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Hotspots and mantle plumes: global intraplate tectonics, magmatism and ore deposits

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Summary

Intraplate tectono-magmatic phenomena, including the emplacement of layered intrusions, and the giant dyke swarms, anorogenic (hotspot) volcanism, oceanic plataeux, rifting processes, basin formation, and geomorphological features are discussed in the context of the mantle plume theory. A review of the relationships between mantle plumes and ore deposits focuses on direct links, proxied by the emplacement of mafic-ultramafic magmas (e.g. PGE and Ni–Cu sulphides associated with flood basalts) and indirectly in rift systems where high geothermal gradients are set up in the crust above the plume, induce large scale circulation of hydrothermal fluids, which result in the generation of a wide range of ore deposits. Peak periods in the deposition of iron formations coincide with plume events in the Archeaen and Proterozoic. Passive margins, which evolve from continental breakups and triple junctions, host abundant mineral and hydrocarbon resources.

Introduction

The mantle plume theory is gaining as much support as did the new plate tectonic theory in the 1960s and 1970s, even though the first ideas on mantle plumes were published at the same time (*Wilson*, 1963; *Morgan*, 1971, 1972). Like plate tectonics, the theory of mantle plumes is a powerful concept that provides testable models for intraplate tectonic, magmatic and ore-forming processes. However, not all geoscientists embrace the mantle plume model. Indeed, alternative models have been proposed whereby intraplate volcanism (or hotspot) would be related either to plate stresses that cause fracturing and volcanism, or to shallow mantle convection (e.g. *Anderson* et al., 1992; *King* and *Anderson*, 1998; *Anderson*, 1994; *Sheth*, 1999).

In this contribution, first I review the salient concepts of the theory of mantle plumes, and examine the geological processes and near-surface manifestations that accompany the impingement of mantle plumes onto the lithosphere. In the second part, I examine the relationships, direct or indirect, between magmatic and hydrothermal ore systems and the geological processes associated with mantle plume activities. It should be pointed out that of the three families of ore deposits, magmatic, hydrothermal and mechanically concentrated (e.g. placers), a link with mantle plumes can be confidently assumed for most, but by no means all, deposits of the first family. The majority of hydrothermal ore deposits form at convergent, divergent and collisional plate boundaries, most notably porphyry and epithermal systems, volcanic-hosted massive sulphides and orogenic lodes, respectively. However, I suggest that a connection with mantle plume processes for those hydrothermal ore systems that form in plate interiors, can be reasonably assumed. These are included in this contribution.

In this context, it should be noted that, apart from *Pirajno* (2000) and *Schissel* and *Smail* (2001), surprisingly little has been published in the recent literature on the relationships between mantle plumes and ore deposits, since *Mitchell* and *Garson* (1981) and *Sawkins* (1990), who discussed mineral deposits formed in continental hotspots, rifts and aulacogens. *Barley* and *Groves* (1992) and *Barley* et al. (1997, 1998) proposed a link between mantle plumes, supercontinent aggregation and dispersal and the distribution of ore deposits through geological time.

Hotspots and mantle plumes

By and large, mantle convection is driven by three fundamental processes: heat loss from the core, internal heating from radioactive decay, and cooling from above (sinking of lithospheric slabs) (*Condie*, 2001). Current ideas on mantle plumes postulate that they are "jets", "narrow upwelling currents", or "narrow cylindrical conduits" of hot, low-density material originating either from the core-mantle boundary (CMB; one-layer mantle model), and/or from the 670 km discontinuity at the base of the upper mantle (two-layered mantle model) (*Davies*, 1999; *Schubert* et al., 2001). These "plumes", account for ~10% of the Earth's heat dissipation. The general consensus is that most large and long-lived plumes originate from the CMB, at the D″ thermal boundary layer, and are caused by heat from the outer core that is focused into a plume, driven upward in response to sinking of cool lithospheric slabs (Fig. 1). The structure of mantle plumes, modelled through numerous laboratory experiments (e.g. *Griffiths* and *Campbell*, 1990) and imaged by seismic tomography (*Woodhouse* and *Dziewonski*, 1989; *Forte* and *Mitrovica*, 2001), consists of a tail or stem and a mushroom-shaped head (Fig. 1). The plume head is cooler than the tail, because it contains entrained material from the surrounding cooler mantle. Recently, *Courtillot* et al. (2003) based on the features of 49 hotspots identified three types of plumes, namely: 1) primary or deep plumes, originating from the D″ layer; 2) secondary plumes, originating from the top of large domes of deep plumes or superplumes; 3) tertiary plumes or Andersonian plumes, originating from near the 670 km discontinuity and linked to tensile

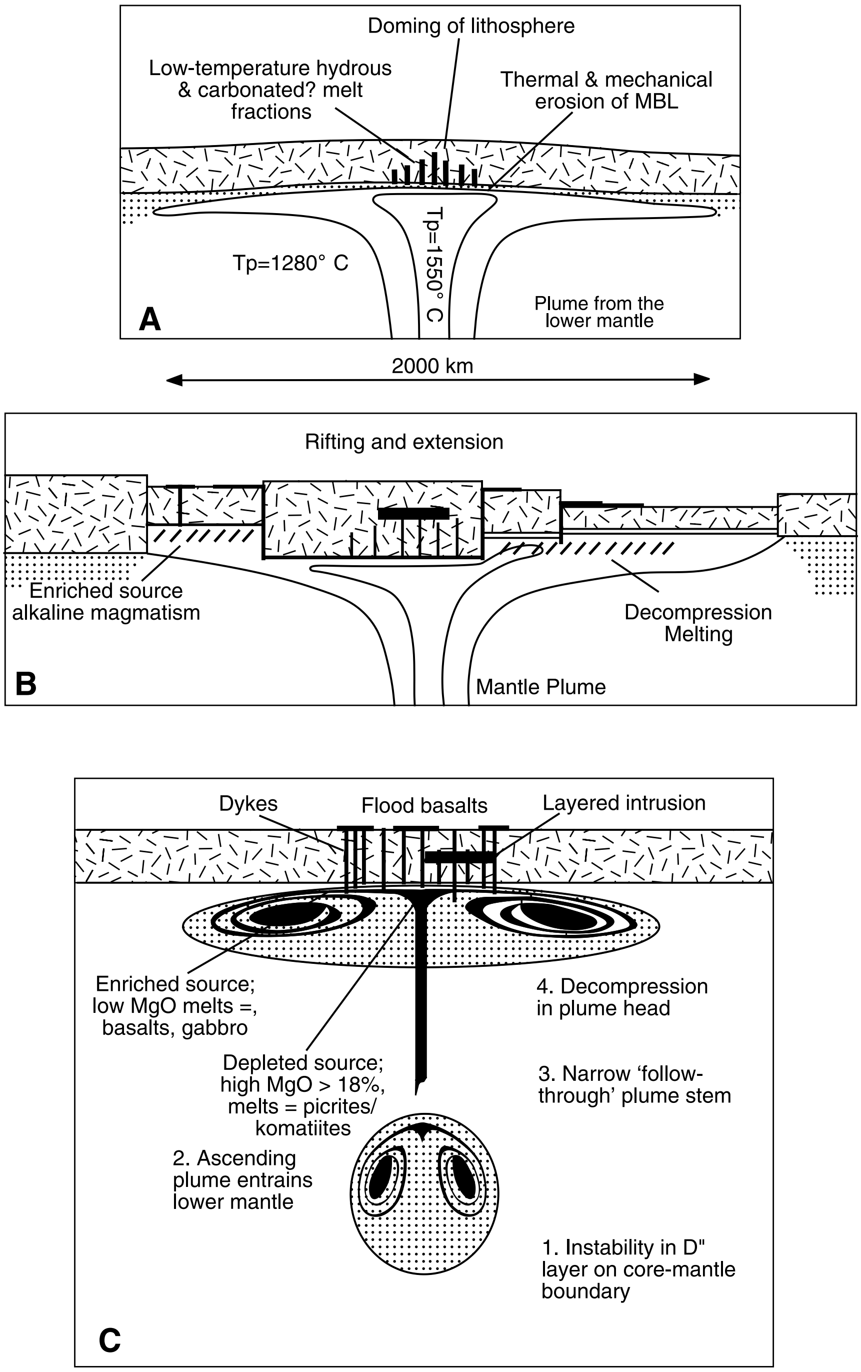


Fig. 1. Schematic illustration showing three types of mantle plume-lithosphere interactions. **A** A plume head impinges the lithosphere causing doming of the overlying crust; the plume has a potential temperature 270 °C greater than the surrounding mantle. **B** Lithospheric extension results in rifting along pre-existing zones of crustal weakness, decompression melting occurs in the plume head. **C** A mantle plume rises from the core-mantle boundary, ascends through the lower mantle to impact onto the base of the lithosphere, followed by decompression melting and emplacement of layered intrusions, dykes and continental flood basalts. **A** and **B** after and slightly modified from *Saunders* et al. (1992): **C** is after *Campbell* and *Griffiths* (1990)

