
1.1 Theory

1.1.1 Introduction

The pattern of crustal evolution in the Phanerozoic Eon, involving the progressive amalgamation of large continental fragments to form the supercontinent of Pangea during Permian and Triassic times, and its subsequent dispersal to form the present day continental geography, is well constrained. A relatively high degree of confidence marks the reconstruction of continental configuration during the Phanerozoic Eon and geoscientists generally agree on the paleogeography of Gondwana, Laurentia, and Pangea. As one goes back in time, however, even into the late Precambrian, the situation becomes progressively more uncertain and disagreement often accompanies the reconstruction of continental geometry. There are many reasons for this, including the difficulties in acquiring and accurately dating apparent polar wander paths, the high rates of recycling (or destruction) of continental and oceanic crust, and progressive deformation and burial of crust with time.

Despite these difficulties, considerable progress has been made in reconstructing continental paleogeography during the latter stages of the Proterozoic Eon. There are indications that a Neoproterozoic supercontinent, generally referred to as Rodinia, was assembled from about 1000 Ma, and then dispersed again by about 700 Ma. There is also some evidence for the existence of a Mesoproterozoic supercontinent, although at this stage there is still considerable uncertainty as to the nature of its constituents, as well as its shape and position. It is also likely that substantial continents existed in the Palaeoproterozoic and even into the late Archaean, although at this stage the configuration of these continents is speculative.

Windley (1995) has compiled a detailed account of the many ideas regarding the evolution of the crust over the entire span of Earth history, and this work provides an excellent platform upon which to base the following discussion. Rogers (1996) has also attempted to reconstruct a history of the continents and their configuration over the past 3000 million years. He has also chosen to name the larger continental fragments over this time period and charted their evolution in terms of major periods of amalgamation and dispersal. In the sections that follow this scheme is summarized and complemented with more recent data pertaining to Neoproterozoic (Rodinia) and Phanerozoic (Pangea) continental evolution. Although still in its early stages of development and, therefore, speculative, the Rogers scheme nevertheless provides a useful framework upon which to base a discussion of metallogeny through time. It is also a working hypothesis that can be used to test ideas relating to crustal evolution and the link to global metallogeny. In brief, the Rogers model commences with an inferred Archaean continent known as Ur, comprising the ancient Kaapvaal and Pilbara cratons of southern Africa and Western Australia, respectively, as well as parts of India and Antarctica. Ur is believed to have coalesced from about 3000 Ma and to
have existed as a separate block throughout much of Earth history until it contributed to the assembly of a supercontinent at around 1000 Ma. Another major continent, called Arctica, now largely preserved in parts of Canada, Greenland, and Siberia, is believed to have amalgamated toward the end of the Archaean or in the early Proterozoic. At least two other substantial continental fragments are believed to have formed during the Palaeoproterozoic, namely Baltica and Atlantica. Arctica (also commonly referred to as Laurentia in earlier reconstructions; Hoffman, 1988) and Baltica are suggested to have amalgamated at around 1500 Ma to form the first substantially consolidated mass of continental material (but not yet a supercontinent), called Nena. Nena, Ur, and Atlantica are considered to have been the founding blocks for what was arguably the first supercontinent, Rodinia, which existed largely in the Neoproterozoic and comprised an amalgamation of most continental mass at that time. Thereafter, the pattern of continental evolution is far better constrained. Rodinia broke up into essentially three large fragments known as Laurentia (i.e. an enlarged version of the Palaeoproterozoic Laurentia) and east and west Gondwana, although it is likely that other smaller continental entities also existed at this time. The two sections of Gondwana amalgamated into a single entity during the Panfrican and Brasiliano orogenies at around 500 Ma. Laurentia collided with Baltica after the closure of the Iapetus Ocean in the early Palaeozoic, at about 400 Ma, to form a transient entity known as Laurasia. It was the subsequent collision of Laurasia and Gondwana that gave rise to the second global supercontinent, Pangea, which formed at around 300 Ma. Since then continental fragments have dispersed to many parts of the globe, forming the present day geographic configuration. Some areas are currently undergoing reamalgamation, with collision of India with Asia, and the partial amalgamation of Africa and Europe, having been particularly active over the past 20 million years. This pattern of continental evolution is used below as the framework around which to discuss global metallogenic trends. The discussion progresses from oldest to youngest, even though much more is known about processes in younger periods of time.

1.1.2 The Archaean Eon

1.1.2.1 The Hadean (>4000 Ma) and Eoarchean (>3600 Ma) stages

The Hadean Era refers to that period of Earth history for which there is very little evidence in the rock record and which is nominally pre-Archaean. It was a time of global differentiation and accretion, as well as intense meteorite bombardment. The Hadean was previously regarded as existing prior to 3800 Ma (Harland, 1989) although the growing evidence (from U–Pb zircon dating) for crustal remnants at close to 4000 Ma suggests that the latter date is perhaps a more accurate reflection of the Hadean boundary (Windley, 1995). The lack of any meaningful preservation of Hadean crust anywhere on the face of the Earth is a feature generally attributed to widespread destruction of this ancient material, either by intense meteorite bombardment or by subduction associated with a turbulent, rapidly convecting mantle, or both. There is also evidence to suggest that the early atmosphere and ocean formed only at the end of the Hadean era, once the main period of accretion and meteorite bombardment had terminated (Kasting, 1993). De Wit and Hynes (1995) have suggested that the Hadean Earth was also characterized by loss of heat direct to the atmosphere, in contrast to later periods of time when heat loss is largely buffered by a liquid hydrosphere. The implications for metallogenesis are that sedimentation and hydrothermal process are likely to have been inconsequential in the Hadean, and any ore deposits that did form at that time were, therefore, probably igneous in character. It is conceivable, for example that oxide and sulfide mineral segregations accumulated from anorthositic and basaltic magmas at this
time. The only preserved record of such rocks within reach of humankind at present is, however, likely to be on the Moon.

The Eoarchean refers to the dawn of Archaean time and to rocks formed prior to 3600 Ma, although for the purposes of this discussion it is considered to extend between 4000 and 3600 Ma. The best preserved section of Archaean crust that falls into this time bracket is the 3800 Ma Isua supracrustal belt and associated Itsaq (previously called Amitsoq) gneisses of western Greenland. The Isua belt comprises mafic and felsic metavolcanics, as well as metasediments, and resembles younger greenstone belts from elsewhere in the world. Although only $4 \times 30$ km in dimension, the Isua belt contains a major chert–magnetite banded iron-formation component as well as minor occurrences of copper–iron sulfides in banded amphibolites and in iron-formation (Appel, 1983). The largest iron-formation contains an estimated 2 billion tons of ore at a grade of $32\%$ Fe. Scheelite mineralization has also been found in both amphibolite and calc–silicate rocks of the Isua belt, an association which suggests a submarine exhalative origin. The coexistence of banded iron formations and incipient volcanogenic massive sulfide style mineralization points to sea-floor processes, not unlike those active throughout much of subsequent Earth history. Although the zones of known mineralization in the Isua belt are sub-economic, at 3800 years old they clearly represent the oldest known ore deposits on Earth.

1.1.2.2 The Palaeo-, Meso-, and Neoarchean stages (3600 to 2500 Ma)

The main stage of Archaean crustal evolution took place over an extended duration of more than 1000 million years, during which time geological processes were probably not too dissimilar to those of today – provided that allowances are made for features such as higher heat flow, thicker oceanic crust, and an anoxic atmosphere. De Wit et al. (1992) described the processes that took place during this period of time in terms of two principal stages, termed “intra-oceanic shield formation” (between about 3600 and 3100 Ma) and “intra- and inter-continental craton formation” (between about 3100 and 2500 Ma). The latter stage terminated, in the Neoarchean, in what might have been the most intense period of crustal development in Earth history. The Neoarchean was also characterized by extremely active ore-forming processes representing igneous, hydrothermal, and sedimentary deposit types.

The Rogers model suggests that by the end of the Archaean (2500 Ma) there might have been two major continental blocks in existence, Ur and Arctica. The two blocks would have comprised segments of ancient crust now preserved in various parts of different present day continents and it is, therefore, doubtful that their configuration in the Archaean can be known with any real certainty. A suggested configuration for Ur at around 3000 Ma shows the Kaapvaal and Pilbara cratons of southern Africa and Western Australia linked via a corridor made up of Archaean segments from India and Antarctica. The existence of Ur receives some support from the similarities that exist in the nature and ages of Archaean greenstone belts and supracrustal sequences on the Pilbara and Kaapvaal cratons (Cheney, 1996; Martin et al., 1998a), a feature that is especially striking in the similarities between the huge Superior-type banded iron-formations of the two regions. The unique occurrence of the Archaean aged Witwatersrand basin on the Kaapvaal Craton, and its apparent absence in Western Australia, on the other hand, would tend to question the relevance of comparative geology that far back in Earth history. Likewise, paleomagnetic data contradict a close fit of Pilbara and Kaapvaal (Wingate, 1998; Evans et al., 2000) so that the real paleogeography and perhaps even the very existence of Ur at all, must
remain questionable. Arctica is considered to have been made up of several ancient continental fragments (including west Greenland and the Slave province with their remnants of circa 4000–3800 million year old crust), although the actual amalgamation and of this continent is considered to have postdated Ur, at around 2500 Ma. The fit between the Siberian and Canadian portions of Arctica was originally suggested on the basis of colinearity in the trends of the Palaeoproterozoic Akitkan and Thelon magmatic arcs.

