant deposits will become even more important, due to the higher costs of exploitation of remote and deeply buried resources. As a consequence, an understanding on the controls on the formation of such giant deposits is essential for effective exploration in remote and/or covered terranes.

Factors that control the formation of giant ore bodies have been the subject of many studies (e.g. Phillips et al., 1996; Kesler, 1997; Jaques et al., 2002; Emsbo et al., 2006; Leach et al., 2010; McCuaig and Hronsky, 2014). One of the key factors identified by these and other studies is the importance of tectonic events that cause changes in the regional far-field stress regime.

Evidence from apparent polar wander paths in Australia (e.g. Idnurm, 2000) and elsewhere (e.g. Elston et al., 2002) suggest that the motion of geological blocks tends to be relatively smooth punctuated by very short and sharp deviations. As shown by Idnurm (2000) in North Australia, major mineral deposits are associated with these deviations, which can be interpreted as major perturbations from steady-state plate motion of the types now associated with tectonic mode switches (Collins, 2002a). An association of major ore deposits with perturbations in plate motion is also noted in younger rocks. For instance, major orogenic gold deposits in the Juneau goldfield in Alaska are temporally associated with a change in Pacific plate motion at ~56 Ma (Miller et al., 1988), and porphyry Cu deposits along the Andean margin of South America area associated with the shallowing of subduction (Sillitoe and Perelló, 2005).

In southeast Australia, the giant orogenic gold system in the Victorian goldfields, and the major Macquarie porphyry province, appear to have formed as a consequence of the oblique docking of VanDieland and ensuing development of an orocline (Cayley, 2012; Moresi et al., 2014). Immediately prior to, and after, this tectono-metallogenic event, when the Paleozoic eastern Australian convergent margin was in steady state, small to moderate-sized VHMS and intrusion related deposits formed (Section 9.5). It is possible that, in many tectono-metallogenic systems, small deposits form during periods of steady state, whereas major deposits form when the system is perturbed, such as during the Lachlan tectonic mode switch (Collins, 2002a). An analysis by McCuaig and Hronsky (2014) came to a similar conclusion. Identification of these perturbances, and their consequences in time and space, which can be provided by a variety of datasets (e.g. geochronology, structural and paleomagnetic), could provide insight and vectors to the locations of giant deposits.

11.5. Exploration implications

As discussed by Schodde (2011); McCuaig and Hronsky (2014) and many other authors, most "easily" discovered deposits have largely been found, and future discoveries will increasingly be made in poorly exposed or under-cover terranes. Consequently, conventional exploration techniques will become less effective and new exploration methods and concepts will be required.

Analysis of the links between the evolution of tectonic systems and metallogeny suggests that an understanding of the tectonic setting can be used to predict characteristics of mineralisation in poorly known terranes. This tectonic understanding can be gained either from better exposed terranes along strike or, in some cases, regional- or national-scale geochemical, isotopic or geophysical datasets. Moreover, the concept of tectonically-linked mineral deposit associations can be used to predict unknown styles of mineralisation in known mineral provinces. If indeed major mineral deposits are the products of perturbations on tectono-metallogenic systems (e.g., McCuaig and Hronsky, 2014), documentation of such perturbations and their spatial and temporal effects could assist the development of regional exploration models that specifically target major deposits.

In many tectono-metallogenic systems, mineralisation is strongly controlled by major crustal breaks (McCuaig and Hronsky, 2014). In Australia, these crustal breaks are being identified using a combination of new-generation geological mapping (e.g. VandenBerg et al., 2000), geochronology, and geophysical (seismic, gravity, aeromagnetic and other dataset (both observational and derived): Korsch and Doublier, 2014; Korsch and Doublier, 2016-in this issue; McCuaig and Hronsky, 2014) and isotopic (Sm–Nd and Pb–Pb: Champion, 2013; Champion and Huston, 2016-in this issue) data. Importantly, the data used to identify these breaks is regional and national in scale, datasets that are commonly available in remote or covered regions. Although an understanding of the relationship between tectonic evolution and mineralisation, combined with national to regional datasets, can vector to potential mineral provinces or districts, more detailed geophysical data and other different data (e.g. lithological, structural and geochemical) are required for district- and prospect-scale exploration.

12. Conclusions

Although the relationships between geodynamics, tectonics and mineral systems are complex, a number of generalisations can be made about this relationship that have bearing on the use of tectonic analysis to predict the temporal and spatial distributions of mineral systems:

- Ore deposits are the products of geological systems that can operate over a long time frame (to hundreds of millions of years) and at scales from the ore shoot to the craton.
- Mineral systems are the product of geodynamic processes that may operate in different tectonic settings, so individual deposit types may not be diagnostic of tectonic setting.
- Lithospheric architecture, which controls the spatial distribution of and fluid flow within mineral systems, is largely determined by current and past geodynamic processes.
- In some cases, mineral deposit *associations* can be used as indicators of tectonic setting.
- Mineral systems that form mostly in divergent tectonic settings include clastic-dominated Zn–Pb–Ag systems, many systems associated with alkalic magmatism, Lake Superior-type banded-iron formation systems (including later upgrading), unconformity-related U systems and shale-hosted Cu–Co–Ag systems.
- Mineral systems that form mostly in convergent tectonic settings include porphyry-epithermal Cu–Au–Mo, VHMS Zn–Pb–Cu–Ag–Au, komatiite-hosted Ni–Cu–PGE, orogenic gold, orogenic base and precious metal (Cu–Au–Zn–Pb–Ag), Mississippi Valley-type Zn–Pb, and intrusion-related W–Sn–Mo–Au–Cu, IOCG Cu–Au–U and Algoma-type BIF (including later upgrading) systems.
- Mineral systems that form mostly in intraplate tectonic settings include LIP-associated orthomagmatic Ni–Cu–PGE systems, and U, V and potash systems associated with intracratonic basins, including salt lakes.
- The duration of mineralising events is commonly relatively short (<5 Myr and commonly <1 Myr) and correlates with local and/or far-field tectonic events such as a tectonic mode switch within the tectonic system.
- Metallogenesis evolves as tectonic setting evolves, leading to the staged development of mineral systems and deposit types within an evolving tectonic environment. This is illustrated well by the evolution of the Ordovician to Silurian Lachlan tectono-metallogenic system in southeastern Australia (Fig. 18).
- The geochemical characteristics of many mineral systems are a consequence of their geodynamic and tectonic settings. Geodynamic and tectonic settings, that are characterised by low heat flow and lack active magmatism, produce low temperature fluids that are oxidised, with ore formation caused largely by redox gradients or the provision of external H_2S. In settings characterised by high heat flow and active magmatism, ore fluids tend to be higher temperature and reduced, with deposition caused by cooling, pH neutralisation, depressurisation and fluid mixing.
- Tectonic processes and, by implication, metallogenesis have evolved as the Earth's interior has cooled, with the most important changes occurring during the Meso- to Neoarchean (probable transition from pre-subduction to proto-subduction) and the Neoproterozoic-Paleozoic boundary (transition from proto-subduction to modern-style subduction). Changes in the characteristics of some deposits and mineral systems can be linked to these transitions.
- Better understanding of geodynamic setting and the development of tectonic/geodynamic models can lead to greater understanding of large scale controls on mineral systems, which will result in greater predictive ability, leading to more targeted exploration and more effective discovery.
- New and/or updated and modern national-scale datasets (e.g. major crustal boundaries, radiogenic isotope maps, tomographic maps, magneto-telluric data) are critical to understanding tectonic architecture and to predict metallogenic provinces under-cover in Australia and elsewhere.

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Appendix A. Supplementary data

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References

Ahmad, M., 1998. Geology and mineral deposits of the Pine Creek Inlier and McArthur Basin, Northern Territory. AGSO J. Aust. Geol. Geophys. 17, 1–17.
Ahmad, M., Scrimgeour, I.R., 2013. Halls Creek Orogen. In: Ahmad, M., Munson, T.J. (Eds.), Geology and Mineral Resources of the Northern Territory 7. Darwin, Northern Territory Government, pp. 1–4.
Ahmad, M., Wygralak, A.S., Ferenczi, P.A., Scrimgeour, I.R., 2009. Gold deposits of the Northern Territory, second edition. North. Territ. Geol. Surv. Rep. 11 (131 pp.).
Ahmat, A.L., 2012. The Ellendale diamond field: exploration history, discovery, geology and mining. Aust. Gemmol. 24, 280–288.
Alkane. Resources, 2014. 2014 Annual Report. (http://www.alkane.com.au/index.php/reports/annual-reports).
Allen, P.A., Allen, J.R., 2005. Basin Analysis — Principles and Applications. second ed. Blackwell Publishing, Oxford (549 pp.).
Anderson, C.A., 1968. Arizona and adjacent New Mexico. In: Ridge, J.D. (Ed.), Ore Deposits in the United States, 1933/1976. The American Institute of Mining, Metallurgical and Petroleum Engineers, New York, pp. 1163–1190.
Anderson, G.M., 1975. Precipitation of Mississippi Valley-type ores. Econ. Geol. 70, 937–942.
Anderson, D.L., 1982. Hotspots, polar wander, Mesozoic convection, and the geoid. Nature 297, 391–393.
Anderson, B.R., Gemmell, J.B., Berry, R.F., 2001. The geology of the Nifty copper deposit, Throssell Group, Western Australia: implications for ore genesis. Econ. Geol. 96, 1535–1565.
Angerer, T., Duuring, P., Hagemann, S.G., Thorne, W., McCuaig, T.C., 2014. A mineral system approach to Archean and Paleoproterozoic banded iron formation hosted iron ore of Western Australia. Geol. Soc. Lond. Spec. Publ. 393.
Angerer, T., Hagemann, S.G., Walde, D., Halverson, G.P., 2015. BIF Chemostratigraphy in the Neoproterozoic Jacadigo Group in the Santa Cruz Hematite Deposit, Brazil SGA extended abstract volume.
Annels, A.E., 1989. Ore genesis in the Zambia copperbelt, with particular reference to the northern sector of the Chambashi basin. Geol. Assoc. Can. Spec. Pap. 36, 427–452.
Anschutz, P., Turner, J.S., Blanc, G., 1998. The development of layering, fluxes through double-diffusive interfaces, and location of hydrothermal sources of brines in the Atlantis II Deep: Red Sea. J. Geophys. Res. 103, 27809–27819.
Arndt, N., 2003. Komatiites, kimberlites, and boninites. J. Geophys. Res. 108 (B6), 2293.
Arndt, N., Lesher, C.M., Barnes, S.J., 2008. Komatiite. Cambridge University Press (487 pp.).
Arndt, N., Lesher, C.M., Czamanske, G.K., 2005. Mantle-derived magmas and magmatic Ni–Cu–(PGE) deposits. Econ. Geol. 5–24 (100th Anniversary volume).
Ashley, P.M., Dudley, R.J., Lesh, R.H., Marr, J.M., Ryall, A.W., 1988. The Scuddles Cu–Zn prospect, an Archean volcanogenic massive sulfide deposit, Golden Grove district, Western Australia. Econ. Geol. 83, 918–951.
Baker, E.M., Andrew, A.S., 1991. Geologic, fluid inclusions, and stable isotope studies of the gold-bearing breccia pipe at Kidston, Queensland, Australia. Econ. Geol. 86, 810–830.
Baker, T., Pollard, P.J., Mustard, R., Mark, G., Graham, J.L., 2005. A comparison of granite-related tin, tungsten, and gold–bismuth deposits: implications for exploration. Soc. Econ. Geol. Newsl. 61, 5–17.
Barley, M.E., 1997. The Pilbara craton. Oxf. Univ. Monogr. Geol. Geophys. 35, 657–664.
Barley, M.E., Groves, D.I., 1992. Supercontinent cycles and the distribution of metal deposits through time. Geology 20, 291–294.
Barley, M.E., Pickard, A.L., Hagemann, S.G., Folkert, S.L., 1999. Hydrothermal origin for the 2 billion year old Mount Tom Price giant iron ore deposit, Hamersley Province, Western Australia. Mineral. Deposita 34, 784–789.
Barley, M.E., Pickard, A.L., Sylvester, P.J., 1997. Emplacement of a large igneous province as a possible cause of banded iron formation 2.45 billion years ago. Nature 385, 55–58.
Barnes, S.J., Cruden, A.R., Arndt, N., Saumur, B.M., 2016. The mineral system approach applied to magmatic Ni–Cu–PGE sulfide systems. Ore. Geol. Rev. 76, 296–316 (in this issue).
Barnes, S.J., Fiorentini, M.L., 2012. Komatiite magmas and Ni sulfide deposits: a comparison of variably endowed Archean terranes. Econ. Geol. 107, 755–780.
Barnes, S.-J., Naldrett, A.J., Gorton, M.P., 1985. The origin of the fractionation of the platinum-group elements in terrestrial magmas. Chem. Geol. 53, 303–325.
Barnicoat, A.C., 2007. Mineral systems and exploration science: linking fundamental controls on ore deposition with the exploration process. In: Andrew, C.J., et al. (Eds.), Digging Deeper: Proceedings of the Ninth Biannual SGA Meeting, Dublin, Ireland 20th-23rd August 2007, pp. 1407–1410.
Barra-Pantoja, L.F., 2005. Applications of the Re–Os isotopic system in the study of mineral deposits: geochronology and source of metals. Unpub. Ph.D. thesis, University of Arizona, 211 p.
Barrie, C.T., 1993. Petrochemistry of shoshonitic rocks associated with porphyry copper–gold deposits of central Quesnellia, British Columbia, Canada. J. Geochem. Explor. 48, 225–258.
Barton, M.D., 2014. Iron oxide (–Cu–Au–REE–P–Ag–U–Co) systems. In: Holland, H.D., Turekian, K.K. (Eds.), Treatise on Geochemistry, Second ed. 13, pp. 515–541.
Beales, F.W., 1975. Precipitation mechanisms for Mississippi Valley-type deposits. Econ. Geol. 70, 943–948.
Begg, G.C., Griffin, W.L., Natapov, L.M., O'Reilly, S.Y., Grand, S.P., O'Neill, C.J., Hronsky, J.M.A., Poudjom Djomani, Y., Swain, C.J., Deen, T., Bowden, P., 2009. The lithospheric architecture of Africa: seismic tomography, mantle petrology and tectonic evolution. Geosphere 5, 23–50.
Begg, G.C., Hronsky, J.A.M., Arndt, N.T., Griffin, W.L., O'Reilly, S.Y., Hayward, N., 2010. Lithospheric, cratonic, and geodynamic setting of Ni–Cu–PGE sulfide deposits. Econ. Geol. 105, 1057–1070.
Behrendt, J.C., 1999. Crustal and lithospheric structure of the West Antarctic Rift System from geophysical investigations — a review. Glob. Planet. Chang. 23, 25–44.
Bekker, A., Barley, M.E., Fiorentini, M.L., Rouxel, O.J., Rumble, D., Beresford, S.W., 2009. Atmospheric sulphur in Archaean komatiite-hosted nickel deposits. Science 326, 1086–1089.
Bekker, A., Slack, J.F., Planavsky, N., Krapež, B., Hofmann, A., Konhauser, K.O., Rouxel, O.J., 2010. Iron formation: the sedimentary product of a complex interplay among mantle, tectonic, oceanic, and biospheric processes. Econ. Geol. 105, 467–508.
Belousova, E.A., Kostitsyn, Y.A., Griffin, W.L., Begg, G.C., O'Reilly, S.Y., Pearson, N.J., 2010. The growth of the continental crust: constraints from zircon Hf-isotope data. Lithos 119, 457–466.
Belperio, A.P., Preiss, W.V., Fairclough, M.C., Gatehouse, C.G., Gum, J., Hough, J., Burtt, A., 1997. Tectonic and metallogenic framework of the Cambrian Stansbury Basin-Kanmantoo Trough, South Australia. AGSO J. Aust. Geol. Geophys. 17, 183–201.
Berkman, D.A., Mackenzie, D.H. (Eds.), 1998. Geology of Australian and Papua New Guinean Mineral Deposits 22. Australasian Institute of Mining and Metallurgy Monograph (894 pp.).
Berry, R.F., Huston, D.L., Stolz, A.J., Hill, A.P., Beams, S.D., Kuronen, U., Taube, A., 1992. Stratigraphy, structure and volcanic-hosted mineralisation of the Mount Windsor Subprovince, North Queensland, Australia. Econ. Geol. 87, 739–763.
Beukes, N.J., Gutzmer, J., 2008. Origin and paleoenvironmental significance of major iron formations at the Archean–Paleoproterozoic boundary. Rev. Econ. Geol. 15, 5–47.
BHP Billiton, 2012. BHP Billiton Annual Report 2012 (277 pp.).
Bierlein, F.P., Arne, D.C., Foster, D.A., Reynolds, P., 2001. A geochronological framework for orogenic gold mineralisation in central Victoria, Australia, Australia. Mineral. Deposita 36, 741–767.
Bierlein, F.P., Groves, D.I., Cawood, P.A.C., 2009. Metallogeny of accretionary orogens—the connection between lithospheric processes and endowment. Ore Geol. Rev. 36, 282–292.
Bierlein, F.P., Groves, D.I., Goldfarb, R.J., Dube, B., 2006. Lithospheric controls on the formation of provinces hosting giant orogenic gold deposits. Mineral. Deposita 40, 874–887.
Bissig, T., Clark, A.H., Lee, J.K.W., Hodgson, C.J., 2002. Miocene landscape evolution and geomorphologic controls on epithermal processes in the El Indio-Pascua Au–Ag–Cu Belt, Chile and Argentina. Econ. Geol. 97, 971–996.
Black, L.P., McClenaghan, M.P., Korsch, R.J., Everard, J.L., Foudoulis, C., 2005. Significance of Devonian–Carboniferous igneous activity in Tasmania as derived from U–Pb SHRIMP dating of zircon. Aust. J. Earth Sci. 52, 807–829.
Bleeker, W., 2012. Targeted Geoscience Initiative 4. Lode gold deposits in ancient deformed and metamorphosed terranes: the role of extension in the formation of Timiskaming Basins and large gold deposits, Abitibi Greenstone Belt — a discussion. Geol. Surv. Can. 47.1–47.12 (Open File. Report 6280).
Blevin, P.L., 2004. Redox and compositional parameters for interpreting the granitoid metallogeny of Eastern Australia: implications for gold-rich ore systems. Resour. Geol. 54, 241–252.
Blevin, P.L., Chappell, B.W., 1992. The role of magma sources, oxidation states and fractionation in determining the granite metallogeny of eastern Australia. Trans. R. Soc. Edinb. Earth Sci. 83, 305–316.
Blevin, P.L., Chappell, B.W., Allen, C.M., 1996. Intrusive metallogenic provinces in eastern Australia based on granite source and composition. Trans. R. Soc. Edinb. Earth Sci. 87, 281–290.
Blewett, R.S., Czarnota, K., Henson, P.A., 2010. Structural-event framework for the eastern Yilgarn Craton, Western Australia, and its implications for orogenic gold. Precambrian Res. 183, 203–229.
Bosence, D., 2005. A genetic classification of carbonate platforms based on their basinal and tectonic settings in the Cenozoic. Sediment. Geol. 175, 49–72.
Bosworth, W., Huchon, P., McClay, K., 2005. The Red Sea and gulf of Aden basins. J. Afr. Earth Sci. 43, 334–378.