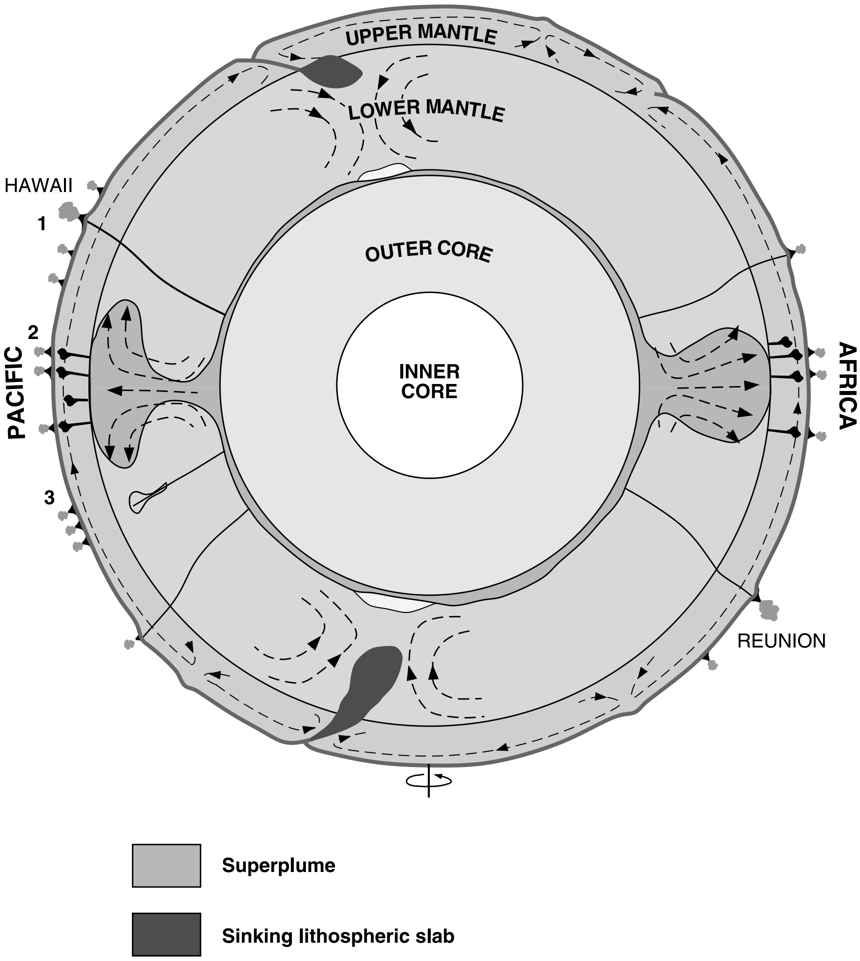


Fig. 2. Idealised cross-section of the Earth showing three types of mantle plumes: 1) primary (e.g. Reunion) rising from the core-mantle boundary, which include superplumes under Africa and the Pacific Ocean, with 2) secondary plumes emerging from them: 3) tertiary plumes rising from the upper mantle regions. More details in text. After *Courtillot* et al. (2003)

stresses in the lithosphere (e.g. *King* and *Anderson*, 1998). The superplumes of *Courtillot* et al. (2003) are located on antipodal regions, Africa and the central Pacific Ocean, where two massive mantle upwellings are evidenced by high crustal elevation (superswells) and by corresponding regions of low shear wave (Vs) velocity anomalies in the mantle (*Gurnis* et al., 2000) (Fig. 2). As mentioned above, from the top of these dome-like upwellings individual or secondary plumes rise from the base of the upper mantle to impinge onto the base of the lithosphere. The African Superplume possibly includes a number of secondary plumes that have acted on the African plate or parts of it for long periods of geological time (Fig. 3).

Partial melting in the plume head occurs by adiabatic decompression yielding lower temperature and lower-Mg melts (tholeiitic basalts), whereas melting in the high-temperature tail yields high Mg-melts (picrites, komatiites) (*Campbell*

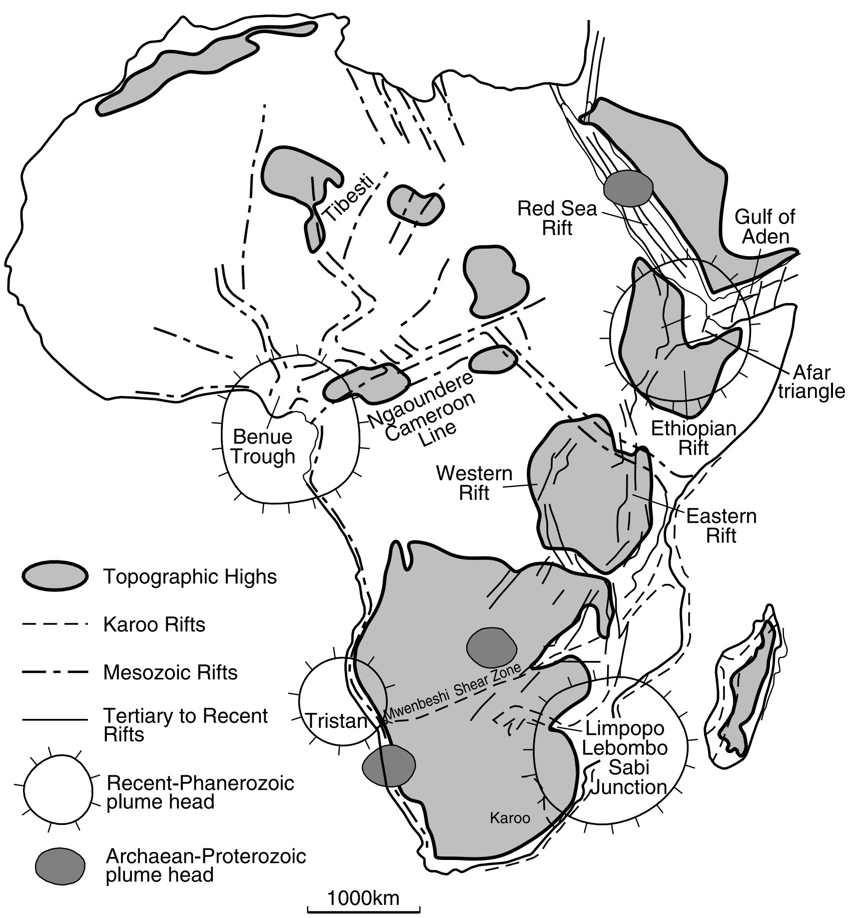


Fig. 3. Topographic highs, rift systems and assumed mantle plumes in the African plate. Details in text

et al., 1989) (Fig. 1). Degree of partial melting, volatile contents, nature of mantle material (enriched, depleted, entraining fragments of subducted slabs) control the products of plume magmatism. For this reason, plume-related rocks are isotopically heterogeneous and commonly characterized by nearly flat chondrite-normalised REE patterns. Ocean island basalts and continental flood basalts have somewhat monotonous major and trace element compositions, but with distinctive trace element patterns and ratios (e.g. Nb/Y, Zr/Y) as compared to mid-ocean ridge and volcanic arc basaltic rocks (*Condie*, 2003). Intraplate basalts are enriched in the most incompatible elements (Rb, Ba, Th, K, Ta, Nb), with positive Nb–Ta anomalies for ocean islands basalts, whereas most continental flood basalts have small negative Nb–Ta anomalies (*Condie*, 2001; *Reiners*, 2002). Good isotopic tracers of mantle plumes include $^{3}He/^{4}He$, $^{187}Os/^{188}Os$, $^{143}Nd/^{144}Nd$, Pb and Ne isotopes (e.g. *Marty*, 1997; *Samuel* and *Farnetani*, 2003).

The surface expression of mantle plumes is typically manifested by doming of the crust, reflected as topographic swells of 1000–2000 km in diameter and 2000–4000 m elevations (above sea level), and intraplate volcanism (*Gurnis* et al., 2000;

Şengör, 2001; *Ernst* and *Buchan*, 2003). Regions of intraplate anorogenic volcanism are commonly called "hotspots", a loose term that essentially refers to the concept of a stationary heat source in the mantle and the high heat flow that is related to magma advection (*Wilson*, 1963; *Schubert* et al., 2001). Geodetic data show that several hotspot regions correlate with rises in the gravitational equipotential surface (geoid high), probably reflecting the buoyancy of heated lithosphere (*Perfit* and *Davidson*, 2000).

Apart from the African and central Pacific Ocean crustal superswells, other regions of the Earth display uplifts that are considered to be related to the impingement of mantle plumes. One of these is the Mongolian plateau with an average elevation of 2000 m. This plateau is characterized by a basin-and-range topography, presence of major domes with altitudes in excess of 3000 m, anomalous low seismic velocities, high heat flow and is dissected by rifts of Cenozoic age that define a triple junction (*Windley* and *Allen*, 1993). The Mongolian plateau is also the focus of intraplate alkaline basaltic volcanism. All these features (uplift and doming, high heat flow, rifts, alkaline volcanism) are indicative of lithospheric thinning and extension above a hot asthenospheric mantle, and are consistent with the plateau representing the initial stages of mantle plume-lithosphere interaction (*Windley* and *Allen*, 1993).

Uplift is followed by subsidence due to loss of buoyancy of the plume head, or removal of magma from the top of the plume, thermal decay, or a combination of all three (*Condie*, 2001). Subsidence and crustal sagging cause the formation of sedimentary basins, characterized by the deposition of extensive aprons of siliciclastics, carbonates and evaporites, commonly overlain by continental flood basalts and/or transected by related dyke swarms. An example is provided by the Centralian Superbasin in Australia (*Walter* et al., 1995), where crustal sagging began as a result of a mantle plume activity at about 826 Ma (*Zhao* et al., 1994), with the deposition of thick successions of marine and fluvial sands. The evolution of this large depositional system continued through to the latest Proterozoic, culminating with the eruption of continental flood basalts from a second plume event at about 510 Ma (Antrim Plateau and Table Hill Volcanics) (*Walter* et al., 1995; *Hanley* and *Wingate*, 2000). Thus, uplifts related to mantle plumes also record pulses of sedimentary successions that are controlled by eustatic sea level fluctuations, due to an interplay of increased oceanic plateau formation (sea level rise) and supercontinent aggregation (sea level fall) (*White* and *Lovell*, 1997; *Williams* and *Gostin*, 2000; *Condie* et al., 2000).

In addition to normal plume events, there appear to be major pulses of heat transfer in the evolution of the Earth, in which a number of plumes impinge on to the base of the lithosphere. These plume events, also called superplumes (*Larson*, 1991; *Ernst* and *Buchan*, 2002), have important implications in terms of possible links with supercontinent cycles and time-space distribution of metalliferous deposits (*Barley* and *Groves*, 1992; *Abbott* and *Isley*, 2002a).