In a metallogenic context, Archaean crustal evolution can be viewed in terms of a two stage model which suggests that early “shield” formation, in which amalgamation of oceanic basaltic terranes and emplacement of early tonalite–trondhjemite–granodiorite (TTG) magmas was followed by “cratonization,” where modern plate tectonic processes such as subduction and continent collision occurred (De Wit et al., 1992).

1.1.2.2.1 Shield formation (circa 3600–3100 Ma)

This period conforms approximately with the Paleoarchean era and is typified by amalgamation of oceanic and arc-formed crust and the incipient stages of continental crust formation by emplacement of early, tonalite–trondhjemite–granodiorite (TTG) plutons (De Wit et al., 1992; Choukroune et al., 1997). This stage of Archaean crustal evolution is characterized by the development of early continental shield areas comprising highly deformed (oceanic) greenstone remnants occurring as megaxenoliths within extensive TTG terranes.

The styles of mineralization that formed during this stage of Earth evolution are limited and best exemplified by the deposits previously described for the Isua belt of west Greenland. Algoma type banded iron-formations are a common component of early greenstone belt assemblages and reflect the low oxygen levels of the atmosphere and the abundance of ferrous iron, sourced from exhalative activity at the mid-ocean ridges. The existence of oceans at this time and the likelihood of exhalative hydrothermal processes on the sea floor would have resulted in the formation of volcanogenic massive base metal sulfide deposits, although examples are rare. An exception is the well preserved Big Stubby VMS deposit, in the 3460 Ma Warrawoona Group metavolcanics of the Pilbara Craton in Western Australia (Barley, 1992).

1.1.2.2.2 Cratonization (circa 3100–2500 Ma)

This stage of Archaean crust formation coincides broadly with the Meso- and Neoarchean eras. It is suggested to be the most prolific period of crustal production in Earth history and is also a time of major global mineralization. The processes active at this time were not unlike those taking place later on in Earth history and involved widespread plate subduction, arc magmatism, continent collision and rifting, and cratonic sedimentation. An earlier sub-stage illustrates consecutive accretion of island arcs onto a previously formed continental shield and stabilization of the latter by intrusion of large granite batholiths. A later sub-stage envisages the existence of Archaean cratons consisting of numerous terranes, each bordered by major suture zones (possibly the fossilized sites of subduction or arc collision) and flanked by both active and passive margins. Sites of intracratonic sedimentation are also envisaged in this stage of development.

From a metallogenic viewpoint this stage of Archaean crustal evolution gave rise to a wide variety of ore-forming processes. Arc-related volcanism associated with plate subduction contributed large volumes of magma to the accreting terranes of the time. Well mineralized
examples of continental crust formed in the period 3100–2500 Ma are represented by the granite–greenstone terranes of the Superior Province of Canada, as well as the Yilgarn and Zimbabwe cratons. Greenstone belts are hosts to numerous important volcanogenic massive sulfide (VMS) Cu–Zn ore bodies, such as those at Kidd Creek and Noranda in the Abitibi greenstone belt of the Superior Province. Off-shore, in more distal environments, chemical sedimentation gave rise to Algoma type banded iron-formations, examples of which include the Adams and Sherman deposits, also in the Abitibi greenstone belt. Greenstone belts formed at this time also often contain komatiitic basalts that, under conditions favourable for magma mixing and contamination, exsolved immiscible Ni–Cu–Fe sulfide fractions to form deposits such as Kambalda in Western Australia and Trojan in Zimbabwe. During periods of compressive deformation, major suture zones became the focus of hydrothermal fluid flow derived from either metamorphic devolatilization or late-orogenic magmatism. This resulted in the formation of the voluminous and characteristic styles of orogenic gold mineralization which are typical of most late Archaean granite–greenstone terranes worldwide. Examples include important deposits such as the Golden Mile, Kalgoorlie district of Western Australia, the Hollinger–McIntyre deposits in the Abitibi greenstone belt, the Sheba–Fairview deposits of the Barberton greenstone belt, and the Freda–Rebecca mine in Zimbabwe. Early intracratonic styles of sedimentation, often in a foreland basinal setting, gave rise to concentrations of gold and uraninite represented by the Witwatersrand basin in South Africa. At least some of this mineralization is placer in origin and was derived by eroding a fertile Archaean hinterland. The passive margins to these early continents would have developed stable platformal settings onto which laterally extensive Superior type banded iron-formations could have been deposited. A very significant period for deposition of iron ores such as those of the Hamersley and Transvaal basins of Western Australia and South Africa respectively, as well as the Mesabi range of Minnesota, seems to have been around the Archaean–Proterozoic boundary at 2500 Ma.

1.1.3 The Proterozoic Eon

The period of time around 2500 Ma represents a major transition in the nature of crustal evolution, involving changes in the volume and composition of the continents, tectonic regimes, and atmospheric make-up. It is also clear from secular metallogenic patterns that these evolutionary changes affected ore-forming processes and characteristics. The Proterozoic Eon spans a vast period of geological time, from 2500 to 540 Ma, including the period between 2000 and 1000 Ma that was marked by a relative paucity of orogenic and/or magmatic-hydrothermal deposit types, but abundant ores hosted in intracontinental sedimentary basins and anorogenic igneous complexes. The reasons for this pattern are multifaceted and complex, but, as a first order approximation, are related to a higher degree of continental stability and the existence of major land masses which amalgamated and dispersed in relatively long-lived Wilson cycles (Windley, 1995). A substantial volume of continental crust must have been in existence by the beginning of the Proterozoic Eon and it is widely held that the first supercontinent came into existence during this time, although its form and evolution are still largely speculative (Piper, 1976; Hoffman, 1988; Park, 1995). The Rogers model envisages that, in addition to Ur and Arctica, the Palaeoproterozoic witnessed continuing amalgamation of land masses to form Atlantica (from about 2000 Ma and comprising parts of West Africa and South America; and Baltica (the Palaeoproterozoic basement to what is now Western Europe). Further amalgamation of two (i.e. Arctica and Baltica) of the four continental land masses in existence at this time is believed to have occurred at about 1500 Ma, to form a new continent that Gower (1990) and
Rogers (1996) called Nena (an acronym for Northern Europe and North America). Laurentia and Baltica were joined together between 1900 and 1500 Ma. The process of continental amalgamation continued through the Mesoproterozoic, until approximately 1000 Ma, when virtually complete consolidation gave rise to the formation of a single supercontinent, now widely referred to as Rodinia (McMenamin and McMenamin, 1990). Rodinia formed largely by accretion of Ur and Atlantica to Nena along a major, almost continuous (in some reconstructions) suture zone known, at least in North America, as the Grenville orogeny. In the Neoproterozoic, and certainly by about 700 Ma, Rodinia was broken apart again, mainly along two major rifts. One of these separated Atlantica from Nena, along a previous suture, to form two fragments known thereafter as Laurentia and west Gondwana. The other split Nena internally to separate Laurentia from east Gondwana. This configuration, although still largely conjectural and the subject of continual modification and further research (Dalziel et al., 2000), sets the scene for the subsequent pattern of crustal evolution and global tectonics in the Phanerozoic Eon.

In summary, crustal evolution through the Proterozoic Eon was characterized by intermittent continental growth (primarily at around 2000 and 1000 Ma) and amalgamation which ultimately gave rise to the existence of a stable and long-lived supercontinent in the Neoproterozoic Era. Although geological processes were not significantly different from those of other periods of Earth history, the interval between 2000 and 1000 Ma is typified by ore deposits related to anorogenic magmatism and intracontinental basin deposition. These ore-forming processes point to a pattern in which a substantial part of Proterozoic crustal evolution was characterized by relatively long-lived periods of continental stability. Continental stability at this time may have been promoted by an asymmetry in the distribution of crustal type, where a largely oceanic hemisphere was antipodal to a static continental domain. This feature might have been further accentuated, at around 1000–700 Ma, by the presence of a Rodinia. A similar situation is suggested to have applied during the existence of the Pangean supercontinent in Permian–Triassic times (Nance et al., 1986). The metallogenic contexts of the three Proterozoic eras are discussed in more detail below.

1.1.3.1 The Palaeoproterozoic Era (2500–1600 Ma)

From a metallogenic viewpoint, the period of earth history between 2500 and 1600 Ma is very significant because of the major changes that occurred to the atmosphere, especially the rise in atmospheric oxygen levels at around 2200 Ma. Prior to this time, the major oxygen sink was the reduced deep ocean where any photosynthetically produced free oxygen was consumed by the oxidation of volcanic gases, carbon, and ferrous iron. In this environment banded iron-formations, as well as bedded manganese ores, developed, as is evident from the widespread preservation of both Algoma and Superior type iron deposits. The increase in ferric/ferrous iron ratio in the surface environment that accompanied oxyatmoinversion at 2200 Ma, and the accompanying depletion in the soluble iron content of the oceans at around this time, resulted in few BIFs forming after this time. The stability of easily oxidizable minerals such as uraninite and pyrite is also to a certain extent dependent on atmospheric oxygen levels and it is, therefore, relevant that major Witwatersrand-type placer deposits did not form after about 2000 Ma. Besides these nonrecurrent changes, which only affected oxygen sensitive ore-forming processes, the pattern of metallogeny in the Palaeoproterozoic followed normal global tectonic constraints.