Both, R.A., Williams, K.L., 1968. Mineralogical zoning in the lead-zinc ores of the Zeehan field, Tasmania; part II, paragenetic and zonal relationships. J. Geol. Soc. Aust. 2, 217–243.

Bown, J.W., White, S., 1995. Effect of finite extension rate on melt generation at rifted continental margins. J. Geophys. Res. 100, 18011–18029.

Boyd, F.R., 1998. The origin of cratonic peridotites: a major-element approach. Int. Geol. Rev. 40, 755–764.

Bradley, D.C., 2008. Passive margins through time. Earth Sci. Rev. 91, 1–26.

Bradshaw, M.T., 1993. Australian petroleum systems. PESA J. 21, 43–53.

Brauhart, C.W., Groves, D.I., Morant, P., 1998. Regional alteration systems associated with volcanogenic massive sulfide mineralization at Panorama, Pilbara, Western Australia. Econ. Geol. 93, 292–302.

Brauhart, C.W., Huston, D.L., Groves, D.I., Mikucki, E.J., Gardoll, S.J., 2001. Geochemical mass transfer patterns as indicators of the architecture of a complete volcanic-hosted massive sulfide hydrothermal alteration system in the Panorama district, Pilbara, Western Australia. Econ. Geol. 96, 1263–1278.

Breit, G.N., Meurnier, J.D., 1990. Fluid inclusion, ^{18}O, and $^{87}Sr/^{86}Sr$ evidence for the origin of fault-controlled copper mineralization, Lisbon Valley, Utah, and the Slick Rock district, Colorado. Econ. Geol. 85, 884–891.

Brown, A.C., 1971. Zoning in the White Pine Copper deposit, Ontonogan County, Michigan. Econ. Geol. 66, 543–573.

Brown, M., 2014. The contribution of metamorphic petrology to understanding lithosphere evolution and geodynamics. Geosci. Front. 5, 553–569.

Brown, M.C., Oliver, N.H.S., Dickens, G.R., 2004. Veins and hydrothermal fluid flow in the Mt. Whaleback Iron Ore District, Eastern Hamersley Province, Western Australia. Precambrian Res. 128, 441–474.

Bucher, K., Frey, M., 1994. Petrogenesis of Metamorphic Rocks. sixth edition. Springer Verlag, New York.

Burgess, P.M., 2008. Phanerozoic evolution of the sedimentary cover of the North American Craton. In: Andrew, D.M. (Ed.), Sedimentary Basins of the World. Elsevier, Amsterdam, pp. 31–63.

Burke, K., Dewey, J.F., 1973. Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. J. Geol. 81, 406–433.

Burke, K., Ashwal, L.D., Webb, S.J., 2003. New way to map old sutures using deformed alkaline rocks and carbonatites. Geology 31, 391–394.

Burnham, C.W., Ohmoto, H., 1980. Late stage processes of felsic magmatism. Min. Geol. Spec. Issue 8, 1–11.

Callaghan, T., 2001. Geology and host-rock alteration of the Henty and Mount Julia gold deposits, western Tasmania. Econ. Geol. 96, 1073–1088.

Campbell, I.H., Griffiths, R.W., 1992. The changing nature of mantle hotspots through time: implications for the chemical evolution of the mantle. J. Geol. 497-523.

Cassidy, K.F., Champion, D.C., McNaughton, N.J., Fletcher, I.R., Whitaker, A.J., Bastrakova, I.V., Budd, A.R., 2002. Characterization and metallogenic significance of Archaean granitoids of the Yilgarn Craton, Western Australia. AMIRA Project No. P482/MERIWA Project M281, MERIWA. Report 222 (514 pp.).

Cathles, L.M., 1983. An analysis of the hydrothermal system responsible for massive sulfide deposition in the Hokuroku Basin of Japan. Econ. Geol. Monogr. 5, 439–487.

Cawood, P.A., Buchan, C., 2007. Linking accretionary orogenesis with supercontinent assembly. Earth-Science Reviews 82, 217–256.

Cawood, P.A., Kröner, A., Pisarevsky, S., 2006. Precambrian plate tectonics: criteria and evidence. GSA Today 16, 4.

Cayley, R.A., 2012. Oroclinal folding in the Lachlan Fold Belt: consequence of southeast-directed Siluro-Devonian subduction rollback superimposed on an accreted Ordovician arc assemblage in eastern Australia. Geol. Soc. Aust. Ext. Abstr. 103, 34–43.

Cayley, R.A., Taylor, D.H., 2001. Ararat 1:100 000 map area geological report. Geological Survey of Victoria Report 115.

Cayley, R.A., Korsch, R., Kennett, B., Skladzien, P., Jones, L., Morand, V.J., Gibson, G., Rawling, T.J., Betts, P.G., 2011b. Results of deep seismic reflection imaging of the eastern Delamerian Orogen, South Australia and western Victoria, Australia. Data CD Version: 4, March, 2011. Department of Primary Industries, GeoScience Victoria.

Cayley, R.A., Korsch, R.J., Moore, D.H., Costeloe, R.D., Nakamura, A., Willman, C.E., Rawling, T.J., Morand, V.J., Skladzien, P.B., O'Shea, P.J., 2011a. Crustal architecture of central Victoria: results from the 2006 deep crustal seismic reflection survey. Aust. J. Earth Sci. 58, 113–156.

Cayley, R.A., Taylor, D.H., VandenBerg, A.H.M., Moore, D.H., 2002. Proterozoic–Early Palaeozoic rocks and the Tyennan Orogeny in central Victoria: the Selwyn Block and its tectonic implications. Aust. J. Earth Sci. 49, 225–254.

Chalmers, D.I., 1990. Brockman multi-metal and rare earth deposit. Aust. Inst. Min. Metall. Monogr. 14, 707–709.

Champion, D.C., 2013. Neodymium depleted mantle model age map of Australia: explanatory notes and user guide. Geosci. Aust. Rec. 2013/044.

Champion, D.C., Cassidy, K.F., 2007. An overview of the Yilgarn Craton and its crustal evolution. Geosci. Aust. Rec. 2007 (14), 8–13.

Champion, D.C., Cassidy, K.F., 2008. Geodynamics: Sm–Nd isotopes and the evolution of the Yilgarn and similar terrains. Geosci. Aust. Rec. 2008 (09), 7–16.

Champion, D.C., Huston, D.L., 2016. Radiogenic Isotopes, ore deposits and metallogenic terranes: novel approaches based on regional isotopic maps and the mineral systems concept. Ore. Geol. Rev. 76, 229–256 (in this issue).

Champion, D.C., Mackenzie, D.E., 1994. Igneous rocks of northern Queensland: 1:500 000 map and explanatory notes. Australian Geological Survey Organisation Metallogenic Atlas Series 2.

Champion, D.C., Kositcin, N., Huston, D.L., Mathews, E., Brown, C., 2009. Geodynamic synthesis of the Phanerozoic of eastern Australia and implications for metallogeny. Geosci. Aust. Rec. 2009/18 (254 pp.).

Chapin, C.E., 1979. Evolution of the Rio Grande Rift — a summary. Am. Geophys. Union Spec. Publ. 14, 1–5.

Chiaradia, M., Schaltegger, U., Spikings, R., 2014. Time scales of mineral systems—advances in understanding over the past decade. Soc. Econ. Geol. Spec. Publ. 18, 37–58.

Christensen, J.N., Halliday, A.N., Vearncombe, J.R., Kesler, S.E., 1995a. Testing models of large-scale crustal fluid flow using direct dating of sulfides: Rb–Sr evidence of early dewatering and formation of MVT deposits, Canning Basin, Australia. Econ. Geol. 90, 877–884.

Christensen, J.N., Halliday, A.N., Vearncombe, J.R., Kesler, S.E., 1995b. Testing models of large-scale crustal fluid flow using direct dating of sulfides: Rb–Sr evidence of early dewatering and formation of MVT deposits, Canning Basin, Australia. Econ. Geol. 90, 877–884.

Clague, D.A., Dalrymple, G.B., 1987. The Hawaiian-Emperor volcanic chain. Part I. Geologic evolution. U. S. Geol. Surv. Prof. Pap. 135 (1), 5–54.

Clout, J.M.F., 2003. Upgrading processes in BIF-derived iron ore deposits: implications for ore genesis and downstream mineral processing. Trans. Inst. Min. Metall. B Appl. Earth Sci. 112, 89–95.

Collins, W.J., 1996. Lachlan Fold Belt granitoids: products of three-component mixing. Trans. Roy. Soc. Edinb. Earth Sci. 87, 171–181.

Collins, W.J., 2002a. The nature of extensional accretionary orogens. Tectonics 21, 6-1–6-12.

Collins, W.J., 2002b. Hot orogens, tectonic switching and creation of continental crust. Geology 30, 535–538.

Collins, W.J., Richards, S.W., 2008. Geodynamic significance of S-type granites in Circum-Pacific orogens. Geology 36, 559–562.

Condie, K.C., Kröner, A., 2008. When did plate tectonics begin? Evidence from the geologic record. Geol. Soc. Am. Spec. Pap. 440, 281–294.

Coney, P.J., 1980. Cordilleran metamorphic core complexes — an overview. Geol. Soc. Am. Mem. 153, 7–34.

Cooke, D.R., Bull, S.W., Donovan, S., Rogers, J.R., 1998. K-metasomatism and base metal depletion in volcanic rocks from the McArthur Basin, Northern Territory—implications for base metal mineralization. Econ. Geol. 93, 1237–1263.

Cooke, D.R., Bull, S.R., Large, R.R., McGoldrick, P.J., 2000. The importance of oxidized brines for the formation of Australian Proterozoic stratiform sediment-hosted Pb–Zn (SEDEX) deposits. Econ. Geol. 95, 1–19.

Cooke, D.R., Hollings, P., Walshe, J.L., 2005. Giant porphyry deposits: characteristics, distribution, and tectonic controls. Econ. Geol. 100, 801–818.

Cooke, D.R., Wilson, A.J., House, M.J., Wolfe, R.C., Walshe, J.L., Lickfold, V., Crawford, A.J., 2007. Alkalic porphyry Au–Cu and associated mineral deposits of the Ordovician to Early Silurian Macquarie Arc, New South Wales. Aust. J. Earth Sci. 54, 445–463.

Coombs, D.S., 1961. Some recent work on lower grades of metamorphism. Aust. J. Sci. 24, 203–215.

Corbett, G.J., Leach, T.M., 1998. Southwest Pacific rim gold-copper systems: structure, alteration, and mineralization. Soc. Econ. Geol. Spec. Pub. 6, 1998 234 p.

Crawford, A.J., Maas, R., 2009. A magmatic-hydrothermal origin for the giant Broken Hill Pb–Zn deposit. Geosci. Aust. Rec. 2009 (28), 28–30.

Crawford, A.J., Falloon, T.J., Green, D.H., 1989. Classification, petrogenesis and tectonic setting of boninites. Boninites and Related Rocks. Unwin Hyaman, pp. 1–49.

Crawford, A.J., Meffre, S., Squire, R.J., Barron, L.M., Falloon, T.J., 2007. Middle and Late Ordovician magmatic evolution of the Macquarie Arc, Lachlan Orogen, New South Wales. Aust. J. Earth Sci. 54, 181–214.

Crosby, A.G., Fishwick, S., White, N., 2010. Structure and evolution of the intracratonic Congo Basin. Geochem. Geophys. Geosyst. 11, 1–20.

Crosby, A.G., White, N.J., Edwards, G.R.H., Thompson, M., Corfield, R., Mackay, L., 2011. Evolution of deep-water rifted margins: testing depth-dependent extensional models. Tectonics TC1004 http://dx.doi.org/10.1029/2010TC002687.