The eruption and intrusion of great volumes of mafic and ultramafic melts is attributed to the rise and impingement of mantle plumes on continental and oceanic lithospheric plates. These large-scale emplacements of mafic rocks are termed Large Igneous Provinces (LIPs; *Coffin* and *Eldhom*, 1992, 1994). The eruptions of mafic melts form vast fields of lava flows and associated igneous complexes

(LIPs), up to $7 \times 10^6\,km^2$ in areal extent (e.g. Central Atlantic province; *Wilson*, 1997; *Marzoli* et al., 1999; Siberian flood basalt province; *Nikishin* et al., 2002; *Kamo* et al., 2003), or lines of oceanic islands, like the Emperor-Hawaiian chain, thousands of kilometres long. In the ocean these vast lava fields are known as oceanic plateaux, such as the Ontong Java and Kerguelen plateaux (*Neal* et al., 1997; *Wallace* et al., 2002). Plumes may also interact with mid-ocean ridges forming large islands, such as Iceland. There is geophysical, geological and bathymetric evidence for the presence of voluminous mafic extrusive and intrusive complexes along the passive, trailing edges of continents that have been rifted and separated by sea floor spreading. These complexes exhibit prominent seismic seaward-dipping reflectors, or seaward-dipping layers at the flanks of, and parallel to the continental slope (*Meissner* and *Köpnick*, 1988). Typically, they are underlain by high-velocity bodies (7.2–7.5 $km\,s^{-1}$), like those observed beneath the oceanic plateaux. Reflectors on the southeastern and eastern Greenland coast, correlate with on-shore Tertiary flood basalts, together forming part of the North Atlantic Igneous Province (*Coffin* and *Eldhom*, 1994; *Klausen* and *Larsen*, 2002). The emplacement of LIPs has also been correlated with climate change and mass extinctions (e.g. *Olsen*, 1999; *Wignall*, 2001). This correlation is possibly due to massive emission of CO_2 from volcanic eruptions and the breakdown of methane clathrates (*Jahren*, 2002). The latest precise age determinations from the great Siberian flood basalts (also known as Siberian Traps or Tunguska volcanic province) show a remarkable coincidence with the Permian-Triassic mass extinction event (*Kamo* et al., 2003).

Where mantle plumes impinge onto subcontinental lithosphere, rifting may occur and may be accompanied by the eruption of continental flood basalts (CFB), exemplified by the well-studied Paraná-Etendeka, Karoo-Ferrar, Siberian and Deccan provinces. It is important to remember that the location of lava flows, sill complexes and layered intrusions are not necessarily indicative of a plume centre, because mafic melts can be transported for great distances from the plume centre or head. An example is provided by the Ferrar igneous province of Antarctica, which is cogenetic with the Karoo province of southern Africa. The emplacement of the Ferrar province was controlled by a huge rift system in the early Jurassic, in which magma dispersal, from a single batch, took place along 3000 km at mid-upper crustal levels (*Elliot* et al., 1999).

Ideally, a plume centre could be identified by the focal point of converging giant dyke swarms, such as the great Mackenzie swarm in Canada (*Ernst* and *Buchan*, 2001; *Ernst* et al., 1995), or converging rift systems (triple junction). The 1.26 Ga Mackenzie dyke swarm is the largest on Earth, and is the best illustration of the spatial and genetic link with layered intrusions and flood basalts. The Mackenzie swarm is 2400 km long and has a maximum width of 1800 km. Dykes have an average thickness of 30 m, but many are up to 150 m thick. The Mackenzie swarm fans out of a focal point in northwestern Canada, where it is suggested that a hotspot and underlying mantle plume existed. This is supported by the presence in the area of regional gravity highs, which may be caused by shallow mafic-ultramafic intrusions (*LeCheminant* and *Heaman*, 1989; *Baragar* et al., 1996). The swarm is spatially associated and coeval with the Muskox layered intrusion and the Coppermine River flood basalts.

An example of a triple junction is the Afar triangle in the horn of Africa (*Mohr*, 1978). The Afar is a depression at the triple junction defined by the Red Sea, the Gulf of Aden and the Ethiopian part of the 5000 km-long East African Rift System (EARS, *McConnell*, 1972; *Tesfaye* et al., 2003; Fig. 3). The Afar triangle marks the transition between the continental and oceanic rifts (Red Sea and the Gulf of Aden) of the EARS (Fig. 3). The Afar region is characterized by fissure-fed tholeiitic volcanism, hot springs and playa lakes with thick evaporite deposits (*Barberi* and *Varet*, 1978). The Afar, together with the Ethiopian and East African plateaux, is the general area, where a mantle plume is impinging beneath the continental lithosphere. This is the conclusion reached by *Ebinger* and *Sleep* (1998), who on the basis of gravity data and geochemistry of the volcanic products, numerically modelled the dynamics of the Afar mantle plume. *Rogers* et al. (2000), on the basis of Sr, Nd, and Pb isotope systematics and geochemical data, suggested that the EARS is underlain by two mantle plumes, one beneath the Kenya rift and the other beneath the Ethiopian rift and Afar region. An interesting concept proposed by *Ebinger* and *Sleep* (1998) is that the effects of the Afar-east African plume might extend along the EARS, to the south, offshore to the Comoros islands, and to the west along the Darfur swell and the Ngaoundere rift (Cameroon Line; Fig. 3). In their model, the authors suggested that the plume head flattens beneath the lithosphere, and flows laterally in zones where the lithosphere is thinner. In this way the behaviour of the plume, in terms of melting, is controlled by pre-existing variations in the thickness of the lithosphere (see Fig. 1), such as those of the west African Mesozoic rift zones and along the passive margins of Africa and Arabia.

The classic scheme of plume-generated three-arm rifting proposed by *Burke* and *Dewey* (1973) remains a valid concept. Indeed, many rift systems are associated with crustal swells, as exemplified by the North Sea Basin and the East African rifts. The timing of extension, igneous activity, uplift, and subsidence, determines the nature of the rifting process, which can be active or passive. Active rifting is believed to be a direct consequence of the impact of mantle plumes onto the lithosphere and is preceded by uplift of the crust and thinning of the lithosphere (e.g. East African rifts); the initial stages of active rifting are typically accompanied by alkaline bimodal (mafic-felsic, both extrusive and intrusive) magmatism. Passive rifting, on the other hand is linked to far-field lithospheric stresses, related to plate boundary forces and may be collision related (e.g. impactogens, such as the Rhine rift system and the Baikal rift; *Şengör* et al., 1978; *Molnar* and *Tapponier*, 1975; *Mats*, 1993). The concept of active and passive rifting is somewhat contentious and a case for interactions of rifting processes associated with both plate boundary forces and mantle plumes has been made for the Late Permian-Triassic rift basins, located between the East European and Siberian Cratons, with which the great Siberian traps (or Tunguska flood basalts) are related (*Nikishin* et al., 2002). Reviews of active and passive rifting processes are given in *Ruppel* (1995) and *Ziegler* and *Cloetingh* (2004), whereas *Şengör* and *Natal'in* (2001) provide a comprehensive listing of rifts world-wide.

Plumes may also interact with passive continental margins, forming thick piles of mafic rocks, as in the North Atlantic volcanic margin, which links the southeast coast of Greenland with Iceland (*Eldhom* and *Grue*, 1994), or they may impinge onto active continental margins or island arcs, resulting in the inception of back-arc basins, as may be the case for some of the back arc systems in SE Asia (*Hall*, 1996). Mantle plume-island arc interactions have been postulated for the Archaean Abitibi-Wawa greenstone belts in the Superior Province (*Wyman* et al., 2002). In addition, the lithological associations of greenstone belts of the Superior province suggest that they represent accreted tectonic domains that include plume-generated oceanic plateaux (*Polat* et al., 1998). These tectonic interpretations that involve mantle plumes have important implications for orogenic precious metal mineralising systems (*Goldfarb* et al., 2001; *Groves* et al., 1998).

A summary and examples of the tectonic, magmatic and geomorphological features, caused by mantle plumes is presented in Table 1. The recognition of mantle plume processes in the geological record is comprehensively treated in *Ernst* and *Buchan* (2003).

The evidence from Mars, Venus and the moon

All terrestrial planets, show volcanic landforms, highlands or continents, plains, impact cratered regions, rifts and other linear features. Mars and Venus have large crustal uplifts on their surfaces that morphologically resemble those that exist today on the African plate. These crustal uplifts are interpreted as hotspot swells, owing to their association with large gravity anomalies and intraplate-like volcanism (*Head* and *Coffin*, 1997). This evidence indicates that mantle plume processes most probably existed on Mars and Venus, both considered as "one-tectonic plate" planets. The Tharsis region on Mars is approximately 4000 km wide and 10 km high, covering an area $>6.5\times10^6\,km^2$ and surmounted by gigantic volcanic constructs (e.g. Olympus Mons, the largest volcano in the solar system, rising some 25 km above the surrounding terrain; *Zimbelman*, 2000). On Earth, the African plate may be considered as the closest analogue of a "one plate" planet, because Africa has been stationary for at least 65 Ma with respect to underlying plumes (*Burke*, 1996), so that the crustal uplifts, rifting and intraplate volcanism have not been recycled by convergent margin tectonics. Therefore, it could be that the African superplume of *Courtillot* et al. (2003) is the terrestrial analogue of the Tharsis plume on Mars.

On Venus there are at least 34 major uplifts, or crustal plateaux, with widths of between 1500 and 3000 km and heights of about 2 km (*Roberts* and *Head*, 1993; *Schubert* et al., 2001). There are nine volcanic rises associated with positive gravity anomalies, ranging in diameter from 1300 to 2300 km and up to 2.5 km high, interpreted to represent the surface signature of deep mantle plumes (*Hansen*, 2003). *Crumpler* and *Aubele* (2000) list 1738 volcanic centers on Venus larger than 20 km. Endogenic circular features on Venus, known as coronae, with diameters ranging from 60 to 2000 km are associated with volcanic rises, rift zones and dyke swarms and are also attributed to mantle plumes (*Roberts* and *Head*, 1993). A genetic model proposed by *Hansen* (2003) suggest that coronae form from clusters of small diapirs that rise from a larger impacting mantle plume. This model is similar to *Courtillot* et al.'s (2003) superplume, referred to above.