The continent of Ur remained tectonically quiescent throughout much of the Proterozoic period although it witnessed episodic growth at around 2000 Ma (possible amalgamation with the
Yilgarn and Zimbabwe cratons) and again at about 1500 Ma. This early stability is reflected in the widespread deposition and preservation of banded iron-formations along shallow continental platforms at the Archaean–Proterozoic boundary, already mentioned. On the Zimbabwe craton, rifting at around 2500 Ma gave rise to intrusion of the Great Dyke, with its significant Cr and PGE reserves. At 2060 Ma on the Kaapvaal craton, the enormous Bushveld complex with its world-class PGE, Cr, and Fe–Ti–V reserves was emplaced, as was the Phalaborwa alkaline complex with its contained Cu–P–Fe–REE mineralization. The period between 2000 and 1800 Ma, however, was characterized by a global orogeny, largely accretionary in nature (Windley, 1995), which also affected Ur. The Australian Barramundi and southern African Kheis orogenies are events which contributed to the growth of Ur and, in the latter region, for example, gave rise to the Haib porphyry Cu deposit and small MVT type Pb–Zn deposits along the western edge of the Kaapvaal craton. A further long period of cratonic stability ensued and Ur was subjected to rifting and intracratonic sedimentation. The period between 1700 and 1600 Ma saw the deposition of large dominantly clastic sedimentary basins that host the world-class SEDEX Pb–Zn ores of eastern Australia (Mount Isa, Broken Hill, and McArthur River) and South Africa (Aggeneys and Gamsberg). Elsewhere on Ur at this time, anorogenic magmatism was also occurring, importantly in the form of the 1600 Ma Roxby Downs granite–rhyolite complex, host to the enormous magmatic-hydrothermal Olympic Dam iron oxide–Cu–Au deposit in South Australia.

A slightly different pattern of Palaeoproterozoic metallogeny is evident with respect to the continental fragments of Arctica and Baltica, which had combined by about 1500 Ma to form Nena. Early cratonic stability of Arctica is evident in the deposition of the Huronian Supergroup at 2450 Ma which contains the palaeoplacer uranium ores of the Eliot Lake–Blind River regions of the Superior province, Canada. Both Arctica and Baltica were, however, subjected to extensive accretionary orogenies between 2000 and 1700 Ma. The Trans-Hudson, Yavapai–Mazatzal and Svecofennian orogenies, for example, produced significant new crust within which volcanogenic massive sulfide Cu–Zn deposits such as Flin Flon in Canada, Jerome, Arizona, and the Skellefte (Sweden)–Lokken (Norway) ores of Scandanavia are preserved. Relative stability followed this period of orogenesis, during which time large intracontinental sedimentary sequences such as the Athabasca basin formed at around 1700 Ma. It should be noted that the very rich uranium ores in the latter are epigenetic and probably formed during several later episodes of fluid flow between 1500 and 1000 Ma (Hecht and Cuney, 2000). The continent of Atlantica was consolidated only after 2000 Ma, subsequent to a major compressional event reflected in west Africa as the Birimian orogeny at 2100–2000 Ma. This major crust-forming event gave rise to the important orogenic or lode-gold deposits of Ghana such as Ashanti. Atlantica was relatively stable for the remainder of the Palaeoproterozoic and saw the emplacement of anorogenic type granite magmatism at around 1900–1800 Ma, with which the large iron oxide–Cu–Au deposits of the Carajas region, Brazil, are associated.

1.1.3.2 The Mesoproterozoic Era (1600–1000 Ma)

The period of geological time after the formation of Nena (i.e. the amalgamation of Arctica and Baltica) at around 1500 Ma appears to have been one of tectonic quiescence and continental stability which lasted for several hundred million years. It culminated at around 1000 Ma in an episode of widespread orogenic activity (the Grenville orogeny and its many analogues worldwide) which resulted in the final amalgamation of the Rodinia supercontinent. Although
this period of orogenesis affected the continental margins of Ur it appears to have contributed little to the formation of mineral deposits. An exception is provided in South Africa by the magmatic Cu–sulfide ores associated with mafic intrusions of the Okiep copper district in the 1060–1030 Ma Namaqualand belt. In Nena, by contrast, enormous volumes of anorogenic magmatism, especially in the period 1500–1300 Ma, provided the host rocks to a number of very important deposits. In a belt stretching from southern California through Labrador into Scandinavia, numerous intrusions of gabbro–anorthosite host the large magmatic Fe–Ti (ilmenite) ore bodies of the Marcy massif in the Adirondacks and Lac Allard in Quebec. The same belt also contains alkali granite–ryholite complexes which give rise to Fe–Au–REE resources such as those of the St Francois Mountains of Missouri. The granite–ryholite magmatic complexes were also eroded to form the sediments of the 1440 Ma Belt basin in the northwest USA, host to the Sullivan Pb–Zn SEDEX deposit in British Columbia in Canada. In addition, intracontinental rifting at around 1100 Ma in Nena gave rise to the formation of the 2000 km long Keweenawan mid-continental rift, stretching from Michigan to Kansas and filled with a thick sequence of bimodal basalt–ryholite volcanics overlain by rift sediments. The latter form the host rocks to the stratiform Cu–Ag White Pine deposit in Michigan.

1.1.3.3 The Neoproterozoic Era (1000–540 Ma)

The Neoproterozoic commenced with the formation, at around 1000 Ma, of the supercontinent Rodinia, arguably the first substantially consolidated land-mass in Earth history. Rodinia was long-lived and only started to partially fragment after a static period of more than 250 million years. Substantial parts of what had been Rodinia then reconvened toward the end of the Proterozoic (at 540 Ma) to form the very substantial Gondwanan land-mass during the Pan-African orogeny. Break-up of Rodinia started at around 750 Ma when east Gondwana (previously referred to as Ur) rifted away from the western edge of Laurentia to initiate the opening of the proto-Pacific ocean. The remaining portion of Rodinia (i.e. Laurentia, west Gondwana, Siberia, and Baltica) drifted southwards, remaining intact while at the same time rotating clockwise (Torsvik et al., 1996). At about 650–600 Ma both Baltica and Siberia started to break away from Laurentia and the Iapetus Ocean formed. By the end of the Proterozoic Eon, the amalgamation of east and west Gondwana had taken place and this large continental mass was situated at polar latitudes and across the Iapetus seaway from an equatorially located Laurentia. In detail, the assembly of Gondwana was long-lived and polyphase and occurred progressively from about 750 to 550 Ma. The Neoproterozoic is, therefore, sometimes referred to as the period of two supercontinents and was notable for its protracted periods of continental amalgamation, enhanced freeboard, and tectonic stability. The time between about 750 and 550 Ma was also characterized by the development of at least two, and in places possibly four, major ice ages, one or more of which was near global in cover and extended to equatorial latitudes. The concept of the Neoproterozoic “Snowball Earth” (Harland, 1965; Hoffman et al., 1998) has important implications for understanding climate change and, especially, for the proliferation and diversification of organic life at the Precambrian–Cambrian boundary. Global glaciations also have implications for the nature and formation of ore deposits in the Neoproterozoic Era.

The major ore deposits of the Neoproterozoic reflect the conditions of continental stability, as well as the periods of near global ice cover and attendant anoxia, that prevailed at this time. The extensively developed ironstone ores of northwest Canada and South Australia, associated with the 750–725 Ma Rapitan and Sturtian glaciogenic rocks respectively, are considered to be the
result of the build-up of ferrous iron derived from offshore hydrothermal vents in the reduced ocean waters that accompanied the development of vast continental and oceanic ice sheets at this time. Receding glaciers and a return to more oxidizing conditions would have resulted in conversion of labile ferrous iron to insoluble ferric iron and precipitation of the latter from the ocean water column, together with clastic and glaciomarine detritus to form the ironstones (Lottermoser and Ashley, 2000). Even more oxidizing conditions would also have resulted in the precipitation of manganese oxides or carbonates in the succession.

In a similar vein, the Precambrian–Cambrian boundary at around 540 Ma is also characterized by the first major global phosphogenic event that resulted in the development of vast deposits of phosphatic sedimentary rock (phosphorites) in several parts of the world (Cook and Shergold, 1984). As with sedimentary iron ores, phosphorites reflect the upwelling of deep, nutrient-rich ocean waters onto shallow continental shelves with the syn-sedimentary precipitation of carbonate–apatite (or collophane) onto the shelf floor. Although the actual formation of phosphate ores is a complex process, there is compelling evidence to suggest that the onset of phosphogenesis at 540 Ma was related to conditions prevailing toward the end of the Neoproterozoic. Many phosphorite deposits worldwide immediately overlie glaciogenic sediments, suggesting that upwelling of phosphorus-rich ocean waters was promoted by overturn of a stagnant ocean during the widespread blanketing of sea-ice associated with a Snowball Earth scenario. The Precambrian–Cambrian phosphogenic event also coincides with the proliferation and diversification of organic life and it is pertinent that a significant proportion of organisms that evolved at this time developed calcium phosphate skeletal structures. Phosphorus is, in addition, a universal nutrient (unlike oxygen) and its concentration in the oceans at the end of the Proterozoic, and in the Cambrian, may also be linked to the evolution of organic life.