Cross, A.J., 2009. SHRIMP U–Pb xenotime geochronology and its application to dating mineralization, sediment deposition and metamorphism. Unpublished Ph.D. thesis, Australian National University, 298 p.

Cross, A., Fletcher, I.R., Crispe, A.J., Huston, D.L., Williams, N., 2005. New constraints on the timing of deposition and mineralisation in the Tanami Group. Northern Territory Geological Survey Record 2005-001, pp. 185–192.

Cross, A., Jaireth, S., Rapp, R., Armstrong, R., 2011. Reconnaissance-style EPMA chemical U–Th–Pb dating of uraninite. Aust. J. Earth Sci. 58, 675–683.

Crouch, S., 1999. Geology, tectonic setting and metllogenesis of the Berserker Subprovince, northern New England Orogen. In: Flood, P.G. (Ed.), Regional Geology, Tectonics and Metallogenesis — New England Orogen. Department of Earth Sciences, University of New England, Armidale, New South Wales, Australia, pp. 233–249.

Cruden, A.R., McCaffrey, K.J.W., 2013. Scaling of tabular igneous intrusions in continental crust. Geol. Soc. Am. Annu. Gen. Meet. Abstr. Programs 86–92.

Cuney, M., 2005. World-class unconformity-related uranium deposits: Key factors for their genesis. In: Mao, J., Bierlein, F.P. (Eds.), Mineral Deposit Research: Meeting the Global Challenge. Springer Berlin Heidelberg, Heidelberg, pp. 245–248.

Currie, C.A., Beaumont, C., 2013. Are diamond-bearing Cretaceous kimberlites related to low-angle subduction beneath western North America? Earth Planet. Sci. Lett. 303, 59–70.

Czarnota, K., Champion, D.C., Goscombe, R., Blewett, S., Cassidy, K.F., Henson, P.A., Groenewald, P.B., 2010. Geodynamics of the eastern Yilgarn Craton. Precambrian Res. 183, 175–202.

Czarnota, K., Hoggard, M.J., White, N., Winterbourne, J., 2013. Spatial and temporal patterns of Cenozoic dynamic topography around Australia. Geochem. Geophys. Geosyst. 14, 634–658.

Dann, J.C., 1991. An Early Proterozoic ophiolite, central Arizona. Geology 19, 590–593.

DeCelles, P.G., Giles, K.A., 1996. Foreland basin systems. Basin Res. 8, 105–123.

DeCelles, P.G., Gehrels, G.E., Quade, J., Ojha, T.P., 1998. Eocene–early Miocene foreland basin development and the history of Himalayan thrusting, western and central Nepal. Tectonics 17, 741–765.

Defant, M.J., Drummond, M.S., 1990. Derivation of some modern arc magmas by melting of young subducted lithosphere. Nature 347, 662–665.
Dewey, J.F., Burke, K.C.A., 1973. Hot spots and continental breakup: implications for collisional orogeny. Geology 2, 57–60.
Dhuime, B., Hawkesworth, C.J., Cawood, P.A., Storey, C.D., 2012. A change in the geodynamics of continental growth 3 billion years ago. Science 335, 1334–1336.
Dickas, A., 1986. Comparative Precambrian stratigraphy and structure along the Mid-Continent Rift. AAPG Bull. 70, 225–238.
Dietz, R.S., 1961a. Continent and ocean basin evolution by spreading of the sea floor. Nature 190, 854–857.
Dietz, R.S., 1961b. Vredefort ring structure: meteorite impact scar. J. Geol. 69, 499–516.
Dietz, R.S., 1964. Sudbury structure as an astrobleme. J. Geol. 72, 412–434.
Downes, P.J., Demény, A., Czuppon, G., Jaques, A.L., Verrall, M., Sweetapple, M., Adams, D., McNaughton, N.J., Gwalani, L.G., Griffin, B.J., 2014. Stable H–C–O isotope and trace element geochemistry of the Cummins Range Carbonatite Complex, Kimberley region, Western Australia: implications for hydrothermal REE mineralization, carbonatite evolution and mantle source regions. Mineral. Deposita 49, 905–932.
Doyle, M.D., Allen, R., 2003. Subsea-floor replacement in volcanic-hosted massive sulfide deposits. Ore Geol. Rev. 23, 183–222.
Dubé, B., Mercier-Langevin, P., Hannington, M.D., Gosselin, G., Gosselin, P., 2007. The LaRonde Penna world-class Au-rich volcanogenic massive sulphide deposit, Abitibi, Quebéc: mineralogy and geochemistry of alteration and implications for genesis and exploration. Econ. Geol. 102, 633–666.
Dubé, B., Mercier-Langevin, P., Kjarsgaard, I., Hannington, M., Bécu, V., Côté, J., Moorhead, J., Legault, M., Bédard, N., 2014. The Bousquet 2-Dumagami world-class Archean Au-rich volcanogenic massive sulfide deposit, Abitibi, Quebec: metamorphosed submarine advanced argillic alteration footprint and genesis. Econ. Geol. 109, 121–166.
Eby, G.N., 1990. The A-type granitoids: A review of their occurrence and chemical characteristics and speculations on their petrogenesis. Lithos 26, 115–134.
Eisenlohr, B.N., Groves, D.I., Partington, G.A., 1989. Crustal scale shear zones and their significance to Archaean gold mineralisation in Western Australia. Mineral. Deposita 24, 1–8.
Elliot, D.H., Fleming, T.H., 2004. Occurrence and dispersal of magmas in the Jurassic Ferrar large igneous province, Antarctica. Gondwana Res. 7, 223–237.
Elston, D.P., Enkin, R.J., Baker, J., Kisilivsky, D.K., 2002. Tightening the belt: paleomagnetic-stratigraphic constraints on deposition, correlation, and deformation of the Middle Proterozoic (ca. 1.4 Ga) Belt-Purcell Supergroup, United States and Canada. Geol. Soc. Am. Bull. 114, 619–638.
Emsbo, P., Groves, D.I., Hofstra, A.H., Bierlein, F.P., 2006. The giant Carlin gold province: a protracted interplay of orogenic, basinal, and hydrothermal processes above a lithospheric boundary. Mineral. Deposita 41, 517–525.
Ernst, R.E., Baragar, W.R.A., 1992. Evidence from magnetic fabric for the flow pattern of magma in the Mackenzie giant radiating dyke swarm. Nature 356, 511–513.
Ernst, R.E., Bell, K., 2010. Large igneous provinces (LIPs) and carbonatites. Mineral. Petrol. 98, 55–76.
Ernst, R.E., Buchan, K.L., Campbell, I.H., 2005. Frontiers in large igneous province research. Lithos 79, 271–297.
Ernst, R.E., Jowitt, S.M., 2013. Large igneous provinces (LIPs) and metallogeny. Soc. Econ. Geol. Spec. Publ. 17, 17–51.
Eugster, H.P., Hardie, L.A., 1978. Saline lakes. In: Lerman, A. (Ed.), Lakes: Chemistry, Geology, Physics. Springer-Verlag, New York, pp. 237–293.
Farmer, G.L., 2003. Continental basaltic rocks. Treatise Geochem. 3, 1–39.
Figueiredo e Silva, R.C., Lobato, L.M., Rosiere, C.A., 2008. A hydrothermal origin for the jaspilite-hosted giant Sierra Norte deposits in the Cajajas Mineral Province, Para State, Brazil. Rev. Econ. Geol. 15, 255–290.
Finch, W.I., Davis, J.F., 1985. Sandstone-type uranium deposits — an introduction. Geological Environments of Sandstone-type Uranium Deposits. TECDOC-328, IAEA, Vienna, pp. 11–20.
Fiorentini, M.L., Beresford, S.W., Barley, M.E., Duuring, P., Bekker, A., Rosengren, N., Cas, R., Hronsky, J., 2012. District to camp controls on the genesis of komatiite-hosted nickel sulfide deposits, Agnew-Wiluna greenstone belt, Western Australia: insights from the multiple sulfur isotopes. Econ. Geol. 107, 781–796.
Foley, S.F., Link, K., Tiberindwa, J.V., Barifaijo, E., 2012. Patterns and origin of igneous activity around the Tanzanian craton. J. Afr. Earth Sci. 62, 1–18.
Franklin, J.M., Gibson, H.L., Jonasson, I.R., Galley, A., 2005. Volcanogenic massive sulfide deposits. Econ. Geol. 609–642 (100th Anniversary volume).
Franklin, J.M., Lydon, J.W., Sangster, D.M., 1981. Volcanic-associated Massive Sulfide Deposits. Econ. Geol. 485–627 (75th Anniversary volume).
Fraser, K.J., Hawkesworth, C.J., Erlank, A.J., Mitchell, R.H., Scott-Smith, B.H., 1985. Sr, Nd and Pb isotope and minor element geochemistry of lamproites and kimberlites. Earth Planet. Sci. Lett. 76, 57–70.
Galley, A.G., 1993. Characteristics of semi-conformable alteration zones associated with volcanogenic massive sulphide districts. J. Geochem. Explor. 48, 175–200.
Galley, A.G., 2003. Composite synvolcanic intrusions associated with Precambrian VMS-related hydrothermal systems. Mineral. Deposita 38, 443–473.
Garven, G., 1995. Continental-scale groundwater flow and geologic processes. Annu. Rev. Earth Planet. Sci. 24, 89–117.
Geological Survey of Queensland, 2011. North-West Queensland Mineral and Energy Province Report. The State of Queensland, Brisbane (123 pp.).
Gerya, T., 2014. Precambrian geodynamics: concepts and models. Gondwana Res. 25, 442–463.
Gibson, G.M., Henson, P.A., Neumann, N.L., Southgate, P.N., Hutton, L.J., 2012. Paleoproterozoic–earliest Mesoproterozoic basin evolution in the Mount Isa region, northern Australia and implications for reconstructions of the Nuna and Rodinia supercontinents. Episodes 35, 131–141.
Gibson, G.M., Morse, M.P., Ireland, T.R., Nayak, G.K., 2011. Arc–continent collision and orogenesis in western Tasmanides: insights from reactivated basement structures and formation of an ocean–continent transform boundary off western Tasmania. Gondwana Res. 19, 608–627.
Gill, J.B., 1981. Orogenic Andesites and Plate Tectonics. Springer-Verlag, Berlin.
Glen, R.A., 2013. Refining accretionary orogen models for the Tasmanides of eastern Australia. Aust. J. Earth Sci. 60, 315–370.
Goldfarb, R.J., Baker, T., Dubé, B., Groves, D.I., Hart, C.J.R., Gosselin, P., 2005. Distribution, character, and genesis of gold deposits in metamorphic terranes. Econ. Geol. 407–450 (100th Anniversary volume).
Goldfarb, R.J., Bradley, D., Leach, D.L., 2010. Secular Variation in Economic Geology. Econ. Geol. 105, 459–712.
Goldfarb, R.J., Groves, D.I., Gardoll, S., 2001. Orogenic gold and geologic time: a global synthesis. Ore Geol. Rev. 18, 1–75.
Goldfarb, R.J., Snee, L.W., Miller, L.D., Newberry, R.J., 1991. Rapid dewatering of the crust deduced from ages of mesothermal gold deposits. Nature 354, 296–298.
Goodfellow, W.D., 2007a. Base metal metallogeny of the Selwyn basin, Canada. Geol. Assoc. Can. Miner. Deposits Div. Spec. Publ. 5, 553–580.
Goodfellow, W.D., 2007b. Metallogeny of the Bathurst mining camp, northern New Brunswick. Geol. Assoc. Can. Miner. Deposits Div. Spec. Publ. 5, 449–469.
Goscombe, B., Blewett, R.S., Czarnota, K., Groenewald, P.B., Maas, R., 2009. Metamorphic evolution and integrated terrane analysis of the eastern Yilgarn Craton: rationale, methods, outcomes and interpretation. Geosci. Aust. Rec. 2007 (23), 270.
Graham, S., Lambert, D., Shee, S., 2004. The petrogenesis of carbonatite, melnoite and kimberlite from the Eastern Goldfields Province, Yilgarn Craton. Lithos 76, 519–533.
Greenfield, J.E., Musgrave, R.J., Bruce, M.C., Gilmore, P.J., Mills, K.J., 2011. The Mount Wright Arc: a Cambrian subduction system developed on the continental margin of East Gondwana, Koonenberry Belt, eastern Australia. Gondwana Res. 19, 650–669.
Griffin, W.L., Begg, G.C., O'Reilly, S.Y., 2013. Continental-root control on the genesis of magmatic ore deposits. Nat. Geosci. 6, 905–910. http://dx.doi.org/10.1038/ngeo1954.
Griffin, W.L., Belousova, E.A., O'Neill, C., O'Reilly, S.Y., Malkovets, V., Pearson, N.J., Spetsius, S., Wilde, S.A., 2014. The world turns over: Hadean–Archean crust–mantle evolution. Lithos 189, 2–15.
Gromet, L.P., Silver, L.T., 1983. REE variations across the Peninsular Ranges Batholith: implications for batholithic petrogenesis and crustal growth in magmatic arcs. J. Petrol. 28, 75–125.
Gromet, P., Silver, L.T., 1987. REE variations across the Peninsular Ranges batholith: Implications for batholithic petrogenesis and crustal growth in magmatic arcs. J. Petrol. 28, 75–125.
Gross, G.A., 1980. A classification of iron formations based on depositional environments. Can. Mineral. 18, 215–222.
Gross, G.A., 1993. Industrial and genetic models for iron ore in iron-formations. Geol. Assoc. Can. Spec. Pap. 40, 151–170.
Groves, D.I., 1993. The crustal continuum model for late-Archaean lode-gold deposits of the Yilgarn Block, Western Australia. Mineral. Deposita 28, 366–374.
Groves, D.I., Bierlein, F.P., 2007. Geodynamic settings of mineral deposit systems. J. Geol. Soc. 164, 19–30.
Groves, D.I., Bierlein, F.P., Meinert, L.D., Hitzman, M.W., 2010. Iron oxide copper–gold (IOCG) deposits through Earth history: implications for origin, lithospheric setting, and distinction from other epigenetic iron oxide deposits. Econ. Geol. 105, 641–654.
Groves, D.I., Condie, K.C., Goldfarb, R.J., Hronsky, J.M., Vielreicher, R.M., 2005. 100th anniversary special paper: secular changes in global tectonic processes and their influence on the temporal distribution of gold-bearing mineral deposits. Econ. Geol. 100, 203–224.
Groves, D.I., Goldfarb, R.J., Gebre-Mariam, M., Hagemann, S.G., Robert, F., 1998. Orogenic gold deposits: a proposed classification in the context of their crustal distribution and relationship to other gold deposit types. Ore Geol. Rev. 13, 7–27.
Gudmundsson, A., 1983. Form and dimensions of dykes in Eastern Iceland. Tectonophysics 95, 295–307.
Gustafson, L.B., Williams, N., 1981. Sediment-hosted deposits of copper, lead, and zinc. Econ. Geol. 139–178 (75th Anniversary volume).
Hackett, J.P., Bischoff, J.L., 1973. New data on the stratigraphy, extent, and geologic history of the Red Sea geothermal deposits. Econ. Geol. 68, 553–564.
Haddad, D., Watts, A.B., Lindsay, J., 2001. Evolution of the intracratonic Officer Basin, central Australia: implications from subsidence analysis and gravity modelling. Basin Res. 13, 217–238.
Hagemann, S.G., Cassidy, K., 2000. Archean orogenic lode gold deposits. In: Hagemann, S.G., Brown, P.E. (Eds.), Gold in 2000, SEG Reviews 13, pp. 9–68.
Hagemann, S.G., Angerer, T., Duuring, P., Rosiere, C.A., Figueiredo e Silva, R.C., Lobato, L., Hensler, A.S., Walde, D.H.G., 2016. BIF-hosted iron mineral system: a review. Ore. Geol. Rev. 76, 317–359 (in this issue).
Hagemann, S.G., Gilg, H.A., Fernando, et al., 2011. Oxygen and hydrogen isotope study of the Achaean granulite-hosted Griffin's Find gold deposit, Western Australia. Proceedings of the 11th Biennial Meeting of the Society for Geology Applied to Mineral Deposits, SGA, Antofagasta, Chile, 26-29th September 2011, pp. 557–559.
Hagemann, S.G., Mavrogenes, J., McCuaig, C., Cassidy, K., Behn, M., 2007. The genesis of the Au–Cu–Mo Boddington gold mine: constraints based on a melt and fluid inclusions study. In: Cassidy, K.F., Hronsky, J. (Eds.), Conference Proceedings, Kalgoorlie '07, 25–27 September 2007. Geoconferences (WA) INC, pp. 23–26.
Haines, P.W., Wingate, M.T.D., 2007. Contrasting depositional histories, detrital zircon provenance and hydrocarbon systems: did the larapintine Seaway link the Canning and Amadeus basins during the Ordovician? North. Territ. Geol. Surv. Spec. Publ. 2, 36–51.
Hand, M., Sandiford, M., 1999. Intraplate deformation in central Australia, the link between subsidence and fault reactivation. Tectonophysics 305, 121–140.