Our moon has no large volcanoes; instead it has vast fields of basaltic lavas, called "maria" by the early astronomers (maria means seas in Latin), which were

Table 1. *Geological processes associated with mantle plumes*

Tectonic, magmatic and geomorphological effects	Selected examples
Doming of the crust; uplift and subsidence: The impact of a mantle plume onto the base of the lithosphere results in doming of the crust. Subsidence follows due to thermal decay, forming sedimentary basins.	Recent to modern examples of mantle-plume induced topographic swells or uplifts, include the East African plateau (Afar hotspot), the Mongolian plateau and the swells in the central Pacific Ocean. The great escarpments of southern Africa, a major physiographic feature in the subcontinent, were formed in response to uplift linked to the Paraná-Etendeka and the Karoo mantle plumes. The Centralian superbasin in Australia was probably formed as a large sag resulting from crustal thinning over a mantle plume.
Intracontinental rifting: Doming can result in rupture and rifting, with development of triple junctions (*Burke* and *Dewey*, 1973).	Examples of continental rifts, all associated with intraplate igneous activity, include the East African Rift System (Afar hotspot); the Permo-Triassic rift system developed between the East European and Siberian Cratons (*Nikishin* et al., 2002); the 1.1 Ga Mid-continent rift system in the USA (*Ojakangas* et al., 2001); the 1.0–0.7 Ga Damara-Irumide rift systems that developed between the Congo and Kalahari Cratons in southern Africa.
Continental breakups and related passive margins: Mantle plumes cause stress during uplift as well as thermal thinning due to conductive heat transfer (*Wilson*, 1993). Extension and rifting may lead to the formation of triple junctions, continental break up and eventually sea floor spreading (*Abbott* and *Isley*, 2002a). Passive margins and associated volcano-sedimentary prisms evolve from continental rifts and the opening of oceans	Best examples of continental breakup are provided by the Atlantic Ocean and North Sea: the South Atlantic (Tristan da Cunha plume; 137–127 Ma Paraná-Etendeka igneous province (*Peate*, 1997), the North Atlantic (Iceland plume and the 60–54 Ma North Atlantic Igneous Province; *Nielsen* et al., 2002), the Central Atlantic (Cape Verde? plume and 200 Ma Central Atlantic Magmatic Province; *Marzoli* et al., 1999). An example of plume-related passive margins is the southwest African (Namibian) margin (Tristan da Cunha hotspot).
Mantle plume magmatism (see *Ernst* and *Buchan*, 2001, for a full list of mantle plume magmatism since 3.5 Ga): • Layered intrusions • Dyke swarms • Oceanic plateaux	Examples of: *Layered intrusions*: Windimurra (2.8 Ga, West. Australia), Stillwater (2.7 Ga, USA), Bushveld Complex (2.06 Ga, South Africa), Giles Complex (1.1 Ga, central Australia), Duluth (1.1 Ga, USA),

(*continued*)

Table 1 (*continued*)

Tectonic, magmatic and geomorphological effects	Selected examples
• Seaward-dipping reflectors • Sill complexes • Continental flood basalts • Anorogenic alkaline complexes • Carbonatites, kimberlites, lamproites	Skaergaard (0.06 Ga, Greenland). *Dyke swarms*: Widgiemooltha (2.4 Ga, West. Australia), Matachewan (2.4–2.5 Ga, Canada), Mackenzie (1.26 Ga, Canada), Gairdner (0.83 Ga, central Australia), Okavango (0.18 Ga, Botswana). *Oceanic plateaux*: Kerguelen, Ontong-Java, Caribeean-Colombia. *Seaward dipping reflectors*: numerous examples in the Atlantic Ocean (west coast of Africa, north American eastern seaboard, the North Sea). *Sill complexes*: Noril'sk-Talnakh; central Western Australia. *Continental flood basalts*: Ventersdorp, Fortescue, Karoo, Siberian Traps, Emeishan, Paraná-Etendeka, Deccan, Ethiopian-Yemen traps, Columbia River. *Anorogenic alkaline complexes*: Damaran province (Namibia); Niger-Nigeria province; many examples in Africa (*Woolley,* 2001); Gardar province (Greenland). *Carbonatites, kimberlites, lamproite*: numerous fields in southern Africa, Canada, northwest Australia, India, Arabian shield, Brazil.
Drainage patterns: Doming of the continental crust results in radial drainage systems along the flanks of the uplifts and in deep incisions with developments of canyons.	Examples include the Orange-Limpopo rivers drainage systems related to the Karoo and Etendeka escarpments in southern Africa; in southern Namibia, the Fish River canyon is due to uplift relating to the Tristan da Cunha plume and opening of the South Atlantic. A radial drainage pattern characterises the Mongolian plateau. In Australia, palaeocanyons, up to 250 km long and 1 km deep, were incised by regional uplift ascribed to a mantle plume (*Williams* and *Gostin*, 2000).

formed between 3 and 4 Ga. The lunar maria, the dark smooth regions visible from Earth, are associated with lava channels (rilles), flow fronts, vents and other volcanic features and are analogous to terrestrial flood basalts. The maria lava fields tend to fill impact-created basins, where again there is an association with large positive gravity anomalies, which are interpreted as a concentration of plugs of dense mantle material. Hence, a possible connection may exist between meteorite impacts and the outpouring of the lavas; a topic that I briefly discuss in the section that follows.

A possible link with meteorite impacts

Some authors have proposed that meteorite impacts can induce mantle plumes, decompression melting and the inception of large igneous provinces (e.g. *Boslough* et al., 1996; *Jones* et al., 2002). There is circumstantial evidence, within the limits of isotopic age dating errors, of a possible link between biological mass extinctions, large meteorite impacts and flood volcanism (*Negi* et al., 1993; *Becker*, 2002). A correlation between major bolide impacts and mantle plume breakouts is advocated by *Abbott* and *Isley* (2002b). These authors have examined the timing of known major impacts for the last 3800 Ma of Earth's history and found that a correlation exists with hotspot volcanism, suggesting that the upwelling of mantle diapirs may have something to do with large meteorite impacts. Support for this hypothesis comes from computer simulations by *Jones* et al. (2002), in which they show that a large impact can cause decompression melting and the rise of a mantle plume. These views, however, are controversial because of the poor age constraints on actual dated impacts and the error limits of isotopic ages of mantle plume proxies, such as large igneous provinces and mafic-ultramafic layered intrusions. Nevertheless, the apparent coincidence of impacts, mass extinctions and volcanism, although poorly-defined statistically and chronologically, is striking.

Ore deposits and mantle plumes

As mentioned above, the rising of mantle plumes results in tectonic and magmatic manifestations that include crustal uplifts, continental rifting, the emplacement of anorogenic magmas, and basalt-dominated large igneous provinces (LIPs) on continents and in the oceans. A great variety of ore deposits, magmatic and hydrothermal, form in these tectono-magmatic settings. We can consider these ore deposits in terms of two end-members: 1) magma-associated ore deposits and; 2) giant hydrothermal systems powered by the thermal energy related to the cooling of anorogenic magmatic systems in the crust.

Magma-associated ore deposits

An overview of magma-associated ore deposits that can be linked to mantle plume activity is shown in Table 2. The closest link with mantle plumes is probably represented by the large layered intrusions and associated LIPs. Erosion of continental flood basalts exposes feeder dykes, sill complexes and magma chambers (layered intrusions). Mafic dyke swarms, sill complexes and layered intrusions therefore represent the remnants of large igneous provinces and can, in the right conditions, host magmatic ore deposits.

Anorogenic magmatism is associated, in space and time, with extensional tectonics, hotspots and intracontinental rifts. Intracontinental alkaline magmatism usually occurs during the early phases of rifting events, in which direct links are

Table 2. *Main geological features of mantle plume-related magma-associated ore deposits*

Ore deposit	Main geological features
Layered mafic-ultramafic complexes Cr, PGE, V, Fe, Ti, (Cu, Ni) (e.g. Bushveld, South Africa; Great Dyke, Zimbabwe; Stillwater, Duluth, USA)	• The main feature is the large size of the bodies, and the nature of layering coupled with remarkable lateral extent. Represent multiple influxes of melts from the mantle. The mafic-ultramafic rocks host variable mineralisation (oxides and sulphides) and generally have high initial Sr ratios (crustal contamination). • Chromite occurs in the Great Dyke in seams and disseminations in olivine-bearing rock. In the Bushveld Complex Cr is concentrated in lower ultramafic parts of the complex (lower Critical Zone). Chromite seams are marked by great lateral extent. • Platinum Group Metals occur in the Merensky Reef and similar layers ± 2000 m above the base of the Bushveld Complex; in similar bodies and levels in the Stillwater Complex, and the Great Dyke. • Titaniferous magnetite $\pm$ vanadium ores in the Bushveld Complex are within the upper zone, and the top of the main zone gabbro-norites. Upper zone has cumulate magnetite disseminated in gabbros and as thick magnetite layers.
Flood basalts and sill complexes: Ni–Cu (PGE) (e.g. Norilsk-Talnakh, Siberia); Ni–Cu in Archaean komatiite lava channels (e.g. Kambalda, WA)	• Related to mafic-ultramafic feeders/conduits located along zones of crustal weakness (i.e. deep seated rift-related faults). • Disseminated to massive Ni–Cu sulphides along the basal contact of sill-like intrusions, or lava channels. • Sulphides may also occur as disseminations and veins in footwall rocks. • Evaporites may be present (e.g. Noril'sk), and are considered important as a source of S. Their presence is also directly related to the sedimentary facies developed on rift margins and aulacogens. S-rich sedimentary rocks form substrate of lava channels.
Anorogenic granites Sn, Nb, Ta, W, U, Th, F, Be, Zn, Cu (e.g. Nigeria, Niger, West Africa; Bushveld granites, South Africa; Rapakivi granites, Scandinavia)	• Granites emplaced in stable intracratonic environments, commonly as ring complexes, and may form extensive "linear" arrays. • Major-rock types are paralkaline albite–riebeckite granites. Sn is associated with less alkaline biotite granites. • In Nigeria, Sn, Ta occurs as disseminations, within greisen zones and quartz veins with base metal sulphides. Commonly concentrated along horizontal roof sections of biotite granite.