The formation of the vast, stratiform Cu–Co clastic sediment hosted ores of the Central African Copperbelt is also considered to have formed in an environment influenced by the Snowball Earth. The host Katangan sediments were deposited on a fertile Palaeoproterozoic basement in an intracontinental rift, the development of which overlapped with both the Sturtian and Marinoan glacial events. The Grand and Petit Conglomerats of the Katangan sequence, for example, represent glaciogenic sediments capped by carbonates which are correlated, respectively, with the Sturtian and Marinoan events (Windley, 1995). Influx of oxide-soluble Cu and Co, perhaps derived from the local basement, might have occurred as diagenetic fluids migrated along growth faults and through the basin during the postglacial stages of deposition. Precipitation of ore sulfides would have occurred when the metal-charged oxidized fluids encountered reduced sediments or fluids.

The enormous proliferation and diversification of organic life in the late Neoproterozoic also resulted in the first substantial development of highly carbonaceous sediments. A good example of these are the Hormuz sediments that flank the eastern Arabian shield and which could represent some of the source rocks for the vast Mesozoic oil and gas fields of the Arabian Gulf (Windley, 1995). The Neoproterozoic is notable for the paucity of orogenic type deposits. Even the extensive Pan-African orogenic belts representing the suturing of continental fragments during Gondwanan assembly are strangely devoid of, for example, world-class volcanogenic massive sulfide deposits, the latter being relatively abundant in the Palaeoproterozoic orogenic belts of the world. A few small examples do nevertheless occur, such as the deposits of the Matchless amphibolite belt in Namibia, Bleïda in Morocco, and the Ducktown, Tennessee
deposit (Titley, 1993). The global shortage of these ores in the Neoproterozoic perhaps reflects either a preservation factor or simply lack of exploration success.

1.1.4 The Phanerozoic Eon

By comparison with earlier eons, crustal evolution during the Phanerozoic is well understood and there is a large measure of confidence that accompanies the interpretation of tectonic, chemical, and biological processes over the past 540 million years. The latter half of this period, the Mesozoic and Cenozoic eras, has been accompanied by a prolific development of mineralization, the likes of which has probably not been seen since the late Archaean bonanza, despite the influence of a preservation factor. Ore-forming processes and their relationships to the pattern of Earth evolution in the Phanerozoic are also reasonably well understood and additional details regarding some of the topics discussed below can be found in Nance et al. (1986), Larson (1991), Barley and Groves (1992), Titley (1993), Kerrich and Cassidy (1994), Windley (1995), and Barley et al. (1998).

A significant proportion of the Phanerozoic was characterized by geological processes that reflect a Wilson cycle, namely the sequence of events that saw the dispersal of Gondwana in the early Palaeozoic, followed by reamalgamation of continental material to form the Pangean supercontinent, by the early Mesozoic. The remainder of the Phanerozoic has witnessed the start of another Wilson cycle, involving the dispersal of Pangea to form the present day continental geography. It seems likely that we are presently about half way through this Wilson cycle (Nance et al., 1986) and that it will be responsible for reassembling a significant proportion of the continents over the next 100–200 million years. Continued continental amalgamation in the future will extend processes currently taking place, such as the collision of the Indo-Australian plate (previously part of east Gondwana) with the combined Baltic–Siberia (now called the Eurasian plate) and closure of the Mediterranean Sea.

As previously mentioned, break-out of Laurentia from Rodinia and rotation of Ur at around 725 Ma led to the formation of a Gondwanan land-mass by the end of the Proterozoic. Dispersal of Gondwana in the Cambrian–Ordovician may have been facilitated by the earlier development of a superplume (Larson, 1991), evidence for which is seen in the formation of major dyke swarms in different parts of Gondwana at 650–580 Ma (Torsvik et al., 1996). The Iapetus Ocean formed as Laurentia drifted away from Baltica and Gondwana from the late Precambrian onwards. The pattern of continental fragmentation was accompanied by a decrease in continental freeboard, sea-level highstand and marine transgression. This was followed by progressive basin closure, terrane accretion, and granitoid magmatism during the period extending from the Silurian through to the early Carboniferous (about 420–300 Ma). This collisional phase commenced with the rapid consumption of Iapetus as Baltica and Avalonia (i.e. a small terrane comprising England, Wales, southern Ireland, and eastern Newfoundland) moved toward lower latitudes and collided with Laurentia in Silurian times. The orogenies that reflect this collision phase are referred to as the Caledonian in Scandinavia and Scotland/Ireland and Appalachian in the eastern USA. Continental amalgamation continued as Gondwana drifted northwards toward lower latitudes, consuming the Rheic Ocean during the Devonian and Carboniferous, and eventually colliding with Laurentia–Avalonia–Baltica in the late Carboniferous–Permian. These complex and polyphase collisions are referred to as the Variscan and Hercynian orogenies (the terms are
essentially synonymous) in present day Europe and the Alleghanian orogeny in the eastern USA. Continental amalgamation continued into the Permian with the accretion of Siberia to Baltica along the Urals suture, after which the Pangean supercontinent had essentially formed.

The break-up of Pangea commenced soon after the time of its maximum coalescence in the Triassic at around 230 ± 5 Ma (Veevers, 1989). It was, therefore, a very short-lived supercontinent compared to Rodinia, which appears to have endured during much of the early Neoproterozoic. One reason for this might have been the development of a superplume, or plumes, that existed for some 70 million years, in Carboniferous and Permian times (320–250 Ma), an interval that also coincided with a protracted period of constant reversed magnetic polarity (Larson, 1991). The effects are seen geologically by increased production of oceanic crust and global volcanism (e.g. the outpouring of the Siberian continental flood basal province and also the central Atlantic magmatic province), eustatic sea-level rise, increased atmospheric greenhouse gas production (with associated warming and organic proliferation), and enhanced deposition of organic-rich black shales. The first land-masses to actually drift apart, more than 70 million years after initial plume-related rifting, were Laurentia from Gondwana in the Jurassic at around 180 Ma. West and east Gondwana also started to split at this time, with the development of the proto-Indian Ocean and outpouring of the Karoo igneous province. The opening of the Atlantic Ocean was particularly prevalent in the early Cretaceous (around 130 Ma) and this extensional phase was also marked by continental flood basalt volcanism of the Etendeka–Parana provinces in Namibia and Brazil respectively. Continued dispersal of Pangea was stimulated by a second superplume event in the mid-Cretaceous between 120 and 80 Ma. The evidence for this event is again seen in a 40 million year period of constant, normal magnetic polarity, as well as the expected increases in oceanic crust production, eustatic sea level, atmospheric temperatures, organic productivity, and black shale deposition. The very large oceanic plateaux of the western Pacific (e.g. Ontong–Java) are a magmatic reflection of this mid-Cretaceous event, as is the development of very significant global oil reserves related to the surges in nutrient supply and organic productivity on the voluminous continental shelves formed by the rise in sea level (Larson, 1991). In post-plume times the late Cretaceous saw the final separation of Indo-Australia from Antarctica, propelling the former northwards and resulting in continent–continent collision with Eurasia and the formation of the Himalayan orogeny in the Cenozoic. The Alpine orogeny of southern Europe was more or less coeval with its Himalayan counterpart and resulted from the collision of the African and Arabian plates with Eurasia after consumption of the Tethyan Ocean.

A very important component of Pangean breakup in the Mesozoic–Cenozoic is reflected in the development of new crust that accompanied the Cordilleran and Andean orogenies along the western margins of North and South America (Windley, 1995). This huge and complex orogeny, extending along the ocean–continent interface of western Pangea, occurred in response to subduction and translocation of the original Panthalassan Ocean (now the eastern Pacific) beneath Laurentia (now North America), the Cocos plate (now central America), and segmented west Gondwana (now South America). An island arc lay to the west of North America during Pangean times and this was accreted onto the continental margin, together with numerous other exotic terranes, during the consumption of several thousand kilometers of the Pacific beneath North America during the Mesozoic Era. The enormous continent-faced magmatic arc that resulted from this subduction gave rise to the 130–80 Ma granite batholiths and felsic volcanic rocks (such as the Sierra Nevada batholith) of the Cordillera. Continued subduction, albeit at a shallower angle than previously, and crustal thickening during the early Cenozoic, gave rise to in-
board extension and exhumation of metamorphic core complexes. Ongoing subduction during the Laramide orogeny (80–40 Ma) continued to feed the magmatic arc and numerous, metallogenically important I-type granite batholiths were emplaced at this time. As subduction waned in the late Eocene–Oligocene the magmatic arc migrated westwards, resulting in continued calc–alkaline magmatism. Thermal collapse commenced in the Neogene and the resulting crustal extension ultimately led to the formation of the Basin and Range province. In South America, Andean evolution was analogous, but less complex, than further north, and was also characterized by a lesser degree of accretionary and extensional processes. Subduction of the Nazca plate beneath a Palaeozoic sedimentary margin, in the late Cretaceous, gave rise to a volcanic arc, behind which back-arc sedimentation took place (Lamb et al., 1997). This arc was the site of protracted magmatism that built the Western Cordillera and gave rise to the volcanic edifices which host the very important porphyry and epithermal styles of polymetallic mineralization in Peru, Bolivia, and Chile. Continued subduction transferred compressional stresses in-board and created a thinskin fold and thrust belt in what is now referred to as the Eastern Cordillera. In the late Oligocene, crust was shortened even more and uplift gave rise to rapid erosion of the mountain belt to form a thick sedimentary plateau of dominantly Miocene-aged gravel and red-bed sequences known as the Altiplano. Subduction-related volcanism continued throughout this period in the Western Cordillera, which was also elevated to form the high Andes and then eroded to contribute sediment into the Altiplano basin as well as westwards onto the coastal plain. Further to the east, magmatism commenced in the early Miocene as a result of convective removal of the basal lithosphere and crustal melting beneath the Altiplano (Lamb et al., 1997). The buoyancy and elevated profile of the Andes is maintained by present day underthrusting of the Brazilian shield beneath the Eastern Cordillera.