Hannington, M.D., de Ronde, C.E.J., Petersen, S., 2005. Sea-floor tectonics and submarine hydrothermal systems. Econ. Geol. 111–142 (100th Anniversary volume).

Hannington, M., Poulsen, H., Thompson, J., Sillitoe, R., 1999. Volcanogenic gold and epithermal-style mineralization in the VMS environment. Rev. Econ. Geol. 8, 325–356.

Hansen, D.A., Herinrichs Jr., W.E., Holmer, R.C., MacDougall, R.E., Rogers, G.R., Sumner, J.S., Ward, S.E. (Eds.), 1966. Society of Exploration Geophysicists' Mining Geophysics; Volume I, Case Histories. Society of Exploration Geophysicists, Tulsa (492 pp.).

Hart, H., Gibson, H.L., Lesher, C.M., 2004. Trace element geochemistry and petrogenesis of felsic volcanic rocks associated with volcanogenic Cu–Zn–Pb massive sulfide deposits. Econ. Geol. 99, 1003–1013.

Hassan, L.Y., 2000. Mineral occurrences and exploration potential of the East Kimberly. Geol. Surv. West. Aust. Rep. 74 (83 pp.).

Hay, R.L., 1989. Holocene carbonatite–nephelinite tephra deposits of Oldoinyo Lengai, Tanzania. J. Volcanol. Geotherm. Res. 37, 77–91.

Haynes, D.W., Cross, K.C., Bills, R.T., Reed, M.H., 1995. Olympic Dam ore genesis: a fluid mixing model. Econ. Geol. 90, 281–307.

Hayward, N., Skirrow, R.G., 2010. Geodynamic setting and controls on iron oxide Cu–Au(±U) ore in the Gawler Craton, South Australia. In: Porter, T.M. (Ed.), Hydrothermal Iron Oxide Copper–Gold and Related Deposits: A Global Perspective, Volume 3, Advances in the Understanding of IOCG Deposits. PGC Publishing, Adelaide, pp. 105–131.

Heaman, L.M., Kjarsgaard, B.A., 2000. Timing of eastern North American kimberlite magmatism: continental extension of the Great Meteor hotspot track? Earth Planet. Sci. Lett. 178, 253–268.

Heaman, L.H., LeCheminant, A.N., Rainbird, R.H., 1992. Nature and timing of Franklin igneous events, Canada: implications for a late Proterozoic mantle plume and the break-up of Laurentia. Earth Planet. Sci. Lett. 109, 117–131.

Heinrich, C.A., Andrew, A.S., Wilkins, R.W.T., Patterson, D.L., 1989. A fluid inclusion and stable isotope study of synmetamorphic copper ore formation at Mt Isa, Australia. Econ. Geol. 84, 529–550.

Heinrich, C.A., Bain, J.H.C., Fardy, J.J., Waring, C.L., 1993. Bromine/chlorine geochemistry of hydrothermal brines associated with Proterozoic metasediment-hosted copper mineralization at Mount Isa, northern Australia. Geochim. Cosmochim. Acta 57, 2991–3000.

Heinson, G.S., Direen, N.G., Gill, R.M., 2006. Magnetotelluric evidence for a deep-crustal mineralizing system beneath the Olympic Dam iron oxide copper–gold deposit, southern Australia. Geology 34, 573–576.

Hergt, J.M., Chappell, B.W., McCulloch, M.T., McDougall, I., Chivas, A.R., 1989. Geochemical and isotopic constraints on the origin of the Jurassic dolerites of Tasmania. J. Petrol. 30, 841–883.

Herzberg, C., Asimow, P.D., Arndt, N., Niu, Y., Lesher, C.M., Fitton, J.G., Cheadle, M.J., Saunders, A.D., 2007. Temperatures in ambient mantle and plumes: constraints from basalts picrites and komatiites. Geochem. Geophys. Geosyst. 8, Q02006. http://dx.doi.org/10.1029/2006GC001390.

Herzberg, C., Condie, K., Korenaga, J., 2010. Thermal history of the Earth and its petrological expression. Earth Planet. Sci. Lett. 292, 79–88.

Herzig, P.M., Petersen, S., Hannington, M.D., 2002. Polymetallic massive sulphide deposits at the modern seafloor and their resource potential. ISA Technical Study 2. International Seabed Authority, Kingston Jamaica (116 pp.).

Hess, H.H., 1962. History of ocean basins. In: Engel, A.E.J., James, H.L., Leonard, B.F. (Eds.), Petrologic Studies: A Volume to Honor A.F. Buddington. Geological Society of America, pp. 599–620.

Hildreth, W.E., 1981. Gradients in silicic magma chambers: implications for lithospheric magmatism. J. Geophys. Res. 86, 10153–10192.

Hitzman, M.W., Valenta, R.K., 2005. Uranium in iron oxide–copper–gold (IOCG) systems. Econ. Geol. 100, 1657–1661.

Hitzman, M., Kirkham, R., Broughton, D., Thorson, J., Selley, D., 2005. The sediment-hosted stratiform copper ore system. Econ. Geol. 609–642 (100th Anniversary volume).

Hitzman, M.W., Oreskes, N., Einaudi, M.T., 1992. Geological characteristics and tectonic setting of Proterozoic iron oxide (Cu–U–Au–LREE) deposits. Precambrian Res. 58, 241–287.

Hoatson, D.M., Jaireth, S., Jaques, A.L., 2006. Nickel sulfide deposits in Australia: characteristics, resources, and potential. Ore Geol. Rev. 29, 177–241.

Hoatson, D.M., Jaireth, S., Miezitis, Y., 2011. The major rare-earth-element deposits of Australia: geological setting, exploration, and resources. Geosci. Aust. (204 pp.).

Holmes, A., 1929. Radioactivity and earth movements. Trans. Geol. Soc. Glasgow 18, 559–606.

Holt, S.J., Holford, S.P., Foden, J., 2014. New insights into the magmatic plumbing system of the South Australian Quaternary Basalt province from 3D seismic and geochemical data. Aust. J. Earth Sci. 60, 797–817.

Hou, Z., Ma, H., Zaw, K., Z, Y., Wang, M., Wang, Z., Pan, G., Tang, R., 2003. The Himalayan Yulong porphyry copper belt: product of large-scale strike-slip faulting in eastern Tibet. Econ. Geol. 98, 125–145.

Hronsky, J.M.A., Schodde, R.C., 2006. Nickel exploration history of the Yilgarn Craton: from the nickel boom to today. Soc. Econ. Geol. Spec. Publ. 13, 1–11.

Hughes, F.E. (Ed.), 1990. Geology of the Mineral Deposits of Australia and Papua New Guinea 14. Australasian Institute of Mining and Metallurgy Monograph (1828 pp.).

Huleatt, M.B., Jaques, A.L., 2005. Australian gold exploration 1976–2003. Resour. Policy 30, 29–37.

Huston, D.L., Taylor, T., Fabray, J., Patterson, D.J., 1992. A comparison of the geology and mineralization of the Balcooma And Dry River South volcanic-hosted massive sulfide deposits, northern Queensland. Econ. Geol. 87, 785–811.

Huston, D.L., Kamprad, J., 2001. Zonation of alteration facies at Western Taarsis: implications for the genesis of Cu–Au deposits, Mount Lyell field, western Tasmania. Econ. Geol. 96, 1123–1132.

Huston, D.L., Stevens, B., Southgate, P.N., Muhling, P., Wyborn, L., 2006. Australian Zn–Pb–Ag ore-forming systems: a review and analysis. Econ. Geol. 101, 1117–1158.

Huston, D.L., Maas, R., Czarnota, K., 2007a. The age and genesis of the Nifty copper deposit: back to the future. Geosci. Aust. Prof. Opin. 2007/03 (26 pp.).

Huston, D.L., Morant, P., Pirajno, F., Cummins, B., Baker, D., Mernagh, T.P., 2007b. Paleoarchean mineral deposits of the Pilbara Craton: genesis, tectonic environment and comparisons with younger deposits. Dev. Precambrain Geol. 15, 411–450.

Huston, D.L., Vandenberg, L., Wygralak, A.S., Mernagh, T.P., Bagas, L., Crispe, A., Lambeck, A., Cross, A., Fraser, G., Williams, N., Worden, K., Meixner, T., Goleby, B., Jones, L., Lyons, P., Maidment, D., 2007c. Lode–gold mineralization in the Tanami region, northern Australia. Mineral. Deposita 42, 175–204.

Huston, D.L., Czarnota, K., Jaireth, S., Williams, N.C., Maidment, D., Cassidy, K.F., Duerden, P., Miggins, D., 2010a. Mineral systems of the paterson region. Geosci. Aust. Rec. 2010 (12), 155–218.

Huston, D.L., Pehrsson, S., Eglington, B.M., Zaw, K., 2010b. The geology and metallogeny of volcanic-hosted massive sulfide deposits through geologic time and with tectonic setting. Econ. Geol. 105, 571–591.

Huston, D.L., Relvas, J.M.R.S., Gemmell, J.B., Drieberg, S., 2011. The role of granites in volcanic-hosted massive sulphide ore-forming systems: an assessment of magmatic–hydrothermal contributions. Mineral. Deposita 46, 473–507.

Huston, D.L., Blewett, R.S., Champion, D.C., 2012. Australia through time: a summary of its tectonic and metallogenic evolution. Episodes 35, 23–43.

Huston, D.L., Champion, D.C., Cassidy, K.F., 2014. Tectonic controls on the endowment of Neoarchean cratons in volcanic-hosted massive sulfide deposits: evidence from lead and neodymium isotopes. Econ. Geol. 109, 11–26.

Huston, D.L., Eglington, B.M., Pehrsson, S., Piercey, S.J., 2015. The metallogeny of zinc through time: links to secular changes in the atmosphere, hydrosphere, and the supercontinent cycle. In: Archibald, S.M., Piercey, S.J. (Eds.), Current Perspectives on Zinc Deposits. Irish Association for Economic Geology, Dublin, pp. 1–16.

Huston, D.L., Champion, D.C., Mernagh, T.P., Downes, P.M., Jones, P., Carr, G., Forster, D., David, V., 2016. Metallogenesis and geodynamics of the Lachlan Orogen: new (and old) insights from spatial and temporal variations in lead isotopes. Ore. Geol. Rev. 76, 257–267 (in this issue).

Hutchinson, R.W., 1973. Volcanogenic massive sulfide deposits and their metallogenic significance. Econ. Geol. 68, 1223–1246.

Hutchinson, D.R., Golmshtok, A.J., Zonenshain, L.P., Moore, T.C., Scholz, C.A., Klitgord, K.D., 1998. Depositional and tectonic framework of the rift basins of Lake Baikal from multichannel seismic data. Geology 20, 589–592.

Hutton, L.J., Denaro, T.J., Dhnaram, C., Derrick, G., 2012. Mineral systems in the Mount Isa Inlier. Episodes 35, 120–130.

Idnurm, M., 2000. Toward a high resolution Late Palaeoproterozoic–Earliest Mesoproterozoic apparent polar wander path for northern Australia. Aust. J. Earth Sci. 47, 405–430.

Ireland, T., Bull, S.W., Large, R.R., 2004. Mass flow sedimentology within the HYC Zn–Pb–Ag deposit, Northern Territory, Australia: evidence for syn-sedimentary ore genesis. Mineral. Deposita 39, 143–158.

Ishihara, S., 1981. The granitoid series and mineralization. Econ. Geol. 458–484 (75th Anniversary volume).

Ivanic, T.J., Van Kraenndonk, M.J., Kirkland, C.K., Wyche, S., Wingate, M.T.D., Belousova, E.A., 2012. Zircon Lu–Hf isotopes and granite geochemistry of the Murchison domain of the Yilgarn Craton: evidence for reworking of Eoarchean crust during Meso-Neoarchean plume-driven magmatism. Lithos 148, 112–127.

Jackson, D.G., Andrew, R.L., 1990. Kintyre uranium deposit. Austr. Inst. Mining Metall. Mon. 14, 653–658.

Jaireth, S., Hoatson, D.M., Miezitis, Y., 2014. Geological setting and resources of the major rare-earth-element deposits in Australia. Ore Geol. Rev. 62, 72–128.

Jaireth, J., Roach, I.R., Bastrakov, E., Liu, S., 2016. Basin-related uranium mineral systems in Australia: a review of critical features. Ore. Geol. Rev. 76, 360–394 (in this issue).

Jaques, A.L., Milligan, P.R., 2004. Patterns and controls on the distribution of diamondiferous intrusions in Australia. Lithos 77, 783–802.

Jaques, A.L., Chappell, B.W., Sun, S.-S., 1989. Geochemistry of the Argyle (AK1) lamproite pipe, Western Australia. Geol. Soc. Aust. Spec. Publ. 14 (2), 171–188.

Jaques, A.L., Huleatt, M.B., Ratajkoski, M., Towner, R.R., 2005. Exploration and discovery of Australia's copper, nickel, lead and zinc resources 1976–2004. Resour. Policy 30, 168–185.

Jaques, A.L., Jaireth, S., Walshe, J.L., 2002. Mineral systems of Australia; an overview of resources, settings and processes. Aust. J. Earth Sci. 49, 623–660.

Jaques, A.L., Lewis, J.D., Smith, C.B., 1986. The kimberlites and lamproites of Western Australia. Geol. Surv. West. Aust. Bull. 132 (268 pp.).

Jelsma, H., Barnett, W., Richards, S., Lister, G., 2009. Tectonic setting of kimberlites. Lithos 112S, 155–165.

John, E.G., 1978. Mineral zones in the Utah copper ore body. Econ. Geol. 73, 1250–1259.

Johnson, J.P., McCulloch, M.T., 1995. Sources of mineralizing fluids for the Olympic Dam deposit, (South Australia): Sm–Nd isotopic constraints. Chem. Geol. 121, 177–199.