(*continued*)

Table 2 (*continued*)

Ore deposit	Main geological features
	• In the Bushveld Complex, Sn, F mineralisation is found in pipes, sheet-like disseminations in coarse-grained porphyry granites, or as fissure veins, breccia zones or replacement bodies in the granites or their volcanic and sedimentary roof zones. Associated minerals include tourmaline, sericite, quartz, and chlorite. • Rapakivi granites in Scandinavia. Enrichment in Sn and/or Be, W, Zn, Cu. Mineralisation is related to the youngest phases, and is especially enriched in the apical parts or at contacts. Mineralisation occurs as disseminations, in pegmatite veins, greisens, and quartz veins and contact skarns.
Anorogenic granitoids and Olympic Dam-style ore deposits; Feoxide–Cu–Au–REE (e.g. Cloncurry district in Queensland; Sweeden Fe districts)	• A class that includes a wide range of ore deposit styles, such as Olympic Dam, Ernest Henry, Bayan Obo, Kiruna. Vergenoeg, ?Palabora; ores are associated with breccia pipes or bodies; alkali metasomatism and hematitic alteration at the regional scale. Granitoids have high K, REE, Zr, Y, F, U, Th.
Massif-type anorthosite: Fe–Ti–V; troctolite-anorthosite Cu–Ni–Co (e.g. Labrador-Grenville Provinces, Canada/USA; Voisey's Bay; ?O'Kiep?, South Africa)	• Massif type intrusives, limited time control (1.5–1.3 Ga). Mostly occur in two linear belts (one in the Southern Hemisphere and the other extending across North America through Scandinavia to Russia). • Commonly intruded into high grade regionally metamorphosed rocks. Bodies are sheet-like and lack layering, with distinctive homogeneous textures. Usually plagioclase-rich (90% plag); deep level emplacement, whereas associated shallow level gabbroic, norite, and troctolitic anorthosites have 78–90% plagioclase. • Cu–Ni–Co in troctolite-anorthosite complexes. • Disseminated Fe and Fe–Ti oxide ores, with occasional lenses or irregular bodies of massive ilmenite. Ti–magnetite commonly enriched in vanadium in gabbroic anorthosite, or noritic gabbro, ilmenite–hematite in cores of anorthosite bodies.
Sudbury Igneous Complex Cu–Ni	• Layered mafic-ultramafic complex containing Cu–Ni sulphides in noritic intrusions in basal zones of complex and dyke-breccia bodies. Good evidence for relationship to meteorite impact. Original crater dimension in the order of 250 km; 1850 Ma. Raises the question of whether or not some mantle upwellings may be caused by giant meteorite impacts.

assumed with melt generation by mantle plume upwellings beneath thinned lithosphere (*Latin* et al., 1993). Anorogenic alkaline complexes and carbonatites are present in many LIPs (e.g. Deccan, Siberian traps, Paraná-Etendeka, Keeweenawan). The Late Jurassic-Early Cretaceous Damaraland alkaline igneous province in Namibia, is linked to the Tristan da Cunha plume in the South Atlantic, from which the Paraná-Etendeka flood basalts were generated (*Pirajno*, 1994; *Pirajno* et al., 2000; *Trumbull* et al., 2000). These alkaline rocks of the province may be the result of lithospheric melting in response to the thermal perturbation caused by this plume (*Ewart* et al., 1998).

Mafic-ultramafic systems. Magmatic sulphide ore deposits are typically hosted by mafic-ultramafic layered intrusions, dykes, sill complexes, continental flood basalts, and Archaean komatiite fields. *Lambert* et al. (1998), in discussing Re–Os isotope systematics, proposed a three-fold classification of magmatic Ni, Cu, PGE and Cr ore deposits: 1) Cu–Ni–PGE-rich sulphides, chromite and Fe–Ti–V oxides, hosted in mafic-ultramafic layered intrusions (e.g. Bushveld Complex, Great Dyke in southern Africa); 2) Cu–Ni sulphides associated with basaltic and gabbroic rocks (e.g. Duluth in the USA, Noril'sk-Talnakh in Russia, Jinchuan in China); 3) Archaean komatiite Ni sulphides (e.g. Kambalda in Western Australia). In terms of metal production, magmatic sulphide deposits can be divided into (*Li* et al., 2001): 1) Ni–Cu deposits with PGE as by-products; 2) PGE deposits with Ni–Cu as by products (Fig. 4). Examples of the former are the Kambalda deposit in

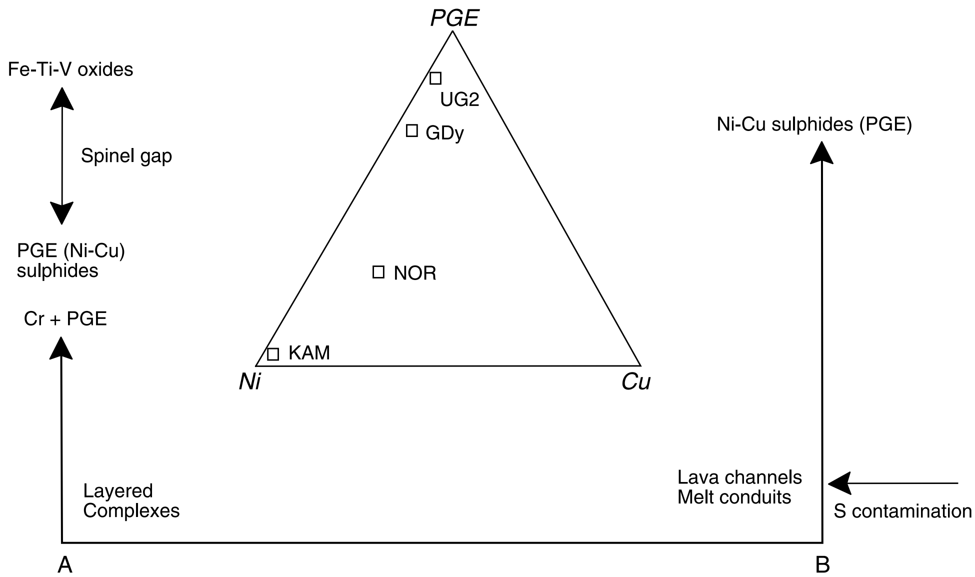


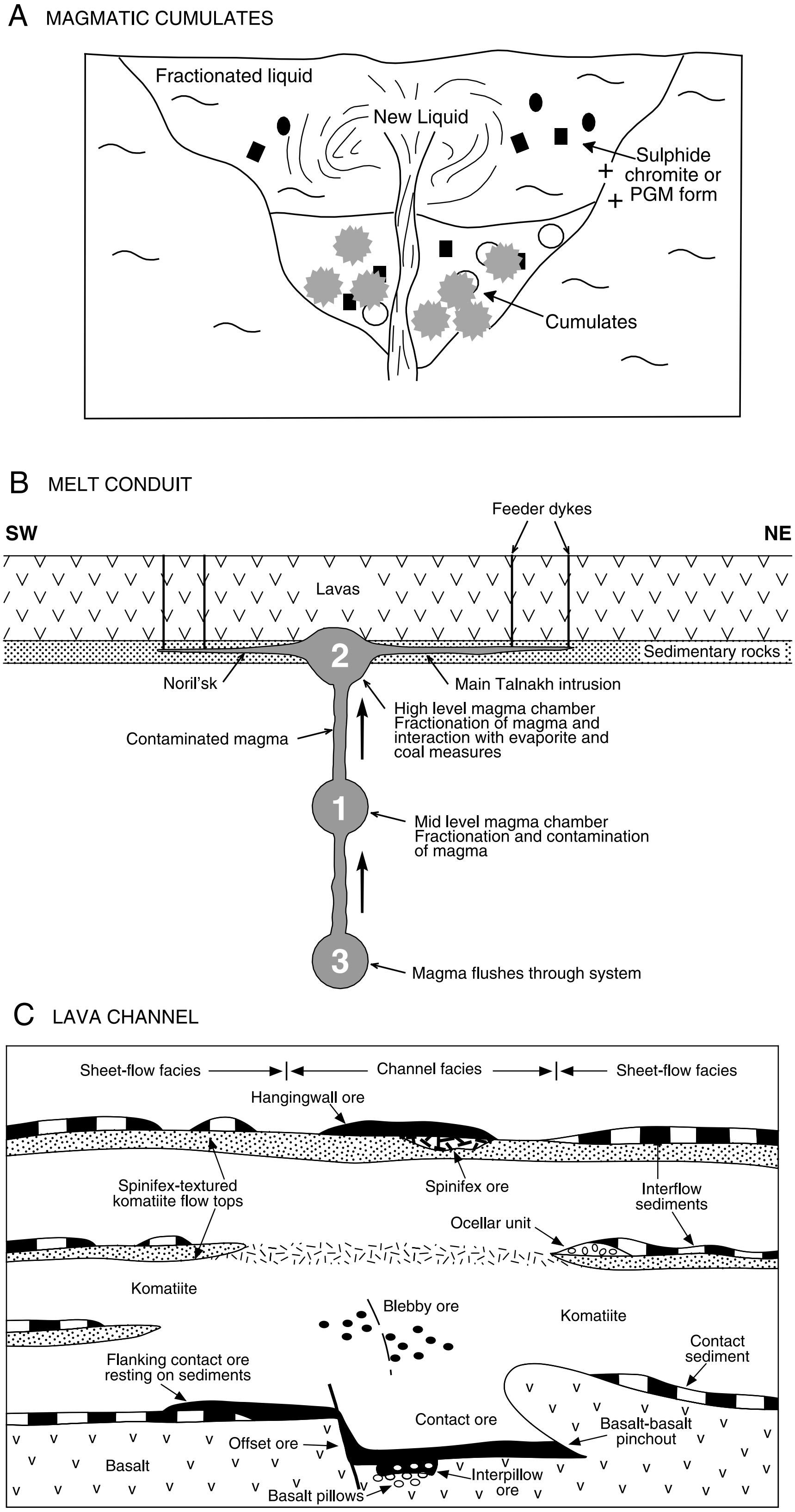
Fig. 4. Schematic illustration showing the two main paths of Ni–Cu and PGE magmatic mineralisation in mafic-ultramafic magmas. Path A relates to layered complexes, with PGE (Ni–Cu), Cr + PGE and Fe–Ti–V oxides being segregated directly from the melts. Path B is where crustal contamination and S addition play a critical role in order to cause sulphides supersaturation. See text for details and references. The triangular diagram qualitatively displays the position of some of the world-class magmatic deposits: KAM is Kambalda, NOR Noril'sk, GDy Great Dyke, UG2 is the platiniferous chromitite in the Bushveld Complex

Western Australia, Jinchuan in China, Noril'sk in Russia, Voisey's Bay in Canada. Important deposits of the second type are the UG-2 chromitite and Merensky Reef of the Bushveld Complex (South Africa), the Great Dyke (Zimbabwe) and the J-M Reef of the Stillwater Complex (USA). *Abbott* and *Isley* (2002a) suggested that layered intrusions formed by deep-mantle plume events contain high level of PGE and Cr, because these plumes are associated with high degrees of melting in the mantle and therefore higher concentrations of compatible elements such as Pt and Cr. It has also been suggested that "second stage" melts of mantle material can be PGE-enriched as a consequence of prior melt extraction from a S-saturated mantle source (*Hamlyn* and *Keays*, 1986).