1.1.4.1 Tectonic cycles and metallogeny

The relatively well defined global tectonic cycles of the Phanerozoic Eon, summarized above, are also clearly reflected in secular metallogenic trends. Tittley (1993), for example, noted that the distribution of stratabound ores (i.e. volcanogenic massive sulfide, clastic sediment hosted Pb–Zn (SEDEX), and Cu (red-bed) deposit types) could be related to the tectonic cycles of Pangean amalgamation and break-up. Preferential development of VMS deposits appears to be associated with periods in the Wilson cycle of elevated sea levels (highstand) associated with continental dispersal, namely, after Gondwana break-up in the early Palaeozoic and in post-Pangean Mesozoic times. This association was considered to reflect the processes of rifting, enhanced ocean crust production, and hydrothermal exhalation that accompany continental dispersal. Similar patterns are also evident in the accumulation of organic-rich shales and oolitic ironstone ores, which preferentially occur in the same two intervals, namely after Gondwana break-up in the Ordovician–Devonian and in post-Pangean Jurassic–Cretaceous times. In these cases, increased exhalative activity, carbon dioxide production, global warming, and organic productivity are interrelated processes which result in suitable conditions for black shale and ironstone precipitation in the oceans. In contrast, clastic sediment hosted base metal ores of the SEDEX Pb–Zn–Ag and red-bed Cu types tend to form at different stages of the Wilson tectonic cycle. These ores are preferentially developed at the time of maximum continental amalgamation and stasis, namely in Gondwana during the Precambrian–Cambrian transition and in Pangea during the late Palaeozoic–early Mesozoic. These ore types appear to have evolved preferentially during intracratonic rifting and sea-level lowstand (Titley, 1993). The two Phanerozoic superplume events described above are also related in time to periods of enhanced organic
productivity and formation of fossil fuels. The mid-Cretaceous superplume, active between 120 and 80 Ma, coincides with the formation of voluminous organic-rich shales deposited largely in the low-latitude Tethyan seaway. These shales are thought to be genetically linked to some 60% of the world’s oil reserves (Larson, 1991). The late Carboniferous–Permian superplume is similarly associated with deposition of a large proportion of the world’s coal reserves, formed between 320 and 250 Ma, again at a time when organic productivity increased and tropical, swampy conditions prevailed along flooded continental margins associated with the corresponding sea-level rise. Continental hotspots, or beheaded plumes, are also possibly related to the preferential development of alkaline and kimberlitic magmatism associated with enhanced igneous activity in the mid-Cretaceous and again in the Cenozoic. Many of these intrusions, important for Cu–REE–P–Fe–Au mineralization and also diamonds, are found on old, stable cratons and located along ancient lineaments that might have been reactivated during extension and crustal thinning associated with hotspot activity.

1.1.4.2 Time-bound and regional aspects of Phanerozoic metallogeny

A summary of ore deposit trends as a function of the major Phanerozoic tectonic events reveals that the metallogenic inventory, especially of convergent margin lithosphere, increases with time and with each tectonic cycle (Barley et al., 1998). The distribution of ore deposit types also very often has a regional character to it and, as mentioned previously, may reflect the extent of erosion in different regions and the preservation potential of ores.

The large-scale development of new crust along the convergent Mesozoic–Cenozoic plate margins of the Americas did not occur on the same scale during the Palaeozoic dispersal stage. Hence, a cyclic pattern in the development of orogenic magmatic-hydrothermal ores is not as evident as it is for statabound deposits, where controls on ore formation are different. There are, for example, few known porphyry Cu–Mo type deposits of Palaeozoic age, although sub-economic examples are recognized along convergent margins such as the Caledonides of Scotland. Orogenic gold mineralization is also related to the late stages of collisional orogens. Cretaceous examples of this style of gold mineralization are evident in California, British Columbia, China, and New Zealand and could conceivably be related to the enhanced thermal regime related to the superplume break-out at this time. It has also been suggested that these gold provinces might rather have been linked to areas where mid-ocean ridge, as opposed to normal oceanic crust, was subducted. Scenarios in which hot crust was subducted, compared to periods when normal subduction of cold oceanic crust prevailed, have been suggested as an explanation for the anomalous thermal conditions required for widespread gold mineralization (Haeussler et al., 1995).

The amalgamation stage of Pangea commenced with the collision of Avalonia–Baltica with Laurentia during the polyphase Palaeozoic Caledonain–Appalachian orogenies. VMS deposits formed in response to these processes in, for example, the Bathurst–Newcastle district of New Brunswick, the Buchans area of Newfoundland, and Scandanavia. New Brunswick also has examples of Devonian aged Sn–W–Mo–F granites, some of which are mineralised. The carbonate-hosted MVT Pb–Zn deposits of the southeastern USA are also related to the Appalachian orogeny and formed when the thrust front expelled basinal fluids into the carbonate platform on the continental margin, giving rise to epigenetic hydrothermal mineralization. The Variscan orogeny, formed when Gondwana collided with an already amalgamated Laurentia–
Baltica–Avalonia, is also characterized by carbonate-hosted ore bodies. Some of these, particularly the Irish deposits at Navan, are transitional in style between MVT and SEDEX-type Pb–Zn deposits, but are also related to compression-related, gravity- and tectonic-driven, basinal fluid flow. The Variscan orogeny of Europe is particularly well known for its association with widespread, 300–275 Myr old, S-type Sn–W–U granites. These are well mineralized in many of the historic European mining districts, such as Cornwall in southwest England (South Crofty, St Just, etc.), the Massif Central of France (Bellezane), the Bohemian massif of the Czech Republic, and Portugal (Panasqueira). Other deposits associated with the Variscan of Europe include the Devonian aged Rammelsberg and Meggen SEDEX-type deposits of Germany and the numerous VMS deposits of the Iberian Pyrite Belt (Rio Tinto, Neves Corvo, etc.). The brief interval in the Permian–Triassic of relative stability that accompanied the existence of the Pangean supercontinent was one in which little mineralization formed. An exception to this was the stratiform red-bed hosted Cu–Ag ores of the Permian Kupferschiefer in Poland and Germany.

Mineralization associated with post-Pangean break-up and orogenesis is both spectacular and widespread. Exceptions to this would appear to be the products of continent–continent collision, such as regions affected by the Himalayan orogeny, which appear to be devoid of any world-class ore deposits. The Karakoram mountains of Pakistan, for example, contain rich gemstone deposits (such as the rubies of the Hunza area), but otherwise only minor MVT and vein-related precious and base metal type ores, as well as a few small hydrothermal uranium deposits, occur (Windley 1995). More significant stratabound and ophiolite-hosted styles of mineralization are contained in rocks caught up in the Himalayan orogeny, but the ore-forming processes probably predate the latter. Regions affected by the Alpine orogeny are, likewise, not particularly well mineralized, although exceptions do occur. MVT-type Pb–Zn deposits occur in mid-Triassic limestones of the eastern Alps and North Africa. The Carpathian arc of southeastern Europe contains porphyry Cu–Mo styles of mineralization, such as at Sar Cheshmeh in Iran and Bor in the former Yugoslavia, and also contains significant potential for epithermal gold mineralization. Important ophiolite-hosted VMS Cu–Zn and chrome mineralization occurs in numerous obducted remnants of ocean floor, in places such as Troodos, Cyprus and in Albania and Turkey. Again, however, these deposits are likely related to ore-forming processes that predate the Alpine orogeny.