Johnson, T.E., Brown, M., Kaus, B., VanTongeren, J.A., 2014. Delamination and recycling of Archean crust caused by gravitational instabilities. Nat. Geosci. 7, 47–52.

Jung, S., Mezger, K., Hauff, F., Pack, A., Hoernes, S., 2013. Petrogenesis of rift-related tephrites, phonolites and trachytes (Central European Volcanic Province, Rhön, FRG): constraints from Sr, Nd, Pb and O isotopes. Chem. Geol. 354, 203–215.

Kalogeropoulos, S.I., Scott, S.D., 1983. Minerology and geochemistry of tuffaceous exhalites (tetsusekiei) of the Fukazawa mine, Hokuroku district, Japan. Econ. Geol. Monogr. 5, 412–432.

Kelley, K.A., Cottrell, E., 2009. Water and the oxidation state of subduction zone magmas. Science 325, 605–607.

Kelley, K.D., Leach, D.L., Johnson, C.A., Clark, J.L., Fayek, M., Slack, J.F., Anderson, V.M., Ayuso, R.O., Ridley, W.I., 2004. Textural, compositional, and sulfur isotope variations of sulfide minerals in the Red Dog Zn–Pb–Ag deposits, Brooks Range, Alaska: implications for ore formation. Econ. Geol. 99, 1509–1532.

Kennard, J.M., Jackson, M.J., Romine, K.K., Shaw, R.D., Southgate, P.N., 1994. Depositional sequences and associated petroleum systems of the Canning Basin, W.A. In: Purcell, P.G., Purcell, R.R. (Eds.), The Sedimentary Basins of Western Australia: Proceedings of the West Australia Basins Symposium, Perth, August, pp. 657–676.

Kerr, A.C., Mahoney, J.J., 2007. Oceanic plateaus: problematic plumes, potential paradigms. Chem. Geol. 241, 332–353.

Kerrich, R., Wyman, D., 1990. Geodynamic setting of mesothermal gold deposits: an association with accretionary tectonic cycles. Geology 18, 882–885.

Kerrich, R., Wyman, D., 1994. The mesothermal gold–lamproite association: significance for an accretionary geodynamic setting, supercontinent cycles, and metallogenic processes. Mineral. Petrol. 51, 147–172.

Kerrich, R., Goldfarb, R.J., Groves, D.I., Garwin, S., 2000. The geodynamics of world class ore deposits: characteristics, space-time distribution, and origins. Rev. Econ. Geol. 13, 501–551.

Kerrich, R., Goldfarb, R.J., Richards, J.P., 2005. Metallogenic provinces in an evolving geodynamic framework. Econ. Geol. 1097–1139 (100th Anniversary volume).

Kesler, S.E., 1997. Metallogenic evolution of convergent margins: selected ore deposit models. Ore Geol. Rev. 153-171.

Kirkham, R.V., 1989. Distribution, setting and genesis of sediment-hosted copper deposits. Geol. Assoc. Can. Spec. Pap. 36, 3–36.

Klein, E.M., 2003. Geochemistry of the igneous oceanic crust. Treatise Geochem. 3, 433–463.

Klein, C., 2005. Some Precambrian banded iron-formations (BIFs) from around the world: their age, geologic setting, mineralogy, metamorphism, geochemistry, and origin. Am. Mineral. 90, 1473–1499.

Klein, C., Beukes, N.J., 1993. Sedimentology and geochemistry of the glaciogenic late Proterozoic rapitan iron-formation in Canada. Econ. Geol. 88, 542–565.

Klein, E.M., Langmuir, C.H., 1987. Global correlations of ocean ridge basalt chemistry with axial depth and crustal thickness. J. Geophys. Res. 92, 8089–8115.

Knight, C.L. (Ed.), 1975. Economic Geology of Australia and Papua New Guinea. Australas. Inst. Min. Metall. Monogr. 5, 1126.

Knox-Robinson, C.M., Wyborn, L.A.I., 1997. Towards a holistic exploration strategy: using geographic information systems as a tool to enhance exploration. Aust. J. Earth Sci. 44, 453–463.

Komiya, T., Maruyama, S., Nohda, S., Masuda, M., Hayashi, H., Okamoto, S., 1999. Plate tectonics at 3.8–3.7 Ga: field evidence from the Isua accretionary complex, southern Greenland. The. J. Geol. 107, 515–554.

Korenaga, J., 2013. Initiation and evolution of plate-tectonics on Earth: theories and observations. Annu. Rev. Earth Planet. Sci. 41, 117–151.

Korsch, R.J., Doublier, M.P., 2014. Major Crustal Boundaries of Australia. first ed. Geoscience Australia, Canberra.

Korsch, R.J., Doublier, M., 2016. Major crustal boundaries of Australia, and their significance in mineral systems targeting. Ore. Geol. Rev. 76, 211–228 (in this issue).

Korsch, R.J., Kositcin, N., Champion, D.C., 2011. Australian island arcs through time: geodynamic implications for the Archean and Proterozoic. Gondwana Res. 19, 716–734.

Krapež, B., 2007. Late-stage basins of the Eastern Goldfields Superterrane. In: Bierlein, F.P., Knox-Robinson, C.M. (Eds.), Proceedings of Geoconferences (WA) Inc. Kalgoorlie '07 Conference, 25–27 September 2007. Kalgoorlie, Western Australia. Geosci. Aust. Rec. 2007 (14), pp. 44–45.

Krapež, B., Barley, M.E., Brown, S.J.A., 2008a. Late Archaean synorogenic basins of the Eastern Goldfields Superterrane, Yilgarn Craton, Western Australia part I. Kalgoorlie and Gindalbie Terranes. Precambrian Res. 161, 135–153.

Krapež, B., Standing, J.G., Brown, S.J.A., Barley, M.E., 2008b. Late Archaean synorogenic basins of the Eastern Goldfields Superterrane, Yilgarn Craton, Western Australia part II. Kurnalpi Terrane. Precambrian Res. 161, 154–182.

Kreuzer, O.P., 2005. Intrusion-hosted mineralization in the Charters Towers goldfield, north Queensland: new isotopic and fluid inclusion constraints on the timing and origin of the auriferous veins. Econ. Geol. 100, 1583–1604.

Kyser, T.K., 2007. Fluids, basin analysis, and mineral deposits. Geofluids 7, 238–257.

Kyser, K., Cuney, M., 2008. Unconformity-related uranium deposits. Mineralogical Association of Canada Short Course Handbook. 37, pp. 161–219.

Kyser, T.K., Hiatt, E.E., Renac, C., Durocher, K., Holk, G., Deckart, K., 2000. Diagenetic fluids in paleo- and meso-Proterozoic sedimentary basins and their implications for long protracted fluid histories. Mineralogical Association of Canada Short Course Handbook. 37, 225–262.

Lang, J.R., Baker, T., Hart, C.J.R., Mortensen, J.K., 2000. An exploration model for intrusion-related gold systems. Soc. Econ. Geol. Newsl. 40.

Langmuir, C.H., Klein, E.M., Plank, T., 1992. Petrological systematics of mid-ocean ridge basalts: constraints on melt generation beneath ocean ridges. Am. Geophys. Union Geophys. Monogr. 71, 183–280.

Large, R.R., Bull, S.W., Cooke, D.R., McGoldrick, P.J., 1998. A genetic model for the HYC deposit, Australia: based on regional sedimentology, geochemistry, and sulfide-sediment relationships. Econ. Geol. 93, 1345–1368.

Large, R., Doyle, M., Raymond, O., Cooke, D., Jones, A., Heasman, L., 1996. Evaluation of the role of Cambrian granites in the genesis of world class VHMS deposits in Tasmania. Ore Geol. Rev. 10, 215–230.

Lawrie, K.C., Hinman, M.C., 1998. Cobar-style polymetallic Au–Cu–Ag–Pb–Zn deposits. AGSO J. Aust. Geol. Geophys. 17, 169–187.

Leach, D.L., Bradley, D.C., Huston, D.L., Pisarevsky, S.A., Taylor, R.D., Gardoll, S.J., 2010. Sediment-hosted lead–zinc deposits in Earth history. Econ. Geol. 105, 593–625.

Leach, D.L., Bradley, D., Lewchuk, M.T., Symons, D.T.A., de Marsily, G., Brannon, J., 2001. Mississippi Valley-type lead–zinc deposits through geologic time — implications from recent age-dating research. Mineral. Deposita 36, 711–740.

Leach, D.L., Landis, G.P., Hofstra, A.H., 1988. Metamorphic origin of the Couer d'Alene base and precious metal veins in the Belt basin, Idaho and Montana. Geology 16, 122–125.

Leach, D.L., Sangster, D.F., Kelley, K.D., Large, R.R., Garven, G., Allen, C.R., Gutzmer, J., Walters, S., 2005. Sediment-hosted lead–zinc deposits: a global perspective. Econ. Geol. 561–608 (100th Anniversary volume).

Lentz, D.R., 1998. Petrogenetic evolution of felsic volcanic sequences associated with Phanerozoic volcanic-hosted massive sulfide systems: the role of extensional geodynamics. Ore Geol. Rev. 12, 289–327.

Lesher, C.M., Goodwin, A.M., Campbell, I.H., Gorton, M.P., 1986. Trace-element geochemistry of ore-associated and barren, felsic metavolcanic rocks in the superior province, Canada. Can. J. Earth Sci. 23, 222–237.

Lewis, S.E., Henderson, R.A., Dickens, G.R., Shields, G.A., Coxhell, S., 2010. The geochemistry of primary and weathered oil shale and coquina across the Julia Creek vanadium deposit (Queensland, Australia). Mineral. Deposita 45, 599–620.

Li, Z.X., Bogdanova, S., Collins, A.S., Davidson, A., DeWaele, B., Ernst, R.E., Fitzsimons, I.C.W., Fuck, R.A., Gladkochub, D.P., Jacobs, J., Karlstrom, K.E., Lu, S., Natapov, L.M., Pease, V., Pisarevsky, S.A., Thrane, K., Vernikovsky, V., 2008. Assembly, configuration, and break-up history of Rodinia; a synthesis. Precambrian Res. 160, 179–210.

Li, C., Lightfoot, P.C., Amelin, Y., Naldrett, A.J., 2007. Contrasting petrological and geochemical relationships in the Voisey's Bay and Mushuau intrusions, Labrador, Canada: implications for ore genesis. Econ. Geol. 95, 771–799.

Lindgren, V., 1933. Mineral deposits. fourth ed. McGraw-Hill, New York (930 pp.).

Lister, G.S., Davis, G.A., 1989. The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the northern Colorado River region, U.S.A. J. Struct. Geol. 11, 65–94.

Lister, G., Forster, M., 2009. Tectonic mode switches and the nature of orogenesis. Lithos 113, 274–291.

Lombaard, A.F., Gunzel, A., Innes, J., Kruger, T.L., 1986. The Tsumeb lead-copper-zinc-silver deposit, south west Africa/Namibia. In: Anhaeusser, C.R., Maske, S. (Eds.), Mineral Deposits of South Africa. Geological Society of South Africa, Johannesburg, pp. 1761–1787.

Lottermoser, B., 1990. Rare-earth element mineralisation within the Mt. Weld carbonatite laterite, Western Australia. Lithos 24, 151–167.

Loutit, T.S., Wyborn, L.A.I., Hinman, M.C., Idnurm, M., 1994. Palaeomagnetic, tectonic, magmatic and mineralisation events in the Proterozoic of northern Australia. Aust. Inst. Min. Metall. Publ. Ser. 5 (94), 123–128.

Lowell, J.D., Guilbert, J.M., 1970. Lateral and vertical alteration–mineralization zoning in porphyry ore deposits. Econ. Geol. 65, 373–408.

Luddington, S., Plumlee, G.S., 2009. Climax-type porphyry molybdenum deposits. US Geol. Surv. 16 (Open-File Report 2009–1215).

Lydon, J.W., 2007. Geology and metallurgy of the Belt-Purcell Basin. Geol. Assoc. Can. Miner. Deposits Div. Spec. Publ. 5, 581–608.

Lyons, P., Goleby, B.R., (compilers), 2005. The 2003 Gawler Craton seismic survey; notes from the seismic workshop. Geosci. Aust. Rec. 2005 (19), 81.

Maas, R., 1989. Nd–Sr isotope constraints on the age and origin of unconformity-type uranium deposits in the Alligator Rivers uranium field, Northern Territory, Australia. Econ. Geol. 84, 64–90.

Machel, H.G., 2001. Bacterial and thermochemical sulfate reduction in diagenetic settings — old and new insights. Sediment. Geol. 140, 143–175.

Magoon, L.B., Dow, W.G., 1994. The petroleum system. AAPG Mem. 60, 3–24.

Maier, W.D., Barnes, S.J., Campbell, I.H., Fiorentini, M.L., Peltonen, P., Barnes, S.-J., Smithies, H., 2009. Mantle magmas reveal progressive mixing of meteoritic veneer into the early Earth's deep mantle. Nature 460, 620–623.

Majorowicz, J.A., Jones, F.W., Lam, H.J., Jessop, A.M., 1984. The variability of heat flow both regional and with depth in southern Alberta, Canada: effect of groundwater flow? Tectonophysics 106, 1–29.

Martin, H., Smithies, R.H., Rapp, R., Moyen, J.-F., Champion, D., 2005. An overview of adakite, tonalite–trondhjemite–granodiorite (TTG), and sanukitoid: relationships and some implications for crustal evolution. Lithos 79, 1–24.

Maruyama, S., Liou, J.G., 1998. Initiation of ultrahigh-pressure metamorphism and its significance on the Proterozoic–Phanerozoic boundary. Island Arc 7, 6–35.

Mathur, R., Marschik, R., Ruiz, J., Munizaga, F., Leveille, R.A., Martin, W., 2002. Age of mineralization of the Candelaria iron oxide Cu–Au deposit, and the origin of the Chilean Iron Belt based on Re–Os isotopes. Econ. Geol. 97, 59–71.

Matthews, K.J., Hale, A.J., Gurnis, M., Muller, R.D., DiCaprio, L., 2011. Dynamic subsidence of eastern Australia during the Cretaceous. Gondwana Res. 19, 372–383.

McClusky, S., Reilinger, R., Ogubazghi, G., Amleson, A., Healab, B., Vernant, P., Sholan, J., Fisseha, S., Asfaw, L., Bendick, R., Kogan, L., 2010. Kinematics of the southern Red Sea–Afar Triple Junction and implications for plate dynamics. Geophys. Res. Lett. 37, L05301. http://dx.doi.org/10.1029/2009GL041127.

McCracken, S.R., Etminan, H., Connor, A.G., Williams, V.A., 1996. Geology of the Admiral Bay carbonate-hosted zinc–lead deposit, Canning Basin, Western Australia. Soc. Econ. Geol. Spec. Publ. 4, 330–349.

McCuaig, T.C., Hronsky, J.M.A., 2014. The mineral system concept: the key to exploration targeting. Soc. Econ. Geol. Spec. Publ. 18, 153–176.

McCuaig, T.C., Kerrich, R., 1998. P–T–t–deformation—fluid characteristics of lode gold deposits: evidence from alteration systematics. Ore Geol. Rev. 12, 381–453.

McCuaig, T.C., Behn, M., Stein, H., Hagemann, S.G., McNaughton, N.J., Cassidy, K.F., Champion, D.C., Wyborn, L., 2001. The Boddington gold mine: a new style of Archaean Au–Cu deposit. Geosci. Aust. Rec. 2001 (37), 453–455.