Magmatic segregations of Cu–Ni sulphides and PGE are due to liquid immiscibility between a silicate magma and a sulphide liquid, probably facilitated by introduction of crustal sulphur into a silicate melt that was originally sulphur-undersaturated (*Keays*, 1995). The immiscible sulphide liquid efficiently scavenges metals such as PGE, Ni and Cu. Iron–Ti–V oxides tend to appear in the same layered intrusions, but at different statigraphic levels in layers of different composition. Thus, a common theme in layered intrusions, is that Cr and PGE $\pm$ Ni–Cu ores are associated with magmas having high Mg/Fe, poor Ca and alkali contents; whereas Fe–Ti–V oxides are associated with magmas that are richer in Fe and alkali + Ca. Consequently, sulphide and Cr mineralisation is commonly confined to the ultramafic layers at the base of the layered sequence, whereas Fe–Ti–V mineralisation is localised within mafic rocks at the top of the sequence.

Critical to these different deposits (Ni–Cu $\pm$ PGE and PGE $\pm$ Ni–Cu) is the style of sulphide accumulation (Fig. 5). In large layered intrusions (PGE $\pm$ Ni–Cu; e.g. Bushveld, Great Dyke) sulphide droplets are segregated from the melt and settle on the floor of the magma chamber (Fig. 5A). It is worth mentioning that a number authors (e.g. *McCandless* and *Ruiz*, 1991; *Boudreau* and *McCallum*, 1992) advocate the role of high-temperature magmatic-hydrothermal fluids in the genesis of PGE mineralisation in layered intrusions. The case for the hydrothermal transport of PGE, at least locally, is supported by numerous experimental studies (e.g. *Mountain* and *Wood*, 1988). In the case of Ni–Cu $\pm$ PGE (e.g. Noril'sk, Voisey's Bay) sulphide deposition is controlled by fluid dynamics in a magma conduit system (*Naldrett*, 1997; *Li* et al., 2001) (Fig. 5B). Addition of sulphur to a metal-rich melt is a necessary prerequisite in order to induce S saturation and the subsequent segregation of sulphides. A three-stage process can be envisaged (e.g. *Sproule* et al., 1999): 1) sulphur-undersaturated melt rises along a conduit till it crosses S-rich crustal rocks; 2) crustal assimilation causes S-saturation and segregation of immiscible sulphides; 3) the sulphide-rich melts are mechanically transported along conduits towards the surface (Fig. 5B), where they may passively accumulate, because of loss of velocity, at bends or structural traps, while the residual metal-depleted melt continues its upward ascent to erupt at the surface or

Fig. 5. **A** Magmatic sulphide segregations in a layered intrusion: after *Barnes* and *Maier* (1999). **B** Model of ore genesis for a melt conduit system, such as that of Noril'sk: after *Naldrett* (1997, 1999). **C** Schematic section (not to scale) showing a lava channel model for the genesis of komatiite-hosted Ni–Cu sulphide ores: after *Hill* et al. (1990)


A MAGMATIC CUMULATES
Fractionated liquid
New Liquid
Sulphide
chromite or
PGM form
Cumulates
B MELT CONDUIT
Feeder dykes
SW
NE
Lavas
Sedimentary rocks
2
Noril'sk
Main Talnakh intrusion
High level magma chamber
Fractionation of magma and
interaction with evaporite and
coal measures
Contaminated magma
1
Mid level magma chamber
Fractionation and contamination
of magma
3
Magma flushes through system
C LAVA CHANNEL
Sheet-flow facies
Channel facies
Sheet-flow facies
Hangingwall ore
Spinifex-textured
komatiite flow tops
Spinifex ore
Interflow
sediments
Ocellar unit
Komatiite
Blebby ore
Komatiite
Contact
sediment
Flanking contact ore
resting on sediments
Contact ore
Basalt-basalt
pinchout
Offset ore
Basalt
Interpillow
ore
Basalt pillows

to be emplaced as metal-depleted mafic sills (*Naldrett*, 1989). In the case of the Noril'sk deposits, the flow of large volumes of magma through overlying S-rich crustal or sedimentary rocks, such as evaporites and coal measures, would have caused the ingestion of S-rich rocks by the melts, thereby promoting sulphide saturation and precipitation (*Naldrett*, 1997, 1999). The hypothesis of an external S introduction into the melt, however, is being re-assessed because, at Noril'sk for example, geochemical signatures and S isotope systematics do not support derivation of S from evaporites, whereas the role of contamination by high-Si and low-S crustal materials at deep levels may have not been adequately considered (*Lightfoot*, 2003; *Ripley* et al., 2003). *Ripley* et al. (2002) suggested the possibility of fractional melting and total fusion of pelitic and graphitic gneiss rocks as a viable mechanism for the addition of S and C to the magma to cause the precipitation of magmatic sulphides.

Archaean komatiites. Komatiite, a term first introduced by *Viljoen* and *Viljoen* (1969), is defined as an ultramafic volcanic rock with >18 wt% MgO, on an anhydrous basis with $TiO_2 < 1\%$ and $Na_2O + K_2O < 1\%$ (*Le Maitre*, 2002). Komatiite lavas are important hosts of Fe–Ni–Cu sulphide mineralisation in Archaean greenstone belts in Australia, Canada, Zimbabwe and Brasil. Komatiite-hosted Ni–Cu (±PGE) deposits have been divided into two types (*Lesher* and *Keays*, 2002; *Beresford*, 2003). Type 1 are stratiform accumulation of Fe–Ni–Cu sulphides in embayments or troughs at the base of komatiite lava flows (e.g. Kambalda, Western Australia) (Fig. 5C). *Beresford* (2003) emphasised that the structural locales that host the massive sulphides are not thermally eroded lava channels. Type 2 are stratabound disseminated sulphides hosted in thick dunitic cumulate rocks (e.g. Mt. Keith, Western Australia). Details of the komatiite-hosted Ni–Cu (±PGE) mineralisation can be found in *Stone* et al. (2003), *Lesher* and *Keays* (2002).

It has been suggested that the Archaean greenstone komatiite–tholeiite sequences are products of mantle plumes (e.g. *Tomlinson* et al. 1998), and furthermore that these sequences may be the fragments of plume-related oceanic plateaux, by analogy with a rare Phanerozoic komatiite on Gorgona Island (*Storey* et al., 1991). The Archaean Abitibi-Wawa granite-greenstone belt (Superior Province, Canada) is a complex collage of accreted oceanic plateaux and island arcs that interacted with a flat-subduction event and the involvement of a mantle plume, which coupled with the entire subduction-accretion system (*Wyman* et al., 2002; *Polat* et al., 1998). The impingement of a mantle plume beneath an accretion-subduction system in the Archaean is reminiscent of the Great Basin geodynamics (Western USA), where the tectono-magmatic history of the region involved shallow subduction and underplating by a mantle plume (Yellowstone hotspot), with uplift (the Basin-and-Range plateau has an average elevation of 1500 m a.s.l.), rifting and anorogenic magmatism (*Ilchik* and *Barton*, 1997; *Oppliger* et al., 1997). This complex 55 Ma-ongoing geodynamic history is reflected in the metallogeny of the Great Basin, in which two temporal stages of epithermal systems have been recognized, one related to subduction, and the other to a mantle plume event (*John*, 2001).

Campbell and *Hill* (1988) and *Hill* et al. (1991) proposed that the Late Archaean greenstone sequences of the Eastern Goldfields of the Yilgarn Craton

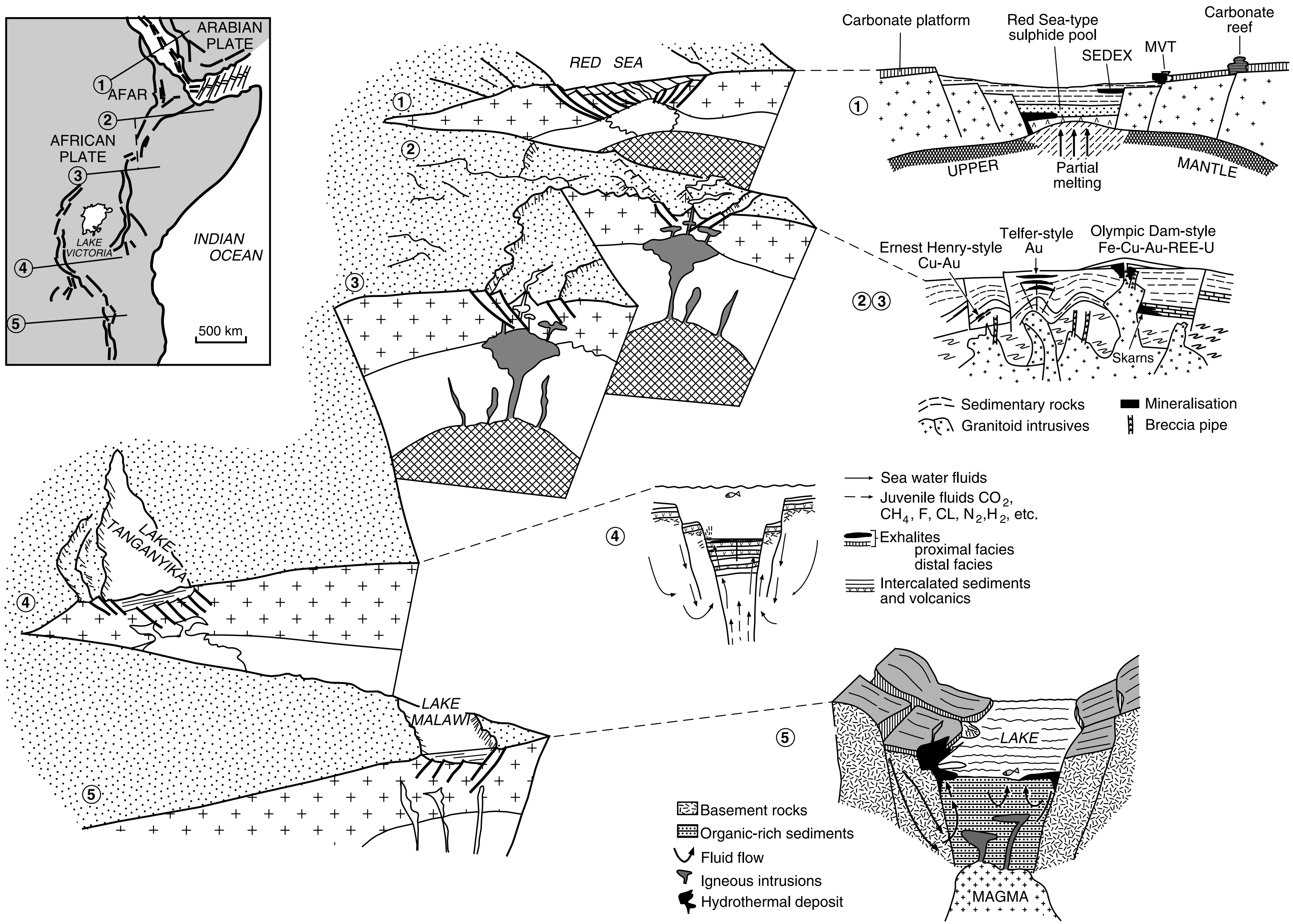
(Western Australia), which contain abundant komatiite rocks, are the result of major thermal disturbances in the mantle. Seen in this light, the komatiites of the Eastern Goldfields (Yilgarn Craton, Western Australia) can be considered as an Archaean flood basalt province (*Hill* et al., 1991). However, there is no consensus regarding the origin of komatiites. In contrast to the mantle plume hypothesis other workers suggested that the Archaean greenstone belts and komatiites were formed in subduction-related oceanic island arcs that were later accreted by plate convergence (e.g. *Myers*, 1995; *de Wit*, 1998). Thus, the origin of komatiites remains a debated topic. It is beyond the scope of this contribution to discuss details of komatiite magmatism, about which there is abundant literature. For details of komatiite magmatism the reader is referred to *Arndt* and *Nisbet* (1982), *Arndt* (1994) and *Barnes* et al. (1999).