The Andean and Cordilleran orogens of the western Americas contain the greatest concentration of metals on Earth, and are pervasively mineralized from one end to the other. During the Jurassic–Cretaceous period of terrane accretion in North America, small VMS-type Cu–Zn deposits were formed, as were the “mother-lode” hydrothermal gold systems, hosted in obducted oceanic material in California, British Columbia, and Mexico. Mesozoic and early Cenozoic plutonism gave rise to the numerous world-class porphyry Cu–Mo deposits of the USA and Canada, as well as the related skarn and polymetallic epithermal vein deposits, such as the Bingham system in Utah. Thermal collapse and the development of Basin and Range extension gave rise to the enormous Eocene (40 Ma), hydrothermal, sediment hosted Carlin-type Au deposits of northeast Nevada, followed by Miocene volcanism and widespread development of epithermal Au–Ag mineralization such as the Comstock Lode of Virginia City, Nevada. Similarly, in South America a wide variety of mineralization styles are associated with Mesozoic–Cenozoic orogeny. During the Jurassic–Cretaceous period island arc volcanism and associated hydrothermal activity gave rise to the Fe oxide–phosphate ores now preserved in the western, coastal portions of the belt. The axis of subduction related magmatism commenced in the west during the late Triassic and early Jurassic and then migrated eastwards with time through
the Cretaceous and early Cenozoic (Sillitoe, 1976). In the Eastern Cordillera, however, minor magmatism occurred early in the Mesozoic, but most of the magmatism took place in the Miocene. The enormous porphyry copper deposits of Chile (such as Chuquicamata and El Teniente) and Peru (Morococha and Toquepala) formed in response to diachronous episodes of subduction-related magmatism. Many of the important porphyry copper deposits in Peru are Paleocene in age, whereas in northern Chile they are Eocene–Oligocene and, further south, Miocene–Pliocene (Sillitoe, 1976). East of the porphyry copper belt a broad zone of vein-related and skarn type Cu–Pb–Zn–Ag mineralization occurs, spatially related to Eocene–Oligocene calc–alkaline plutons. The deposits are most prolific in Peru (Antamina and Cerro de Pasco) but the belt continues southwards into Bolivia and Argentina. Finally, a distinct zone of Sn–W–Ag–Bi mineralization occurs mainly in the Eastern Cordillera of Bolivia and southern Peru, in the elbow of the central Andes. These ores occur as both vein types associated mainly with Mesozoic intrusions, and disseminated porphyry type deposits (such as Llallagua and Potosi in Bolivia) associated with Oligocene–Miocene granites.

Also extremely important with respect to mineralization associated with post-Pangean orogenesis is the extensive chain of island arcs in the northern and western Pacific. The northern and western Pacific is predominantly characterized by collisions of oceanic crust (as opposed to the ocean–continent collisions of the western Americas) which form island arcs that build new crust on oceanic (simatic) basement rather than on continental (sialic) basement. There are several types of island arc, each with its particular metallogenic character. These include intraoceanic arcs (such as Tonga, New Hebrides, and the Solomons), island arcs separated from continental crust by a narrow back-arc sea (such as Japan), and those built directly against a continent (such as Java–Sumatra). Soon after initiation of ocean–ocean collision in the western circum-Pacific region, early, relatively mafic (andesitic–dacitic), stages of calc–alkaline magmatism resulted in porphyry Cu–Au deposit formation, examples of which include Grasberg, Indonesia, and Bougainville in Papua New Guinea. Besshi-type VMS deposits also occur in back arc settings where andesite–dacite volcanism and ocean floor exhalative activity occurs synchronously with deep water sedimentation. During the main stages of arc construction and calc–alkaline magmatism, dacite–rhyolite volcanism occurred and is associated with the development of Kuroko-type VMS deposits, examples of which are known from the Miocene-aged Green Tuff belt of Japan. Also very important, in geologically more recent times, is the formation of large epithermal Au–Ag deposits. Several world-class ore deposits of this type occur in the western circum-Pacific region, such as Baguio and Lepanto in the Philippines, Hishikari in Japan, Ladolam and Porghera in Papua New Guinea, and Emperor in Fiji. Porphyry Cu–Au styles of mineralization are commonly known to occur beneath such epithermal deposits, such as at Baguio and Ladolam.

1.2 References


### 2. Geochronological history of mineralisation in the DRC


#### 2.1 Archaean rocks in the DRC

I. Congo Méridional  
   a. Kwango-Kasai-Lomami  
      - Gneiss de la Haute Luanyi / Complexe gabbro-noritique et charnockitique du Kasai-Lomami / Complexe granitique et migmatitique de Dibaya  
   b. Haute Lomami  
      - Formations anté-Kibariennes indifférenciées, y compris les granites  

II. Congo Sud-Oriental  
   a. Région Méridional  
      - Anté-Kibaran of Central Katanga  
   b. Région Septentrional  
      - Complexe métasédimentaire et cristallin des Muhila  

III. Congo Oriental  
   a. Région Méridional  
      - Formations Anté-Rusizianes et/ou Rusizianes  
   b. Région Septentrional  
      - Gneiss du Mont Speke
IV. Congo Nord Oriental
   - Complexe gneissique de la Garamba
V. Congo Nord Occidental
   a. Région Orientale
      - Complexe amphibolitique et gneissique du Boma
   b. Région Occidental
      - Complexe métasédimentaire et migmatique de l’Ubangi
VI. Congo Occidental
   - Complexe de gneiss et migmatites de Mpozo-Tombagadio

2.2 Proterozoic rocks in the DRC

2.2.1 Palaeoproterozoic rocks in the DRC

I. Congo Méridional.
   a. Kwango-Kasai-Lomami
      - Luizien
   b. Haute Lomami
      - Formations anté-Kibariennes indifférenciées, y compris les granites
II. Congo Sud-Oriental
   a. Région Méridional
      - Anté-Kibaran of Central Katanga et Shaba Méridional
   b. Région Septentrional
      - Rusizian
III. Congo Oriental
   a. Région Méridional
      - Rusizian
   b. Région Septentrional
      - Kibalien
IV. Congo Nord Oriental
   - Kibalien
V. Congo Nord Occidental
   a. Région Orientale
   b. Région Occidental
      - Ganguen
VI. Congo Occidental
   - Zadinien

2.2.2 Mesoproterozoic rocks in the DRC

I. Congo Méridional.
   a. Kwango-Kasai-Lomami
      - Complexe Sédimentaire et volcanique de la Lulua
   b. Haute Lomami
      - Kibarien
II. Congo Sud-Oriental
   a. Région Méridional
      - Kibarien
   b. Région Septentrional
      - Rhyolites des Marungu et sediments associés

III. Congo Oriental
   a. Région Méridional
      - Burundien
   b. Région Septentrional
      - Formations de la Bilati, Formations de la Luhule-Mobissio

IV. Congo Nord Oriental
   ///////////////

V. Congo Nord Occidental
   a. Région Orientale
      ///////////////
   b. Région Occidentale
      - Liki-Bembien

VI. Congo Occidental
   - Mayumbien

2.2.3 Neoproterozoic rocks in the DRC

I. Congo Méridional.
   a. Kwango-Kasai-Lomami
      - Bushimay
   b. Haute Lomami
      - Luamba

II. Congo Sud-Oriental
   - Katanga Supergroup

III. Congo Oriental
   - Lindien

IV. Congo Nord Oriental
   - Lindien

V. Congo Nord Occidental
   - Lindien (Ubangien)

VI. Congo Occidental
   - Ouest Congolien

2.3 Phanerozoic rocks in the DRC

2.3.1 Palaeozoic rocks
   - Permien inférieur – Carbonifère Superieur : Série de la Lukuga

2.3.2 Mesozoic rocks
   - * Triassic
     * Trais – Lias : - Série de la haute Lueki
- Série des « Roches Rouges »

- **Jurassic**
  * Jurassique supérieur : Série de Stanleyville

- **Cretaceous**
  **Undifferentiated**
  * Crétacé indifférencié, principalement inférieur (Int)
  * Wealdien: Série de la Loia (Int)
  * Crétacé inférieur : continental, grès sublittoraux (ZL)

**Lower Cretaceous**
* Albien – Aptien : Série de Bokungu (In)
* Albien marin : Mavumu inférieur (ZL)
* Albien – Aptien continental : Mavumu supérieur (ZL)

**Upper Cretaceous**
* (supra)-cénomanien :Série de Kwango, couches de Boende (In)
* Maestrichtien, campanien, santonien, coniacien, turonien (ZL)

2.3.3 Cenozoic rocks

- **Paleogene** : - Paleocene
  - Eocene
  - Oligocene
  * Paleogene : série des grès polymorphes (« Kalahari » inférieur)

- **Neogene** : - Miocene
  - Pliocene
  * Neogene : série des sable ocre (« Kalahari » supérieur)
  * Miocene du fossé tectonique centre africain
  * Pliocene : alluvions, éluvions et colluvions
  * Pleistocene, pliocene et miocene supérieur :« série des Cirques », Quelo (ZL)

- **Quaternary** : - Pleistocene
  - Holocene
  * Pleistocene : alluvions, éluvions et colluvions
  * Holocene : alluvions modernes


About 60 years ago, the understanding of global tectonics experienced a revolutionary advance (Kearey et al. 2009). Until then, the Earth's crust was considered to move either up or down, but rarely in a horizontal direction. The new concept of plate tectonics recognized that the lithosphere is divided into a number of rigid plates, which display considerable lateral movement. The engine of plate tectonics is convective cooling of the mantle. The resulting lithosphere is in part recycled back into the mantle. Extensional and compressional interactions at plate boundaries are the cause of profusely fertile metallogenetic systems. The Theory of Plate Tectonics was worked out only recently, but its foundations are much older. The similarity of the coastal geometry of South
America, Africa and India, and their sharing Permian sediments with the striking Glossopteris flora made already Eduard Suess (1831-1914) speculate that continents were not fixed in time and that in the geological past, the three formed a large supercontinent that he called Gondwana (Suess 1851. Building on this, Alfred Wegener developed the hypothesis of continental drift (Wegener 1924) that is in large parts still valid. Modern understanding of plate tectonics led to great progress in many fields of the earth sciences, including metallogeny (Robb 2005, Sawkins 1990b). Several lines of evidence indicate that plate tectonics may have started to operate as early as 4.4 Ga when a stiff lithosphere had been established (Moyen et al. 2006, Furnes et al. 2007). Already in the Archaean, ore deposits are known that suggest a supra-subduction zone setting. Main elements of plate tectonic process systems that are "metallogenetic factories" include:

3.1 The formation of intracontinental rifts, aulacogens and large sedimentary basins (incipient divergent plate boundaries)

Rifts originate by extensional deformation of lithospheric plates and may or may not evolve into a new plate boundary. Very often, rifting causes thinning of the crust, upflow of hot mantle and updoming of rift shoulders (Tackley, 2000). Volcanic activity within the rifts is a frequent consequence often organized into large volcanic centres (“hot spots”, Foulger & Natland 2003). Hot spots can be the origin of three diverging rifts (triple junction). Two of the three rift arms may widen to form a new ocean, whereas the third remains inactive and is called a failed rift arm. Several failed arms display thick sediments with bimodal volcanic rock suites, which were later folded by horizontal shortening. Considerable intrusive activity may occur. Settings like this have been called aulacogens (Eriksson & Chuck 1985).

Sediments of continental rifts include early, mainly terrestrial, alluvial clastic infill that can contain uranium, placers and coal deposits. In many cases, a freshwater, saline or marine-influenced lake stage succeeds with beds of salt, gypsum, magnesite, phosphate, valuable clays or oil shale. Full marine ingestion into the widening rift and inception of oceanic spreading can induce submarine metalliferous exhalation of the black smoker or brine pool type (Red Sea) and the deposition of thick marine sedimentary sequences. Later, as diagenesis is enforced by rising temperature and pressure, oil and natural gas deposits are generated.

Hot spot-related ore-forming systems include the Bushveld in South Africa, tin-fertile A-granites in Nigeria and worldwide, many alkali-carbonatite igneous complexes. When rifting reaches the stage of a deep graben with vertical displacement at marginal faults approaching kilometres (Seholz & Contreras 1998), hydrothermal convection systems may form, based on the permeable tensional structures, the heat contrast and the hydraulic head imposed by rift shoulder mountains. The ascending branch of these hydrothermal systems typically results in deposits of lead, zinc, silver, manganese, fluorine and barite, which take the form of veins and metasomatic replacement bodies in rift margin rocks, or of ore beds in the graben sediments. Good examples are many Pb-Zn and Mn occurrences in Tertiary sediments on both sides of the Red Sea, and part of the Ag-Pb-Zn-F-Ba veins along the Rhine graben in France and Germany. Carbonatites and alkali intrusions with apatite, fluorine, niobium and rare earth element ores characterize the Cretaceous-Tertiary rifts in Eastern and Central Africa. Submarine, epicontinental rifts and half-grabens are related to base metal deposits of the sedex type. Sullivan in British Columbia, Canada (base metals in the Neoproterozoic Alberta Rift Mt Isa in Queensland, Australia [Pb-Zn-Cu in the early
Mesoproterozoic) and the large deposits of native copper in basalts and of chalcocite in fine sands of the Nonesuch Shale in the Keweenawan Rift (USA, Late Mesoproterozoic) were proposed as remarkable examples of mineralization in aulacogens. The Panafrikan Damara Orogen in southern Africa has also been interpreted as an aulacogen, although with exceptionally strong tectonic shortening. Its main mineralizations are late to posttectonic, including the giant hydrothermal karst pipe Tsumeb with polymetallic ores of Pb, Zn, Cu, Cd and Ge (Chetti & Frimmel 2000), and the uranium-deposit Rössing in alaskitic granite.

Intracontinental basins with prominent ore provinces include the European Copper Shale (Mesozoic), Witwatersrand gold (Late Archaean) and Mississippi Valley type lead-zinc-barite-fluorite deposits (Palaeozoic) in North America. Major plate reorganizations affect both continental and oceanic systems intensely (Whittaker et al. 2007). Within short periods of a few million years, new subduction zones are installed, vectors of plate drift change (wander paths form "loops") and the plates are subjected to new stress fields. Oceanic and continental crust is stretched or sheared, new mantle regions experience partial melting, resulting in magma underplating, the formation of hotspots and the rise of mantle volatiles. Flood basalt volcanism may be a consequence, producing giant Cu-Ni-PGE deposits such as Noril'sk, as well as climate change and global extinction of life due to huge emissions of sulphur and chlorine such as those of the Dekkan traps at the end of the Cretaceous (Self et al. 2008).

Enhanced heat flow and elevated permeability of the crust are favourable factors for the formation of deep convective hydrothermal systems and of mineral deposits. Mantle volatiles (mainly water and CO2, with solutes like fluorine, arsenic, etc.) may rise, leach metals, mix with crustal fluids and form ore deposits. Examples of mineralization caused by plate reorganization include the hydrothermal "Saxonian Mineralization" of Europe north of the Alps (Walther 1983), several kimberlite provinces and unconformity uranium ore deposits in Canada and Australia.

3.2 The evolution of passive continental margins and the disruption of older ore provinces (divergent plate boundaries)

The opening of new oceans passes from a high heat-flow rift stage into a marine transgression and thermal contraction phase. Relatively shallow, epicontinental seas may form. As the young ocean widens, passive continental margins develop. Sediments include salt, phosphate and hydrocarbon source rocks. Manganese ore beds of the Tertiary Black Sea province, Quaternary metalliferous marine placers and Palaeoproterozoic banded iron ores of the Superior type represent typical marine epicontinental shelf ore deposits.

The separation of continents by rifting and seafloor spreading may cut across older orogenic belts, cratons and other crustal-scale structures. With them, older ore provinces are ruptured and the fragments can be found on remote coasts across an ocean (e.g. the Atlantic borderlands of Africa and South America). In these cases, metallogenetic knowledge acquired on one coast is a valuable tool for work in its twin across the seas.

Seafloor spreading and the production of new lithosphere at mid-ocean ridges (oceanic divergent, or "constructive" plate boundaries)

This is the domain of ore formation at mid-ocean ridges that was presented earlier in more detail. After obduction, the products of these processes are ophiolite-hosted deposits. Many ophiolites,
however, were not formed at mid-ocean rifts but in tensional supra-subduction settings including back-arc spreading systems, or rifts of primitive island arcs (e.g. the Cyprus ophiolite). Yet, there is no doubt that all midocean rifts display segments of hydrothermal activity, including black smokers. Related ores are sulphide mounds or mud-pools in a proximate position, iron-manganese oxides (ochres and numbers) and distal manganese crusts and nodules with important contents of Cu, Ni and Co. Oceanic transform faults that offset ridges are apparently not metallotects for mid-ocean metallogenesis.

3.2.1 Subduction of lithospheric plates at convergent (“destructive”) plate boundaries

Subduction recycles oceanic lithosphere back into the mantle. The trace of subduction on the seafloor is marked by deep oceanic trenches. Volcanic arcs develop on the overriding plate. Between trench and arc, four structural zones are typically developed: Nearest to the trench an accretionary complex of low-grade metamorphic sediments is followed by a wedge of mainly continental crust with minor oceanic and hydrated mantle material of medium to high-pressure metamorphic grade. This is overlain by a mega-scale “mélange” composed of high-pressure and ultra-high-pressure oceanic and continental crust fragments that are extruded from the subduction channel. Finally follows the frontal part of the upper plate that carries the volcanic arc. Volcanic arcs in dominantly oceanic settings form island arcs, whereas active continental margins display continental or Cordilleran arcs. Recent primitive island arcs are geologically young (Tonga Scotia), because maturation sets in quickly and produces arcs with a partially continental character (Japan, Kurile Islands). Other Island arcs pass into continental collision belts (Sumatra-Malaysia-Himalaya). Andean volcanic arcs build upon older, strong upper crust that is largely of Precambrian age in both Americas. Behind the magmatic arcs appear back-arc spreading systems that include the back arc basins of Island arcs, the continental “molasse” basins and broad distended regions such as the Basin and Range Province of North America.

There is a great diversity of subduction zone configurations, due to many variables including slab density, thickness and length (Schellart, et al. 2007). Subduction zones show variably high or low trenchward plate velocities, trench retreat (or more rarely trench advance) velocities, slab dip angles and so forth. Trench retreat (“subduction rollback”) is caused by the negative buoyancy of the cold, dense descending slab. This places the overriding lithosphere into a state of tension as the subduction zone moves oceanwards and facilitates movement of magmas and fluids (Hamilton 1995). Slab rollback, slab breakoff and delamination of mantle lithosphere allow asthenospheric upwelling that can provide the heat pulses required for ore forming processes, including magmatism and regional hydrothermal fluid systems. Extensive intracontinental compressional deformation migrating cratonwards is explained by flat-slab subduction. This is probably caused by the subduction of oceanic plateaus and plume tracks (Livaccari et al. 1981), which typically ends in delamination (foundering) of the slab from the continental lithosphere. This is the environment of Basin-and-Range type tectonic and magmatic provinces (e.g. in Mesozoic South China: Li & Li 2007). Fertile anorogenic magmatism including alkaline basalts, bimodal volcanic rocks and I- and A-type granitoids are characteristic for this setting.