McCuaig, T.C., Beresford, S., Hronsky, J., 2010. Translating the mineral systems approach into an effective exploration targeting system. Ore Geol. Rev. 38, 128–138.

McCulloch, M.T., Gamble, J.A., 1991. Geochemical and geodynamical constraints on subduction zone magmatism. Earth Planet. Sci. Lett. 102, 358–374.

McDonough, W.F., 2003. Compositional model for the Earth's core. Treatise Geochem. 2, 547–568.

McKay, A.D., 1990. Evaluation of the uranium mineralisation and uranium resource potential at Coronation Hill. In: Curtis, R.E.S. (Ed.), Assessment of the Identified Resource

at Coronation Hill and its Possible Extensions, BMR Geological and Mineral Resource Studies Resource Assessment Commission Kakadu Conservation Zone Inquiry Consultancy Series. Australian Government Publishing Service, Canberra.
McKenzie, D., 1978. Some remarks on development of sedimentary basins. Earth Planet. Sci. Lett. 40, 25–32.
McKenzie, D., Bickle, M.J., 1988. The volume and composition of melt generated by extension of the lithosphere. J. Petrol. 29, 623–679.
Menand, T., 2011. Physical controls and depth of emplacement of igneous bodies: a review. Tectonophysics 500, 11–19.
Mercier-Langevin, P., Dubé, B., Hannington, M.D., Davis, D.W., Lafrance, B., Gosselin, G., 2007a. The LaRonde Penna Au-rich volcanogenic massive sulphide deposit, Abitibi greenstone belt, Quebec: part I. Geology and geochronology. Econ. Geol. 102, 585–609.
Mercier-Langevin, P., Dubé, B., Hannington, M.D., Richer-Laflèche, M., Gosselin, G., 2007b. The LaRonde Penna Au-rich volcanogenic massive sulfide deposit, Abitibi Greenstone Belt, Quebec: part II. Lithogeochemistry and paleotectonic setting. Econ. Geol. 102, 611–631.
Mercier-Langevin, P., Gibson, H.L., Hannington, M.D., Goutier, J., Monecke, T., Dubé, B., Houlé, M.G., 2014. A special issue on Archean magmatism, volcanism, and ore deposits: part 2. Volcanogenic massive sulfide deposits preface. Econ. Geol. 109, 1–9.
Mercier-Langevin, P., Hannington, M.D., Dubé, B., Bécu, V., 2011. The gold content of volcanogenic massive sulfide deposits. Mineral. Deposita 46, 509–539.
Mernagh, T.P., Wygralak, A.S., 2011. A fluid inclusion study of uranium and copper mineral systems in the Murphy Inlier, Northern Australia. Russ. Geol. Geophys. 52, 1421–1435.
Mernagh, T.P., Bastrakov, E.N., Clarke, J.D.A., de Caritat, P., English, P.M., Howard, F.J.F., Jaireth, S., Magee, J.W., McPherson, A.A., Roach, I.C., Schroder, I.F., Thomas, M., Wilford, J.R., 2013. A Review of Australian Salt Lakes and Assessment of Their Potential for Strategic Resources. Geoscience Australia Canberra (243 pp.).
Mernagh, T.P., Heinrich, C.A., Leckie, J.F., Carville, D.P., Gilbert, D.J., Valenta, R.K., Wyborn, L.A.I., 1994. Chemistry of the low-temperature hydrothermal gold, platinum, and palladium ($\pm$ uranium) mineralization at Coronation Hill, Northern Territory, Australia. Econ. Geol. 89, 1053–1073.
Mernagh, T.P., Wyborn, L.A.I., Jagodzinski, E.A., 1998. Unconformity-related U $\pm$ Au $\pm$ platinum group-element deposits. AGSO J. Aust. Geol. Geophys. 17, 197–205.
Messenger, P.R., Taube, A., Golding, S.D., Hartley, J.S., 1998. Mount Morgan gold–copper deposits. Aust. Inst. Min. Metall. Monogr. 22, 715–721.
Meyer, C., 1999. Ore-forming processes in geologic history. Econ. Geol. 6–41 (75th Anniversary volume).
Meyer, C., 1988. Ore deposits as guides to geologic history of the Earth. Annu. Rev. Earth Planet. Sci. 16, 147–171.
Miller, J., Blewett, R., Tunjic, J., Connors, K., 2010. The role of early formed structures on the development of the world class St Ives Goldfield, Yilgarn, WA. Precambrian Res. 183, 292–315.
Miller, L.D., Goldfarb, R.J., Gehrels, G.E., Snee, L.W., 1988. Genetic links among fluid cycling, vein formation, regional deformation, and plutonism in the Juneau gold belt, southeastern Alaska. Geology 22, 203–206.
Miller, J.M., Phillips, D., Wilson, C.J.L., Dugdale, L.J., 2005. Evolution of a reworked orogenic zone: the boundary between the Delamerian and Lachlan fold belts, southeastern Australia. Aust. J. Earth Sci. 52, 921–940.
Mitchell, R.H., 1986. Kimberlites: Mineralogy, Geochemistry and Petrology. Plenum Press, New York (442 pp.).
Mitchell, R.H., 1995. Kimberlites, Orangeites and Related Rocks. Plenum Press, New York (410 pp.).
Miyashiro, A., 1973. Metamorphism and Metamorphic Belts. Allen and Unwin, London (492 pp.).
Miyashiro, A., Shido, F., Ewing, M., 1971. Metamorphism in the Mid-Atlantic Ridge near 24° 30°N. Philos. Trans. R. Soc. Lond. A Math. Phys. Sci. 268, 589–603.
Mole, D.R., Fiorentini, M.L., Cassidy, K.F., Kirkland, C.L., Thebaud, N., McCuaig, T.C., Doublier, M.P., Duuring, M.L., Romano, S.S., Maas, E.A., Belousova, E.A., Barnes, S.J., Miller, J., 2013. Crustal evolution, intra-cratonic architecture and the metallogeny of an Archaean craton. Geol. Soc. Lond., Spec. Publ. 393. http://dx.doi.org/10.1144/SP393.8.
Mole, D.R., Fiorentini, M.L., Thebaud, N., Cassidy, K.F., McCuaig, T.C., Kirkland, C.L., Romano, S.S., Doublier, M.P., Belousova, E.A., Barnes, S.J., Miller, J., 2014. Archean komatiite volcanism controlled by the evolution of early continents. Proc. Natl. Acad. Sci. 111, 10083–10088. http://dx.doi.org/10.1073/pnas.1400273111.
Mole, D.R., Fiorentini, M.L., Thebaud, N., McCuaig, T.C., Cassidy, K.F., Kirkland, C.L., Wingate, M.T.D., Romano, S.S., Doublier, M.P., Belousova, E.A., 2012. Spatio-temporal constraints on lithospheric development in the southwest–central Yilgarn Craton, Western Australia. Aust. J. Earth Sci. 59, 625–656.
Moores, E.M., 2002. Pre-1 Ga (pre-Rodinian) ophiolites: their tectonic and environmental implications. Geol. Soc. Am. Bull. 114, 80–95.
Moresi, L., Betts, P.G., Miller, M.S., Cayley, R.A., 2014. Dynamics of continental accretion. Nature 508, 245–248.
Moreto, C.P.N., Monteiro, L.V.S., Xavier, R.P., Creaser, R., Dufrane, A., Melo, G.H.C., Delinardo Silva, M.A., 2013. U–Pb and Re–Os geochronological constraints on the timing of iron oxide–Cu–Au systems in the Southern Copper Belt, Carajás Province. Anais do 13° Simpósio de Geologia da Amazônia 22 a 26 de setembro de 2013, pp. 1032–1035.
Morganti, J., 1981. Sedimentary-type stratiform ore deposits: some models and a new classification. Geosci. Can. 8, 65–75.
Morris, R.C., Ramanaidou, E.R., 2007. Genesis of the channel iron deposits (CID) of the Pilbara region, Western Australia. Aust. J. Earth Sci. 54, 733–756.
Morrison, G.W., 1980. Characteristics and tectonic setting of the shoshonite rock association. Lithos 13, 97–108.
Mortensen, J.K., Gemmell, J.B., McNeill, A.W., Friedman, R.M., 2015. High-precision U–Pb zircon chronostratigraphy of the Mount Read Volcanic Belt in western Tasmania, Australia: implications for VHMS deposit formation. Econ. Geol. 110, 445–468.
Moyen, J.-F., van Hunen, J., 2012. Short-term episodicity of Archean plate tectonics. Geology 40, 451–454.
Müller, D., Groves, D.I., 1993. Direct and indirect associations between potassic igneous rocks, shoshonites and gold–copper deposits. Ore Geol. Rev. 8, 383–406.
Müller, S.G., Krapež, B., Barley, M.E., Fletcher, I.R., 2005. Giant iron-ore deposits of the Hamersley province related to the breakup of Paleoproterozoic Australia: new insights from in situ SHRIMP dating of baddeleyite from mafic intrusions. Geology 33, 577–580.
Müller, R.D., Sdrolias, M., Gaina, C., Roest, W.R., 2008. Age, spreading rates and spreading symmetry of the world's ocean crust. Geochem. Geophys. Geosyst. 9, Q04006. http://dx.doi.org/10.1029/2007GC001743.
Murphy, B., 2007. Structural architecture, potential field gradients and exploration potential in the Cobar region, NSW. In: van der Wielen, S., Korsch, R. (Eds.), 3D Architecture and Predictive Mineral System Analysis of the Central Lachlan Subprovince and Cobar Basin, New South Wales. Final Report for pmd*CRC Project, pp. 53–75.
Mustard, R., 2001. Granite-hosted gold mineralization at Timbarra, northern New South Wales, Australia. Mineral. Deposita 36, 542–562.
Naldrett, A.J., 1981. Nickel sulfide deposits: classification, composition, and genesis. Econ. Geol. 628–685 (75th Anniversary volume).
Naldrett, A.J., 1989. Magmatic sulfide deposits. Oxf. Monogr. Geol. Geophys. 14 (186 pp.).
Naldrett, A.J., 2004. Magmatic Sulfide Deposits: Geology, Geochemistry and Exploration. Springer, Heidelberg (727 pp.).
Nelson, D.R., Chivas, A.R., Chappell, B.W., McCulloch, M.T., 1988. Geochemical and isotopic systematics in carbonatites and implications for the evolution of ocean-island sources. Geochim. Cosmochim. Acta 52, 1–17.
Neumann, N.L., Southgate, P.N., Gibson, G.M., McIntyre, A., 2006. New SHRIMP geochronology for the Western Fold Belt of the Mt Isa Inlier: developing a 1800–1650 Ma event framework. Aust. J. Earth Sci. 53, 1023–1039.
Neumayr, P., Hagemann, S.G., 2002. Hydrothermal fluid evolution within the Cadillac Tectonic Zone, Abitibi Greenstone belt, Canada: relationship to auriferous fluids in adjacent second- and third-order shear zones. Econ. Geol. 97, 1203–1225.
Newman, R., White, N., 1999. The dynamics of extensional sedimentary basins: constraints from subsidence inversion. Philos. Trans. R. Soc. London, Ser. A Math. Phys. Eng. Sci. 357, 805–834.
Nicholson, R., Pollard, D., 1985. Dilation and linkage of echelon cracks. J. Struct. Geol. 7, 583–590.
Oliver, N.H.S., Cleverley, J.S., Mark, G., Pollard, P.J., Fu, B., Marshall, L.J., Rubenach, M.J., Williams, P.J., Baker, T., 2004. The role of sodic alteration in the genesis of iron oxide–copper–gold deposits, eastern Mt Isa Block, Australia. Econ. Geol. 99, 1145–1176.
O'Neill, H.S.C., Dingwell, D.B., Borisov, A., Spettel, B., Palme, H., 1995. Experimental petrochemistry of some highly siderophile elements at high temperatures, and some implications for core formation and the mantle's early history. Chem. Geol. 120, 255–273.
O'Neill, C.J., Moresi, L., Jaques, A.L., 2005. Geodynamic controls on diamond deposits: implications for Australian occurrences. Tectonophysics 404, 217–236.
Oreskes, N., Einaudi, M.T., 1992. Origin of hydrothermal fluids at Olympic Dam: preliminary results from fluid inclusions and stable isotopes. Econ. Geol. 87, 64–90.
Orth, K., Meffre, S., Davidson, G., 2014. Age and paragenesis of mineralisation at Coronation Hill uranium deposit, Northern Territory, Australia. Mineral. Deposita 49, 595–623.
Oszczepalski, S., 1999. Origin of the Kupferschiefer polymetallic mineralization in Poland. Mineral. Deposita 34, 599–613.
Page, R.W., Conor, C.H.H., Stevens, B.P.J., Gibson, G.M., Preiss, W., Southgate, P.N., 2005a. Correlation of Olary and Broken Hill domains, Curnamona Province: possible relationship to Mount Isa and other north Australian Pb–Zn–Ag-bearing successions. Econ. Geol. 100, 663–676.
Page, R.W., Stevens, B.P.J., Gibson, G.M., 2005b. Geochronology of the sequence hosting the Broken Hill Pb–Zn–Ag orebody, Australia. Econ. Geol. 100, 633–661.
Palme, H., O'Neill, H.S.C., 2004. Cosmochemical estimates of mantle composition. Treatise Geochem. 2, 1–38.
Parkinson, I.J., Arculus, R.J., 1999. The redox state of subduction zones: insights from arc-peridotites. Chem. Geol. 160, 409–423.
Passchier, C.W., Trouw, R.A.J., 2005. Microtectonics. 2nd edition. Springer-Verlag, Berlin Heidelberg (366 pp.).
Pearce, J.A., 1982. Trace element characteristics of lavas from destructive plate boundaries. In: Thorpe, R.S. (Ed.), Andesites. John Wiley and Sons, New York, pp. 525–548.
Pearce, J.A., 1986. Geochemical fingerprinting of oceanic basalts with applications to ophiolite classification and the search for Archean oceanic crust. Lithos 100, 14–48.
Pearce, J.A., Cann, J.R., 1973. Tectonic setting of basic volcanic rocks determined by trace element analysis. Earth Planet. Sci. Lett. 19, 290–300.
Pearce, J.A., Norry, M.J., 1979. Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks. Contrib. Mineral. Petrol. 69, 33–47.
Pearce, J.A., Harris, N.B.W., Tindle, A.G., 1984. Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. J. Petrol. 25, 956–983.
Pease, V., Percival, J., Smithies, H., Stevens, G., Van Kranendonk, M., 2008. When did plate tectonics begin? Evidence from the orogenic record. Geol. Soc. Am. Spec. Pap. 440, 199–228.
Peltonen, P., Mänttäri, I., Huhma, H., Kontinen, A., 2003. Archean zircons from the mantle: the Jormua ophiolite revisited. Geology 31, 645–648.