Ore deposits associated with anorogenic magmatism. The nature of anorogenic magmatism is complex and varied, but for the purpose of this contribution, we consider three groups: 1) intracontinental alkaline complexes, 2) anorthosite-gabbro-troctolite (or massif-type) complexes, 3) kimberlites and carbonatites.

Anorogenic granitoids host or are associated with a great variety of ore deposits, which include both magmatic and hydrothermal types (Table 2). Anorogenic magmas are enriched in incompatible elements (e.g. Ti, P, Y, Nb, K, Th, U, F, Ba, REE) and produce peraluminous and peralkaline granitoids, which commonly contain greisen or late-magmatic sub-solidus Sn, W, Zn, Cu, U, Nb mineralisation (*Pirajno*, 1992, 1994). Hydrothermal ore deposits include the economically important Fe oxide–Cu–Au–REE class (*Porter*, 2000 and papers therein) (Table 2). The Fe oxide–Cu–Au–REE class (*Hitzman* et al., 1992; *Oreskes* and *Hitzman*, 1993), includes giant ore deposits such as Olympic Dam, Ernest Henry in Australia, La Candelaria in Chile, the Carajas deposits in Brazil, the Fe–Cu deposits at Boss-Bixby and Pea Ridge (USA) and perhaps the world-class Bayan Obo REE deposit of Inner Mongolia (*Chao* et al., 1997). The common theme in all these deposits is the enrichment in Fe, P, F and the widespread alkali metasomatism in the host rocks (*Hitzman* et al., 1992; *Oreskes* and *Hitzman*, 1993). Other deposits have been recognised in intracontinental settings that share some important features of the Fe oxide Cu–Au–REE class. These include the Vergenoeg magnetite–fluorite (*Borrok* et al., 1998), and the Palabora (a carbonatite) copper (*Groves* and *Vielreicher*, 2001) deposits in South Africa.

Typically, Fe oxide–Cu–Au–REE hydrothermal systems form in shallow crustal environments (4–6 km) and are the expression of volatile-rich, alkaline magmas (*Hitzman* et al., 1992). Their global occurrence covers the time span of approximately 1.8 to 1.1 Ga and they appear to be linked to planetary-scale rifting events and the assembly and breakup of supercontinents, like Rodinia (*Unrug*, 1997). An idealised setting for the Fe oxide Cu–Au–REE deposits is illustrated in Fig. 6.

Massif-type gabbro-anorthosite-troctolite intrusive complexes which appear to be related to rift settings and/or terrane boundaries, mainly host magmatic ore deposits. Massif-type anorthosite magmatism is widespread in the 1.5–1.3 Ga time span, forming a belt 5000 km long and 1000 km wide that extends across the Laurentian shield, from present-day California through to Scandinavia (*Windley*, 1995). The origin of these anorthosite intrusions, many of which also contain


ARABIAN PLATE
AFAR
AFRICAN PLATE
LAKE VICTORIA
INDIAN OCEAN
500 km
RED SEA
LAKE TANGANYIKA
LAKE MALAWI
Carbonate platform
Red Sea-type sulphide pool
SEDEX
MVT
Carbonate reef
UPPER
Partial melting
MANTLE
Ernest Henry-style Cu-Au
Telfer-style Au
Olympic Dam-style Fe-Cu-Au-REE-U
Skarns
Sedimentary rocks
Granitoid intrusives
Mineralisation
Breccia pipe
Sea water fluids
Juvenile fluids CO_2, CH_4, F, CL, N_2, H_2, etc.
Exhalites
proximal facies
distal facies
Intercalated sediments and volcanics
LAKE
MAGMA
Basement rocks
Organic-rich sediments
Fluid flow
Igneous intrusions
Hydrothermal deposit

important resources of Fe–Ti–V and Ni–Cu–Co, is not clearly understood. *Windley* (1995, p. 264) proposed a model that attempts to explain this type of anorogenic magmatism. He suggested that supercontinent breakup is initiated as a result of extension above a mantle plume head, with the production of lower crustal melts, including anorthosites.

Iron–Ti–V ores (ilmenite–magnetite) form by magmatic segregation triggered by episodic increases in f_{O_2}, resulting in the development of a Fe–Ti oxide liquid that becomes trapped in the interstices of plagioclase and clinopyroxene crystals, a mechanism not dissimilar to that responsible for the formation of chromitite layers (*Reynolds*, 1985). A giant Fe–Ti–V deposit is sited in the 1.5 Ga Suwalki anorthosite massif in Poland (*Morgan* et al., 2000). In the southern hemisphere, one of the world's largest massif-type anorthosite complex is the north-trending, 300-km long, ca 2.0 Ga Kunene Intrusive Complex, exposed across the Kunene River, marking the border between Angola and Namibia (*Ashwal* and *Twist*, 1994). The Complex hosts large bodies of massive titaniferous magnetite and disseminated ilmenite (*Schneider*, 1992) as well as economically important sodalite (*von Seckendorff* et al., 2000).

Mesoproterozoic (ca 1.3 Ga) anorthosite-leuconorite-troctolite intrusions are present in Labrador (Canada). In Eastern Labrador, these intrusions are part of a larger igneous body of batholitic dimensions, known as the Nain Plutonic Suite, which straddles the Abloviak Shear Zone and intruded at approximately 1.33–1.29 Ga. The Nain Suite includes granites, anorthosite, diorite and troctolite rocks. Members of the Suite contain the Voisey's Bay Ni–Cu–Co deposit, hosted in troctolite rocks that are interpreted to be the feeder to overlying intrusions (*Naldrett*, 1999). This feeder is from 10 to 100 m thick, and within its wider parts massive sulphides form elongate lenses. Re–Os isotope systematics indicate that the parental magma of the Voisey's Bay intrusion was probably basaltic, which is consistent with the low Ni/Cu ratio of the ore (*Lambert* et al., 1999). Recent research, based on S, C, O isotopic data and Se/S ratios, supports the idea that a combination of S-rich pelitic country rock and melting of xenoliths was a strong contaminant to the basaltic magma that gave rise to the Voisey's Bay intrusion (*Ripley* et al., 1999, 2002). *Li* and *Naldrett* (1999) proposed a tentative model for the origin of the Voisey's Bay deposit. According to their model, the Voisey's Bay magma rose to the lower chamber and became S-saturated by interaction with S-rich and graphite-bearing gneissic rocks. Fractionation of the magma with associated loss of Ni and Cu to sulphides and olivine, was followed by a pulse of new

Fig. 6. Schematic illustration showing the East African Rift System and hypothetical ore-system models. 1) represents the Red Sea advanced rift, where oceanic crust is forming and where brine and sulphide pools are actively forming. 2 and 3) represent a continental rift with anorogenic alkaline magmas at depth, from which magmatic and hydrothermal deposits of the Fe oxide–Cu–Au–REE style (names of deposits in figure refer to well-known Australian examples) may be forming. 4) represents a rift system with a deep lake, in which sediments and volcanic materials accumulate together with subaqueous exhalites. 5) also a rift-lake system, where sediments are organic-rich and where subaqueous hydrothermal venting occurs. This figure is based on and modified from *Stager* (1990)

magma into the lower chamber. This led to the injection of the Ni and Cu depleted melt and the segregated sulphides along the feeder to form the upper chamber (olivine gabbro). The sulphides were forced into fractures of the feeder walls and were also precipitated within the feeder in zones where the flow of the melt decreased (Fig. 5B). Redeposition and upgrading of sulphides took place during the continuing influx of magma into the upper chamber (*Li* and *Naldrett*, 1999).

The unusual and enigmatic Okiep copper district of Namaqualand (South Africa) is possibly an anorthosite magmatic system, within a mantle-plume generated mid-Proterozoic rift system (Namaqua Metamorphic Complex) (*Willner* et al., 1990). The 1060 Ma Koperberg Suite includes anorthosites, diorites and pyroxenites, forming about 1700 pipes and dyke bodies (known as basic bodies) in granulite facies gneisses and granitoids of the Okiep Group (*Lombaard* et al., 1986). The Okiep copper ores contain abundant bornite, which is not considered typical of magmatic sulphides (*Maier*, 2000). The origin of the sulphide ores of the Okiep basic bodies still defies explanation, but it is possible that original primary magmatic sulphides were modified, perhaps as a result of high-temperature metamorphism (*Maier*, 2000).

Included in anorogenic magmatism are kimberlites, lamproites and carbonatites. These rocks are known to have isotopic signatures (He, Os, Sr, Nd, Pb, O) similar to ocean island basalts (e.g. *Nelson* et al., 1988), and as such are considered as part of plume magmatism (*Bell*, 2001). They may represent distal expressions resulting from the channelling of plume material along pre-existing lithopheric breaks, small degrees of melting of enriched/metasomatised lithosphere and/or crustal melts. A link with mantle plumes is invoked by a number of workers (e.g. *Bell*, 2001; *Schissel* and *Smail*, 2001). *Ebinger* and *Sleep* (1998), suggested that a mantle plume may focus its flow towards craton-mobile belt boundaries, where at depths of >150 km small volumes of melt can be produced by decompression (Fig. 1). Although a connection with mantle plumes is by no means certain, it may be that thermal perturbation of anomalously hot mantle triggers melting, particularly in region of thinned or weakened lithosphere, even if these are a long distance away.