It is important to stress that most of the Earth's richest ore provinces are found above subduction zones. This is conspicuously so along the margins of the Pacific Ocean, which are largely formed by long-lived destructive plate boundaries. Associated are numerous active volcanoes, accounting
for the term "ring of fire". Reconstruction of similar settings for stages in the geological past is crucial for strategic exploration planning (Haeberlin et al. 2003).

Island arc ore deposits may be either allochthonous, which implies tectonic transport, for example of slivers of oceanic lithosphere, or autochthonous, formed within the arc. Allochthonous are first of all the ophiolite-related ores, including chromite (Cuba, Luzon) and platinum placers; Cyprus type sulphide deposits are infrequent. Lateritic nickel ore deposits (New Caledonia) are autochthonous formations. Major autochthonous deposits are associated with the large mass of calc-alkaline to potassic intrusive and volcanic rocks. Of outstanding economic prominence are porphyry and skarn copper-gold deposits, epithermal gold deposits and volcanogenic massive sulphides. Similar to continental margin arcs, sources of the metals may be subducted oceanic crust or the mantle wedge above the subduction zone. The latter was confirmed for Lihir, Papua New Guinea (McInnes et al. 1999), which is a giant epithermal gold deposit of very recent geological age (ca. 690 ka). It is significant that some metals such as tin and mercury appear only in older, more complex island arcs with a partially continental character.

Active continental margin ore deposits are often more clearly zoned compared with island arcs, as a function of increasing distance from the subduction zone. In the apparently simple geotectonic setting of Central South America, the results are long and narrow ore provinces (Sillitoe 1972). Prominent along the western coast within Precambrian basement and Cretaceous plutons are iron-apatite and Iron oxide-copper-gold (IOCG) deposits associated with hydrous intermediate magmatism (Sillitoe 2003, Oyarzun et al. 2003). A belt of giant porphyry Cu-Ma-Au deposits follows towards the east, roughly along the Neogene-Recent volcanic arc. These mines currently dominate world copper production. Near the Eastern margin of the Cordillera, a Sn-Ag belt is developed. However, neither this zonation nor all deposits in single belts are synchronous, but are the product of several regional metallogenic Episodes that range from Late Triassic (earliest tin deposits) to Cretaceous (iron and part of copper) and Tertiary ages (most of the copper and tin silver). Subduction configurations during this time changed considerably (James & Sacks 1999). The structure of the North American Cordillera is even more complex. One example is the subducting East Pacific Ridge, which is a factor that enhances metallogenic processes. Partial melting of young, hot subducting oceanic plates favours the formation of oxidized adakitic magmas and of important gold and copper-gold deposits (Cooke et al. 2005, Mungall 2002). South Alaskan gold deposits are thought to be related to ridge subduction (Haeussler et al. 1995). Another difference is the collage-like nature of the North American Cordillera that consists of many "suspect" or "exotic" terranes, which preserved distinct but interrelated geological records (Colpron & Nelson 2006). This complicates metallogenic interpretation.

Ore deposit formation above subduction zones is causally tied to the fate of the subducting lithospheric slab of oceanic crust and mantle. At midocean ridges, the crust is largely hydrated and oxidized. When the oceanic slab bends before entering the subduction zone, additional hydration appears to take place (Faccenda et al. 2009). Altered basalts, gabbros and depleted mantle peridotites enter the subduction zone as a “cold” slab at geothermal gradients of 15°C/km or less. Along the subduction plane, continental material can be scraped off (“subduction erosion”) and taken down to the zone of dehydration and melting. The highpressure low-temperature metamorphism of subduction zones converts the rocks to the typical blueschist and eclogite lithologies. Mantle rocks, oceanic crust and its sedimentary cover incur devolatilization
and possibly, partial anatexis. Dehydration processes control the structure of slabs from ca. 40 to 150 km depth (Rondenay et al. 2008). As a function of T and P, hydrous fluids, anatectic melts or supercritical liquids may be set free (Kessel et al. 2005). The latter are characterized by high trace element solubilities and consist of H$_2$O, Cl, S, CO$_2$ etc., including large ion lithophile elements (LILE: Ba, K, Rb, Cs, Ca, Sr) and other incompatible elements (U and Pb). This transfer "metasomatizes" the mantle wedge above the subduction zone and triggers widespread melting. Because of relatively high fO$_2$ (roughly from fayalite-magnetite-quartz (FMQ to FMQ +2) sulphide (S$^{2-}$) and sulphate (S$^{6+}$) coexist and combine to high total sulphur contents in melts, which favours sulphur saturation and mineralization in the upper crust (Jugo 2009). Extensive formation of sulphide melt during metasomatism and partial melting of the hot mantle wedge would be detrimental, because sulphide melts scavenge chalcophile and siderophile elements such as copper and gold from silicate melt and being heavy, tend to remain trapped in the deep crust (Mungall 2002).

Magma batches rising through the crust continue to change by complex assimilation and contamination processes, until they reach the surface as calc-alkaline melts of andesitic-dioritic nature. These magmas have only ~50% material from the mantle, the other half is derived from the crust. Intrusive and extrusive activity of continental arcs is concentrated in short pulses of 10-15 My ("flare ups") that occur during and after tectonic shortening (Ducea & Barton 2007). Porphyry copper-molybdenum-gold deposits are direct products of these processes within and above the subduction zone (Richards 2003, 2009). Among many other arguments, this can be substantiated by the observation that localization and metal contents of porphyries are largely independent of their specific setting (e.g. primitive or evolved island arcs, diverse types of active margins). The precise source of the chalcophile metals and gold – oceanic crust or mantle wedge - remains obscure (Dreher et al. 2005). A continental source, however, is implied for the metals tin, tungsten and tantalum because deposits appear only in regions with thick and old crust.

3.2.2 Continental Collision

Oceans that were consumed by subduction leave a suture in the newly welded continent, which is marked by ophiolites. One of the most remarkable and metal-endowed sutures worldwide is the Palaeozoic accretionary orogenic collage of the Altaids in Central Asia, with a length of 3000 km (Xian et al. 2009). Usually, continental collision results in the subduction of continental crust (Ampferer or A-subduction), although this is limited by the buoyancy of crustal rocks. The process in thickened crust below collisional belts and the formation of anatetic S-type granitoid melts. Less frequent are post-subduction Cu-Au porphyries and related epithermal gold deposits, which are formed where former magmatic arcs are involved in the collision (Richards 2009). Continental crust of the lower plate can be subducted to depths of more than 100 km and exhumed alter ultrahigh-pressure metamorphism. Also, collision causes giant systems of hydrothermal fluid flow involving metamorphic, basinal and meteoric fluids (Mark et al. 2007, Craw et al. 2002, Oliver, 1986). Similar features are reported from intracontinental mountain belts involving very narrow oceans (Alps, European Variscan Belt) and from purely intracontinental orogens (Kibarides in Central Africa: Pohl, 1994). Typically, collisional orogens exhibit: granitoid-related deposits of tin, tungsten, gold and rare metals, and deposits formed by migrating metamorphic fluids. Gold is especially common in this setting (orogenic gold deposits: Groves et al. 2003). Mineralization in orogenic belts is favoured by phases of extension, because melts and fluids can
more easily rise to shallow depths. Extension may be related to orogenic collapse and other post-collisional processes.

### 3.2.3 Assemblage and break-up of supercontinents

The plate-tectonic evolution of the Earth's crust follows not only the relatively short Wilson cycles (opening and closure of oceans) but also a trend of large-scale cycles of amalgamation of all continental plates into supercontinents and the following break-up. The Phanerozoic supercontinent Pangaea is well-known, existing from the Permian into the Jurassic (~300-175Ma). Mesoproterozoic Rodinia (~1100-800 Mal is generally accepted but its assemblage is more contentious because data are insufficient for a unique solution (Torsvik, 2003). Older supercontinents are even less well-defined. Supercontinents can be related to specific characteristics of the metallogenetic evolution, including the incidence of anorogenic ore formation (e.g. titaniferous anorthosite-ferrodiorite complexes) and the prevalence of continental sediment-hosted deposits (Kupferschiefer: Robb 2005). When Gondwana and Laurasia finally fused into the supercontinent Pangaea, the Variscan belt in Europe experienced a short-lived metallogenetic peak of unique fertility. Deep processes inducing initial crustal distension and break-up of Pangaea, at about the Triassic/Jurassic boundary, again produced an ore-forming heat and fluid pulse across much of Europe.

Apart from the relatively simple plate tectonic model situations described above, many quite complex interaction fields are known today. One recent example is the Gulf of California, where a subducted oceanic ridge passes along strike into a continental rift and ultimately into an intra-continental transform structure (San Andreas Fault). Only rarely, connections such as these can be reconstructed for the geological past, so that the precise plate-tectonic setting of some ore deposits may never be fully understood. Yet, the quest for solving a given plate-tectonic puzzle is always scientifically fascinating and results benefit applications of economic geology.

### 3.3 References