Perkins, C., Heinrich, C.A., Wyborn, L.A.I., 1999. $^{40}Ar/^{39}Ar$ geochronology of copper mineralization and regional alteration, Mount Isa, Australia. Econ. Geol. 94, 23–36.
Perkins, C., Walshe, J.L., Morrison, G., 1995. Metallogenic episodes of the Tasman fold belt system, eastern Australia. Econ. Geol. 90, 1443–1466.
Phillips, D., Fu, B., Wilson, C.J.L., Kendrick, M.A., Fairmaid, A.M., Miller, J.M., 2012. Timing of gold mineralisation in the western Lachlan Orogen, SE Australia: a critical overview. Aust. J. Earth Sci. 59, 495–525.
Phillips, G.N., Groves, D.I., Kerrich, R., 1996. Factors in the formation of the giant Kalgoorlie gold deposit. Ore Geol. Rev. 10, 295–317.
Phillips, G.N., Hughes, M.J., Arne, D.C., Bierlein, F.P., Carey, S.P., Jackson, T., Willman, C.E., 2003. Gold historic wealth, future potential. Geol. Soc. Aust. Spec. Publ. 23, 377–432.
Piercey, S.J., 2011. The setting, style, and role of magmatism in the formation of volcanogenic massive sulfide deposits. Mineral. Deposita 46, 449–471.
Piercey, S.J., Paradis, S., Murphy, D.C., Mortensen, J.K., 2001. Geochemistry and paleotectonic setting of felsic volcanic rocks in the Finlayson Lake volcanic-hosted massive sulfide (VHMS) district, Yukon, Canada. Econ. Geol. 96, 1877–1905.
Piestrzyński, A., 1990. Uranium and thorium in the Kupferschiefer formation, Lower Zechstein, Poland. Mineral. Deposita 25, 146–151.
Pirajno, F., Hoatson, D.M., 2012. A review of Australia's Large Igneous Provinces and associated mineral systems: implications for mantle dynamics through geological time. Ore Geol. Rev. 48, 2–54.
Plumb, K.A., Derrick, G.M., 1975. Geology of the Proterozoic rocks of the Kimberley to Mount Isa region. Aust. Inst. Min. Metall. Monogr. 5, 217–252.
Polito, P.A., Kyser, T.K., Alexandre, P., Hiatt, E.E., Stanley, C.R., 2011. Advances in understanding the Kombolgie Subgroup and unconformity-related uranium deposits in the Alligator Rivers Uranium Field and how to explore for them using lithogeochemical principles. Aust. J. Earth Sci. 58, 453–474.
Polito, P.A., Kyser, T.K., Marlatt, J., Alexandre, P., Bajwah, Z., Drever, G., 2004. Significance of alteration assemblages for the origin and evolution of the Proterozoic Nabarlek unconformity-related uranium deposit, Northern Territory, Australia. Econ. Geol. 99, 113–139.
Polito, P.A., Kyser, T.K., Thomas, D., Marlatt, J., Drever, G., 2005. Reevaluation of the petrogenesis of the Proterozoic Jabiluka unconformity related uranium deposit, Northern Territory, Australia. Mineral. Deposita 40, 257–288.
Powell, C.M., Oliver, N.H., Li, Z.X., Martin, D.M., Ronaszeki, J., 1999. Synorogenic hydrothermal origin for giant Hamersley iron oxide ore bodies. Geology 27, 175–178.
Preiss, W.V., Drexel, J.F., Reid, A.J., 2009. Definition and age of the Kooringa Member of the Skillogalee Dolomite: host of Neoproterozoic (c. 790 Ma) porphyry-related copper mineralisation at Burra. MESA J. 55, 19–33.
Price, R.C., Nicholls, I.A., Day, A., 2014. Lithospheric influences on magma compositions of late Mesozoic and Cenozoic intraplate basalts (the Older Volcanics) of Victoria, south-eastern Australia. Lithos 206-207, 179–200.
Prosser, S., 1993. Rift-related linked depositional systems and their expression. Geol. Soc. Lond. Spec. Publ. 21, 283–308.
Qui, Y., Groves, D.I., McNaughton, N.J., 1997. Deep-seated granitoids: implications for Late Archean subduction–collision–lithospheric delamination and gold mineralization in the Yilgarn Craton. Aust. Geol. Surv. Organ. Rec. 1997 (41), 65–69.
Ramsden, A.R., French, D.H., Chalmers, D.I., 1994. Volcanic-hosted rare-metals at Brockman. Western Australia. Mineral. Deposita 28, 1–12.
Rawlings, D.J., Korsch, R.J., Goleby, B.R., Gibson, G.M., Johnstone, D.W., Barlow, M., 2004. The 2002 southern Mcarthur Basin seismic reflection survey. Geosci. Aust. Rec. 2004 (17), 78.
Richards, J.P., 1995. Alkalic-type epithermal gold deposits—a review. Mineral. Assoc. Can. Short Course Ser. 23, 367–400.
Richards, J.P., 2003. Tectono-magmatic precursors for porphyry Cu–(Mo–Au) deposit formation. Econ. Geol. 96, 1515–1533.
Richards, J.P., 2009. Postsubduction porphyry Cu–Au and epithermal Au deposits: products of remelting of subduction-modified lithosphere. Geology 37, 247–250.
Richards, J.P., 2015. The oxidation state, and sulfur and Cu contents of arc magmas: implications for metallogeny. Lithos 233, 27–45.
Richards, J.P., Mumin, A.H., 2013. Magmatic-hydrothermal processes within an evolving Earth: iron oxide–copper–gold and porphyry Cu ± Mo ± Au deposits. Geology 41, 767–770.
Richardson, S.M., Moy, A.D., 1998. Gunpowder copper deposits. Australas. Inst. Min. Metall. Monogr. 22, pp. 743–752
Robert, F., 1989. Internal structure of the Cadillac tectonic zone southeast of Val d'Or, Abitibi greenstone belt, Quebec. Can. J. Earth Sci. 26, 2661–2675.
Roberts, G.G., White, N., 2010. Estimating uplift rate histories from river profiles using African examples. J. Geophys. Res. Solid Earth 115 (B2).
Rose, A.W., 1976. The effect of cupreous chlorite complexes in the origin of red-bed copper and related deposits. Econ. Geol. 71, 1036–1048.
Rose, A.W., Hawkes, H.E., Webb, J.S., 1979. *Geochemistry in Mineral Exploration, second ed.* Academic Press, London (657 pp.).
Rosière, C.A., Spier, C.A., Rios, F.J., Suckau, V.E., 2008. The itabirites of the Quadrilátero Ferrífero and related high-grade iron ore deposits: an overview. Rev. Econ. Geol. 15, 223–254.
Rozendaal, A., 1986. The Gamsberg zinc deposit, Namaqualand district. In: Anhaeusser, C.R., Maske, S. (Eds.), Ore Deposits of Southern Africa. Geological Society of South Africa, Johannesburg, pp. 1477–1488.
Rudnick, R.L., Gao, S., 2003. Composition of the continental crust. Treatise Geochem. 3, 1–64.
Ryan, P.J., Lawrence, A.L., Lipson, R.D., Moore, J.M., Paterson, A., Stedman, D.P., Van Zyl, D., 1986. The Aggeneys base metal sulphide deposits, Namaqualand, South Africa. In: Anhaeusser, C.R., Maske, S. (Eds.), Ore Deposits of Southern Africa. Geological Society of South Africa, Johannesburg, pp. 1447–1474.
Sandiford, M., 2007. The tilting continent: A new constraint on the dynamic topographic field from Australia. Earth Planet. Sci. Let. 261, 152–163.
Sato, T., 1972. Behaviors of ore-forming solutions in seawater. Min. Geol. 22, 31–42.
Sawkins, F.J., 1984. Metal Deposits in Relationship to Plate Tectonics. Springer-Verlag, Berlin (325 pp.).
Schellart, W.O., 2008. Subduction zone trench migration: slab driven or overriding-plate-drive? Phys. Earth Planet. Inter. 170, 73–88.
Schodde, R., 2011. Recent trends in Australian exploration. Presentation to the AMEC 2011 National Mining Congress, Perth, June 2011. http://www.minexconsulting.com/publications/Australian%20Discovery%20Performance%20-%20MinEx%20UPDATED%20-%20June%202011.pdf.
Schofield, N.J., Brown, D.J., Magee, C., Stevenson, C.T., 2013. Sill morphology and comparison of brittle and non-brittle emplacement mechanisms. J. Geol. Soc. Lond. 169, 127–141.
Scrimgeour, I.R., Kinny, P.D., Close, D.F., Edgoose, C.J., 2005. High-T granulites and polymetamorphism in the southern Arunta region, central Australia: evidence for a 1.64 Ga accretional event. Precambrian Res. 142, 1–27.
Seat, Z., Beresford, S.W., Grguric, B.A., Waugh, R.S., Hronsky, J.M., Gee, M.M., Groves, D.I., Mathison, C.I., 2007. Architecture and emplacement of the Nebo–Babel gabbronorite-hosted magmatic Ni–Cu–PGE sulphide deposit, West Musgrave, Western Australia. Mineral. Deposita 42, 551–581.
Seedorf, E., Dilles, J.H., Proffett Jr., J.M., Einaudi, M.T., Zurcher, L., Stavast, W.J.A., Johnson, D.A., Barton, M.D., 2005. Porphyry deposits: characteristics and origin of hypogene features. Econ. Geol. 251–298 (100th Anniversary volume).
Selley, D., Broughton, D., Scott, R., Hitzman, M., Bull, S., Large, R., McGoldrick, P., Croaker, M., Pollington, N., Barra, F., 2005. A new look at the geology of the Zambian copperbelt. Econ. Geol. 965–1000 (100th Anniversary volume).
Seymour, D.B., Green, G.R., Calver, R.R., 2006. The geology and mineral deposits of Tasmania: a summary. Miner. Resour. Tasmania Geol. Surv. Bull. 72.
Sharpe, R., Gemmell, J.B., 2002. The Archean Cu–Zn magnetite-rich Gossan Hill volcanic-hosted massive sulfide deposit, Western Australia: genesis of a multistage hydrothermal system. Econ. Geol. 97, 517–539.
Shirey, S.B., Richardson, S.H., 2011. Start of the Wilson Cycle at 3 Ga shown by diamonds from subcontinental mantle. Science 333, 434–436.
Sibson, R.H., Moore, J.M.M., Rankin, A.H., 1975. Seismic pumping—a hydrothermal fluid transport mechanism. J. Geol. Soc. Lond. 131, 653–659.
Sillitoe, R.H., 1972. A plate tectonic model for the origin of porphyry copper deposits. Econ. Geol. 67, 184–197.
Sillitoe, R.H., 1993. Gold-rich porphyry copper deposits: geological model and exploration implications. Geol. Assoc. Can. Spec. Pap. 40, 465–478.
Sillitoe, R.H., 1998. Major regional factors favoring large size, high hypogene grade, elevated gold content and supergene oxidation and enrichment of porphyry copper deposits. In: Porter, T.M. (Ed.), Porphyry and Hydrothermal Copper and Gold Deposits: A Global Perspective, Perth, 1998, Conference Proceedings: Glenside. South Australia, Australian Mineral Foundation, pp. 21–34.
Sillitoe, R.H., Perelló, J., 2005. Andean Copper Province: Tectonomagmatic Setting, Deposit Types, Metallogeny, Exploration, and Discovery. Econ. Geol. 845–890 (100th Anniversary volume).
Sillitoe, R.H., Hannington, M.D., Thompson, J.F., 1996. High sulfidation deposits in the volcanogenic massive sulfide environment. Econ. Geol. 91, 204–212.
Simmons, S.F., White, N.C., John, D.A., 2005. Geological characteristics of epithermal precious and base metal deposits. Econ. Geol. 485–522 (100th Anniversary volume).
Sinclair, H.D., Coakley, B.J., Allen, P.A., Watts, A.B., 1991. Simulation of foreland basin stratigraphy using a diffusion model of mountain belt uplift and erosion: an example from the central Alps, Switzerland. Tectonics 10, 599–620.
Sizova, E., Gerya, T., Brown, M., 2014. Contrasting styles of Phanerozoic and Precambrian continental collision. Gondwana Res. 25, 522–545.
Sizova, E., Gerya, T., Brown, M., Perchuk, L.L., 2010. Subduction styles in the Precambrian: insight from numerical experiments. Lithos 116, 209–229.
Skirrow, R.G., 2010. 'Hematite-group' IOCG ± U ore systems: tectonic settings, hydrothermal characteristics, and Cu–Au and U mineralizing processes. Geol. Assoc. Can. Short Course Notes 20, 39–57.
Skirrow, R.G., Bastrakov, E.N., Barovich, K., Fraser, G.L., Creaser, R.A., Fanning, C.M., Raymond, O.L., Davidson, G.J., 2007. Timing of iron oxide Cu–Au–(U) hydrothermal activity and Nd isotopic constraints on metal sources in the Gawler Craton, South Australia. Econ. Geol. 102, 1441–1470.
Skirrow, R.G., Huston, D.L., Mernagh, T.P., Thorne, J.P., Dulfer, H., Senior, A.B., 2013. Critical Commodities for a High-tech World: Australia's Potential to Supply Global Demand. Geoscience Australia, Canberra.
Skirrow, R.G., Mercadier, J., Armstrong, R., Kuske, T., Deloule, E., 2016. The Ranger uranium mineral system, and temporal-spatial controls on the formation of unconformity-related uranium deposits in the Pine Creek Orogen. Ore. Geol. Rev. 76, 463–503 (in this issue).
Sleep, N.H., 2000. Evolution of the mode of convection within terrestrial planets. J. Geophys. Res. Planets 105, 17563–17578.
Sloss, L.L., 1963. Sequences in the cratonic interior of North America. Bull. Geol. Soc. Am. 74, 93–114.
Smith, C.B., 1983. Pb, Sr and Nd isotopic evidence for sources of southern African Cretaceous kimberlites. Nature 304, 51–54.
Smith, S.G., 1996. Geology and geochemistry of the Warrabarty carbonate-hosted Zn–Pb prospect, Paterson Orogen, Western Australia. Unpublished Ph.D. thesis, University of Tasmania.

Smithies, R.H., Champion, D.C., 2000. The Archaean high-Mg diorite suite: links to tonalite–trondhjemite–granodiorite magmatism and implications for early Archaean crustal growth. J. Petrol. 41, 1653–1671.
Smithies, R.H., Van Kranendonk, M.J., Champion, D.C., 2005. It started with a plume — early Archaean basaltic proto-continental crust. Earth Planet. Sci. Lett. 238, 284–297.
Sobolev, A.V., Hofmann, A.W., Kuzmin, D.V., Yaxley, G.M., Arndt, N.T., Chung, S.-L., Danyushevsky, L.V., Elliott, T., Frey, F.A., Garcia, M.O., Gurenko, A.A., Kamenetsky, V.S., Kerr, A.C., Krivolutskaya, N.A., Matvienkov, V.V., Nikogosian, I.K., Rocholl, A., Sigurdsson, I.A., Sushchevskaya, N.M., Teklay, M., 2007. The amount of recycled crust in sources of mantle-derived melts. Science 316, 412–417.
Solomon, M., Groves, D.I., 2000. The Geology and Origin of Australia's Mineral Deposits. Centre for Ore Deposit Research. University of Tasmania, Hobart (1002 pp.).
Southgate, P.N., Scott, D.L., Sami, T.T., Domalga, J., Jackson, M.J., James, N.P., Kyser, T.K., 2000. Basin shape and sediment architecture in the Gun Supersequence: a strike-slip model for Pb–Zn–Ag ore genesis at Mt Isa. Aust. J. Earth Sci. 47, 509–531.
Squire, R.J., Miller, J.M., 2003. Synchronous compression and extension in East Gondwana; tectonic controls on world-class gold deposits at 440 Ma. Geology 31, 1073–1076.
Stanton, R.L., 1960. General features of the conformable pyritic orebodies. Part I — field association. Can. Inst. Min. Metall. Bull. 53, 24–29.
Steckler, M.S., Berthelot, F., Lyberis, N., Le Pichon, X., 1988. Subsidence in the Gulf of Suez: implications for rifting and plate kinematics. Tectonophysics 153, 249–270.
Stegman, C.L., 2001. Cobar deposits: still defying classification. SEG Newsl. 44, 15–25.
Stein, H.J., Markey, R.J., Morgan, J.W., Selby, D., Creaser, R.A., McCuaig, T.C., Behn, M., 2001. Re–Os dating of Boddington molybdenite, SW Yilgarn: two Au mineralization events. Geosci. Aust. Rec. 2001 (37), 469–471.
Stern, R.J., 2002. Subduction zones. Rev. Geophys. 40, 1012.
Stern, R.J., 2007. When and how did plate tectonics begin? Theoretical and empirical considerations. Chin. Sci. Bull. 52, 578–591.
Stern, R.J., 2008. Modern-style plate tectonics began in Neoproterozoic time: an alternative interpretation of Earth's tectonic history. Geol. Soc. Am. Spec. Pap. 440, 265–280.
Stern, R.J., 2010. The anatomy and ontogeny of modern intra-oceanic arc systems. Geol. Soc. Lond. Spec. Publ. 338, 7–34.
Storey, B.C., 1995. The role of mantle plumes in continental breakup: case histories from Gondwanaland. Nature 377, 301–308.
Stuwe, K., Sandiford, M., 1995. A description of metamorphic PTt paths with implications for low-P and high-T metamorphism. Phys. Earth Planet. Inter. 88, 211–221.
Sun, S.-S., Carr, G.R., Page, R.W., 1994. Lead isotope-based stratigraphic correlations and ages of Proterozoic sediment-hosted Pb–Zn deposits in the Mount Isa Inlier. AGSO Res. Newsl. 20, 1–2.
Sun, S.-S., Carr, G.R., Page, R.W., 1996. A continued effort to improve lead-istope model ages. AGSO Res. Newsl. 24, 19–20.
Sweet, I.P., Hutton, L.J., Leggo, N., Pietsch, B.A., Ahmad, M., Rawlings, D.J., Haines, P.W., 1993. The AusIMM (Northwest Queensland and Darwin branches) Carpentaria zinc belt geological excursion — Gulf region, Queensland and Northern Territory. The Gulf Classic — Excursion Guide (73 pp.).
Symons, D.T.A., 2007. Paleomagnetism of the HYC Zn–Pb SEDEX deposit, Australia: evidence of an epigenetic origin. Econ. Geol. 102, 1295–1310.
Tappe, S., Foley, S.F., Jenner, G.A., Heaman, L.M., Kjarsgaard, B.A., Romer, R.L., Stracke, A., Joyce, N., Hoeffs, J., 2006. Genesis of ultramafic lamprophyres and carbonatites at Aillik Bay, Labrador: a consequence of incipient lithospheric thinning beneath the North Atlantic Craton. J. Petrol. 47, 1261–1315.
Tatsumi, Y., Eggins, S., 1995. Subduction zone magmatism. Frontiers in Earth Sciences. Blackwell Science, Boston (211 pp.).
Taylor, B., Martinez, F., 2003. Back-arc basin basalt systematics. Earth Planet. Sci. Lett. 210, 481–497.
van Thienen, P., Vlaar, N.J., van den Berg, A.P., 2004. Plate tectonics on the terrestrial planets. Phys. Earth Planet. Inter. 142, 61–74.
Thompson, J.F.H., Sillitoe, R.H., Baker, T., Lang, J.R., Mortensen, J.K., 1999. Intrusion-related gold deposits associated with tungsten-tin provinces. Mineral. Deposita 34, 323–334.
Thorne, W., Hagemann, S.G., Webb, A., Clout, J., 2008. BIF-related deposits of the Hamersley province, Western Australia. Rev. Econ. Geol. 15, 197–222.
Thorne, W.S., Hagemann, S.G., Sepe, D., Dalstra, H.J., Banks, D.A., 2014. Structural control, hydrothermal alteration zonation, and fluid chemistry of the concealed, high-grade 4EE iron orebody at the Paraburdoo 4E deposit, Hamersley Province, Western Australia. Econ. Geol. 109, 1529–1562.
Titley, S.R., Beane, R.E., 1981. Prorphyry copper deposits. Econ. Geol. 214–269 (75th Anniversary volume).
Tomkins, A.G., Grundy, C., 2009. Upper temperature limits of orogenic gold deposit formation: constraints from the granulite-hosted Griffin's Find deposit, Yilgarn Craton. Econ. Geol. 104, 669–685.
Tomkins, A.G., Mavrogenes, J.A., 2002. Mobilization of gold as a polymetallic melt during pelite anatexis at the Challenger deposit, South Australia: a metamorphosed Archean gold deposit. Econ. Geol. 97, 1249–1271.
Tomkins, A.G., Pattison, D.R.M., Zaleski, E., 2004. The Hemlo gold deposit, Ontario: an example of melting and mobilization of a precious metal-sulfosalt assemblage during amphibolite facies metamorphism and deformation. Econ. Geol. 99, 1063–1084.
Tonn, H., Schmidt, F.-P., Porada, H., Horn, E.-E., 1987. Untersuchungen von Flűssigkeitseinschlűssen im Zechstein als Beitrag zur Genese des Kupferschiefers. Fortschr. Mineral. 65 (Supplement 1), 184.
Trendall, A.F., Nelson, D.R., De Laeter, J.R., Hassler, S.W., 1998. Precise zircon U–Pb ages from the Marra Mamba Iron Formation and Wittenoom Formation, Hamersley Group, Western Australia. Aust. J. Earth Sci. 45, 137–142.
Twelvetrees, W.H., Ward, L.K., 1910. The orebodies of the Zeehan field. Geol. Surv. Tasmania Bull. 8.
Unternehr, P., Péron-Pinvidic, G., Manatschal, G., Sutra, E., 2010. Hyper-extended crust in the South Atlantic: in search of a model. Pet. Geosci. 16, 207–215.
van Hunen, J., Moyen, J.-F., 2012. Archean subduction: fact or fiction? Annu. Rev. Earth Planet. Sci. 40, 195–219.
van Hunen, J., van der Berg, A.P., 2008. Plate tectonics on the early Earth: limitations imposed by strength and buoyancy of subducted lithosphere. Lithos 103, 217–235.
Van Kranendonk, M.J., Philippot, P., Lepot, K., Bodorkos, S., Pirajno, F., 2008. Geological setting of Earth's oldest fossils in the c. 3.5 Ga Dresser Formation, Pilbara Craton, Western Australia. Precambrian Res. 167, 93–124.
Van Kranendonk, M.J., Smithies, R.H., Griffin, W.L., Huston, D.L., Hickman, A.H., Champion, D.C., Anhaeusser, C.R., Pirajno, F., 2014. Making it thick: a volcanic plateau origin of Palaeoarchean continental lithosphere of the Pilbara and Kaapvaal cratons. Geol. Soc. Lond. Spec. Publ. 389, 83–111.
Van Wagoner, J.C., Posamentier, H., Mitchum, R., Vail, P., Sarg, J., Loutit, T., Hardenbol, J., 1988. An overview of the fundamentals of sequence stratigraphy and key definitions. SEPM Spec. Publ. 42, 39–45.
VandenBerg, A.H.M., Willman, C.E., Maher, S., Simons, B.A., Cayley, R.A., Taylor, D.H., Morand, V.J., Moore, D.H., Radojkovic, A., 2000. The Tasman Fold Belt System in Victoria. Geological Survey of Victoria Special Publication (462 pp.).
Vine, F.J., Matthews, D.H., 1963. Magnetic anomalies over oceanic ridges. Nature 199, 947–949.
Von Pechmann, E., 1986. Mineralogy, age dating and genesis of the Kylie and Spring Creek uranium prospects, Waterhouse Complex, Northern Territory, Australia. In: Craig, J.R., Hagni, R.D., Kiesl, W., Lange, I.M., Petrovskaya, N.V., Shadlun, T.N., Udubasa, G., Augustithis, S.S. (Eds.), Mineral Paragenesis. Theophrastus Publications, SA, Athens, pp. 303–343.
Waller, G.A., 1904. Report of the Zeehan Silver–Lead Mining Field. Report of Secretary of Mines, Tasmania.
Walter, M.R., Veevers, J.J., Calver, C.R., Grey, K., 1995. Neoproterozoic stratigraphy of the Centralian Superbasin, Australia. Precambrian Res. 73, 173–195.
Waring, C.L., Heinrich, C.A., Wall, V.J., 1998. Proterozoic metamorphic copper deposits. AGSO J. Aust. Geol. Geophys. 17, 239–246.
Warren, J.K., 2010. Evaporites through time: tectonic, climatic and eustatic controls in marine and nonmarine deposits. Earth Sci. Rev. 98, 217–268.
Watts, A., 2001. Isostasy and Flexure of the Lithopshere. Cambridge University Press, Cambridge.
Wedepohl, K.H., Hartmann, G., 1994. The composition of the primitive upper earth's mantle. In: Meyer, H.O.A., Leonardos, O.H. (Eds.), Kimberlites, Related Rocks and Mantle Xenoliths. Companhia de Pesquisa de Recursos Minerais, Rio de Janeiro, pp. 486–495.
Wegener, A., 1912. Die entstehung der kontinente. Geol. Rundsch. 3, 276–292.
Wernicke, B., 1985. Uniform-sense normal simple shear of the continental lithosphere. Can. J. Earth Sci. 22, 108–125.
White, N., McKenzie, D., 1988. Formation of the "steer's head" geometry of sedimentary basins by differential stretching of the crust and mantle. Geology 16, 250–253.
White, R., McKenzie, D., 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. J. Geophys. Res. 94 (B6), 7685–7729.
White, N., Thompson, M., Barwise, T., 2003. Understanding the thermal evolution of deep-water continental margins. Nature 426, 334–343.
Whitmarsh, R.B., Manatschal, G., Minshull, T.A., 2001. Evolution of magma-poor continental margins from rifting to seafloor spreading. Nature 413, 150–154.
Williams, N., 1978. Studies of base metal sulfide deposits at McArthur River, Northern Territory, Australia: I. The Cooley and Ridge deposits. Econ. Geol. 73, 1005–1035.
Williams, P.J., Barton, M.D., Johnson, D.A., Fontboté, L., de Haller, A., Mark, G., Oliver, N.H.S., Marschik, R., 2005. Iron oxide copper–gold deposits; geology, space–time distribution, and possible modes of origin. Econ. Geol. 371–405 (100th Anniversary volume).
Willis, I.L., Brown, R.E., Stroud, W.J., Stevens, B.J.P., 1983. The early Proterozoic Willyama Supergroup: stratigraphic subdivision and interpretation of high- to low-grade metamorphic rocks in the Broken Hill block, New South Wales. J. Geol. Soc. Aust. 30, 195–224.
Willman, C.E., Korsch, R.J., Moore, D.H., Cayley, R.A., Lisitsin, V.A., Rawling, T.J., Morand, V.J., O'Shea, P.J., 2010. Crustal-scale fluid pathways and source rocks in the Victorian Gold Province, Australia: insights from deep seismic reflection profiles. Econ. Geol. 105, 895–915. http://dx.doi.org/10.2113/econgeo.105.5.895.
Willocks, A.J., Haydon, S.J., Asten, M.W., Moore, D.H., 1999. Geophysical signatures of base metal deposits in Victoria. Geological Survey of Victoria Report 119 and Australian Society of Exploration Geophysicists Special Publication 11.
Wilson, M., 1989. Petrogenesis — A Global Tectonic Approach. Springer, Amsterdam.
Winchester, J.A., Floyd, P.A., 1977. Geochemical discrimination of different magma series and their differentiation products using immobile elements. Chem. Geol. 20, 325–343.
Wingate, M.T.D., Campbell, I.H., Compston, W., Gibson, G.M., 1998. Ion microprobe U–Pb ages for Neoproterozoic basaltic magmatism in south-central Australia and implications for the breakup of Rodinia. Precambrian Res. 87, 135–159.
Winkler, H.G.F., 1979. Petrogenesis of Metamorphic Rocks. 5th Edition. Springer Verlag, New York.
Winterbourne, J., White, N., Crosby, A., 2014. Accurate measurements of residual topography from the oceanic realm. Tectonics 33, 982–1015.
Wood, D., 2012. Discovery of the Cadia deposits, NSW, Australia (part 1). SEG Newsl. 88 (1), 13–18.
Wood, D.A., Joron, J.-L., Treuil, M.A., 1979. Re-appraisal of the use of trace elements to classify and discriminate between magma series erupted in different tectonic settings. Earth Planet. Sci. Lett. 45, 326–336.
Woodall, R., 1990. Gold in Australia. Aust. Inst. Min. Metall. Monogr. 14, 45–68.
Woodhead, J., Eggins, S., Gamble, J., 1993. High field strength and transition element systematics in island arc and back-arc basin basalts: evidence for multi-phase melt extraction and a depleted mantle wedge. Earth Planet. Sci. Lett. 114, 491–504.

Woolley, A.R., 1989. The spatial and temporal distribution of carbonatites. In: Bell, K. (Ed.), Carbonatites: Genesis and Evolution. Unwin Hyman, London, pp. 15–37.
Woolley, A.R., Bailey, D.K., 2012. The crucial role of lithospheric structure in the generation and release of carbonatites: geological evidence. Mineral. Mag. 76, 259–270.
Woolley, A.R., Kjarsgaard, B.A., 2008. Carbonatite occurrences of the world: map and database. Geol. Surv. Can. 28 (Open File 5796).
Wyborn, L.A.I., Heinrich, C.A., Jaques, A.L., 1994. Australian Proterozoic mineral systems: essential ingredients and mappable criteria. AusIMM Publication Series 5, pp. 109–115.
Wyman, D.A., Kerrich, R., Groves, D.I., 1999. Lode gold deposits and Archean mantle plume–island arc interaction, Abitibi subprovince, Canada. J. Geol. 107, 715–725.
Wyman, D.A., Kerrich, R., Polat, A., 2003. Assembly of Archean cratonic mantle lithosphere and crust: plume–arc interaction in the Abitibi-Wawa subduction–accretion complex. Precambrian Res. 15, 37–62.
Yeats, C.J., Groves, D.I., 1998. The Archaean Mount Gibson gold deposits, Yilgarn Craton, Western Australia: products of combined synvolcanic and syntectonic alteration and mineralisation. Ore Geol. Rev 13, 103–129.
Yeats, C.J., McNaughton, N.J., Groves, D.I., 1996. SHRIMP U–Pb geochronological constraints on Archean volcanic-hosted massive sulfide and lode gold mineralization at Mount Gibson, Yilgarn craton, Western Australia. Econ. Geol. 91, 1354–1371.
Zhang, M., O'Reilly, S., Wang, K.L., Hronsky, J., Griffin, W., 2008. Flood basalts and metallogeny: the lithosphere–mantle connection. Earth-Sci. Rev. 86, 145–174.
Zhao, J.X., McCulloch, M.T., Korsch, R.J., 1994. Characterisation of a plume-related ~800 Ma magmatic event and its implications for basin formation in central-southern Australia. Earth Planet. Sci. Lett. 121, 349–367.
Ziegler, P.A., Dèzes, P., 2005. Evolution of the lithosphere in the area of the Rhine Rift System. Int. J. Earth Sci. 94, 594–614.