Hydrothermal ore deposits

The wider link between hydrothermal ore deposits and mantle plumes is explored in this section. In addition to direct generation of magmas, thermal anomalies associated with mantle plumes constitute powerful heat sources in the crust. These may be responsible for the inception of crustal scale hydrothermal circulation and high-T and low-P metamorphism, which may result in a wide range of ore deposits in hotspot related rift systems (*Pirajno*, 2000, 2004).

Numerous vein deposits in the sedimentary basins of the Capricorn Orogen in Western Australia are considered to have been formed as a result of regional scale hydrothermal convection related to heat energy resulting from the emplacement of sill complexes of the 1070 Ma Warakurna LIP in central Western Australia (*Wingate* et al., 2004; *Pirajno*, 2004). Similarly, the thermal energy released during cooling of the Bushveld Igneous Complex in South Africa, activated crustal-scale hydrothermal convection cells from which many Sn, Cu, Au and W vein deposits formed in the surrounding rocks of the Transvaal basin (*Pirajno*, 2000). The geological record shows several rift systems associated with LIPs, at least since the Late Archaean-Early Proterozoic (~2.7–2.5 Ga; e.g. rift systems of the Fortescue, Western Australia, and Ventersdorp, South Africa). Rift systems act as major conduits for both magmas and hydrothermal fluids and rift-related ore deposits are characterized by metal associations that reflect the interaction of anorogenic magmas and the rifts' volcano-sedimentary successions. Examples of rift systems that are well-endowed with hydrothermal ore deposits are, the 1.1 Ga Mid-Continent rift system (USA; *Nicholson* et al., 1992), the Jurassic Limpopo-Sabi-Lebombo triple junction rifts (southern Africa; *Pirajno*, 2000), the Mesozoic Panxi rift and the mid-Proterozoic Langshan-Bayunobo rift (both in China; *Gilder* et al., 1991), which hosts the world-class Bayan Obo REE deposit, referred to earlier. Volcano-sedimentary successions of rift basins host a great variety of hydrothermal ore deposits as well as hydrocarbon reservoirs. Sedimentary-rock hosted metalliferous ore deposits (e.g. Pb–Zn–Ba–Cu–Au–Ag; SEDEX) typically occur in intracontinental rift basins. Well-known examples of SEDEX deposits of rift basins include, Gamsberg-Aggenays in South Africa; McArthur River, Broken Hill, Hilton, Lady Loretta, Mt. Isa in Australia; Sullivan in Canada, to mention a few. It is interesting to note that the majority of SEDEX deposits formed between the early and mid-Proterozoic (~1.9–1.0 Ga), which coincide with plume events at 1.9, 1.6, 1.2–1.3, 1.1 Ga (*Ernst* and *Buchan*, 2002). Perhaps the best modern analogues of at least some SEDEX type deposits are provided by Red Sea brine pools (*Pottorf* and *Barnes*, 1983). The Red Sea can be considered as the northern arm of the great, 5000 km-long, East African Rift System (EARS), which represents a geodynamic setting for a wide range of rift-related ore deposits. The EARS provides a window into what must be a giant intracontinental ore making factory, comparable to present-day oceanic spreading centers. The tectonic and magmatic environments of the EARS are schematically illustrated in Fig. 6. The EARS includes massive sulphide deposits in the Red Sea brine pools, hydrothermal sediments deposits in rift lakes, and deep-seated anorogenic magmatic systems beneath the rift floors.

Iron-formations

Banded iron-formations (BIF) may owe their origin to hydrothermal effluents from submarine LIPs (oceanic plateaux) (e.g. *Barley* et al., 1997). BIF are chemical – sedimentary units containing in excess of 15% Fe, or 30% Fe-oxides, consisting of Fe oxides (hematite and/or magnetite) alternating with chert and silica bands. Clastic iron-formations, due to the re-working of BIF in shallow waters, are known as granular iron-formation (GIF). In the Precambrian there are two main types of iron-formations: 1) Algoma-type associated with volcanic sequences in back-arc environments; and 2) Superior-type associated with sedimentary sequences in continental shelves of passive tectonic margins. Algoma-type iron-formations are relatively thin, discontinuous and typically exhibit facies changes from oxide to carbonate, silicate and sulphide. Their origin is related to volcanic exhalations vented on the sea floor in back-arc settings. Algoma-type iron formations are not considered further in this paper.

The best-known Superior-type iron-formations, in terms of both geological interest and economic value are those of the Transvaal Supergroup (South Africa), the Hamersley Group (Western Australia), the Minas Gerais (Quadrilatero Ferrifero Brazil), the Krivoy Rog basin (Ukraine) and the classic Gunflint, Biwabik and Sokoman iron-formations of the Lake Superior region in North America. Literature on iron-formations is voluminous, and useful and comprehensive reviews are provided by *Kimberley* (1989a, b) and *Trendall* (2002). Superior-type iron-formations are most abundant in the Late Archaean-Early Proterozoic, although iron-formations are also known from the Early Archaean and the Phanerozoic. The Proterozoic iron-formations are commonly associated with giant manganese deposits, such as those of the 2.2 Ga Kalahari Mn field in South Africa (the largest in the world with about 13 billion tonnes of ore) and Minas Gerais in Brazil. This close spatial relationship is related to the chemical affinity of Fe and Mn and therefore the same solutions that are enriched in Fe are also enriched in Mn (*Schissel* and *Aro*, 1992; *Cornell* and *Schütte*, 1995). The origin of the iron and manganese formations requires that large amounts of these metals be brought into solution as reduced species (Fe^{2+} and Mn^{2+}), which are then oxidised (Fe^{3+} and Mn^{3+}, Mn^{4+}) and precipitated as Fe and Mn oxides and carbonates. Deposition of BIF reached maxima at about 2.7 and 1.9 Ga, which coincide with maxima in mantle plume activity (*Isley* and *Abbott*, 1999). The 1.9 Ga period also coincides with peak production of black shales and the deposition of intracratonic sediments in sag basins and continental shelves (*Condie* et al., 2000). Two issues are contentious for the genesis of these formations: one is the source of the metals, the other is the amount of oxygen that is necessary to induce oxidation of the Fe and Mn. For the source of the metals, two possibilities are considered: one is that the Fe is derived from the weathering of iron-rich rock types (e.g. continental flood basalts); the second is that the Fe (and Mn) is introduced by extensive subaqueous hydrothermal discharges, in lakes or in ocean basins, or in Red Sea-type brine pools (*Krapež* et al., 2003). Common to both theories is the necessity of a density-stratified system, in which upwelling currents bring the reduced iron from anoxic deeper waters into the oxygenated environment of near surface waters, such as a continental shelf, where the Fe^{2+} is oxidised to Fe^{3+} and precipitated as oxides, carbonates or sulphides. In summary, the required submarine volcanism and related extensive hydrothermal venting, presence of appropriate depositional systems (e.g. continental shelf), and deposition of black shales, suggest a cause and effect relationship, as advocated by *Barley* et al. (1997), *Condie* et al. (2000) and *Abbott* and *Isley* (2001).

Hydrocarbon and mineral resources at passive margins

The sedimentary and volcanic successions of continental passive margins contain both hydrocarbon and mineral resources, typically salt domes, petroleum and gas traps, gas hydrates, diamond and gold placers, phosphorites, metal-rich black shales, ironstones, carbonate-hosted Pb–Zn deposits (*Mitchell* and *Garson*, 1981; *Magoon* and *Dow*, 1994; *Henriet* and *Mienert*, 1998; *White* et al., 2003).

Passive margins evolve from continental rifting, breakups, advanced rifting and sea floor spreading. Subsequent dynamic processes that control their development, structural style and depositional environment may not have any direct connection with mantle plume events. However there are well-documented cases, as for example passive margins that border both sides of the Atlantic Ocean and the west coast of Australia, which are associated with flood volcanisms and are related to rifting and breakup caused by mantle plumes (*Menzies* et al., 2002). The best example of conjugate passive continental margins is that of west Africa and the eastern seaboard of North America, which are temporally and spatially associated with the rifting of the Central and South Atlantic above mantle plumes at about 200 Ma and 138 Ma respectively (*Wilson*, 1997). In the South Atlantic, the Namibian passive margin, investigated in detail using on-shore and off-shore seismic transects and petrophysical modelling, is predominantly volcanic (*Corner* et al., 2002; *Trumbull* et al., 2002). The Namibian passive margin is related to continental rifting linked to the Tristan da Cunha hot spot and the Parana-Etendeka LIP. *Prendergast* (in press) recognised a passive margin associated with a LIP in the 2.7 Ga Bulawayan Supergroup, Zimbabwe Craton. This Archean LIP, known as Reliance Unit, consists of komatiitic sill-flow complexes and a submarine flood basalt sequence, is part of a passive continental margin succession and may be the ancient analogue of a Namibian-style margin. It is important to note that the ultramafic components of the Reliance Unit host several significant Ni–Cu sulphide deposits (*Prendergast*, 2003). In addition, there are Ni–Cu sulphide deposits in the intrusive phases of mafic lavas in the Bulawayan Supergroup (Prendergast, written comm. 2003). If the analogy with a Namibian-style margin is valid, then volcanic-dominated passive margin successions in the younger geological record may constitute good exploration targets for Ni–Cu sulphide deposits.

Conclusions

The mantle plume theory is based on a combination of surface observations of large topographic swells, intraplate magmatism and geophysical data (e.g. seismic tomography). The theory holds that plumes of hotter than normal mantle material may originate from thermal boundary layer instabilities, at the 670 km discontinuity and from the core-mantle boundary (CMB). The deep plumes rising from the CMB are thought to be responsible for intraplate volcanism. In the Archaean, mantle plumes were hotter and underwent higher degrees of melting, hence the common komatiitic components. It follows that the pattern of mantle covection changed with time, perhaps in a cyclic fashion, from two-layer to single- and back to two-layer convection, causing changes in the driving mechanisms that control plate movements and consequently the diverse patterns of Archaean, Proterozoic and Phanerozoic tectonics. The mantle plume model, although not universally agreed upon, provides a global framework for the understanding of intraplate tectono-magmatic and ore-forming processes, as well as continental assembly, breakup and rifting. As discussed in this paper, mantle plume events result in the emplacement of large igneous provinces or LIP, characterised by vast outpourings of dominantly basaltic lavas in, geologically, very short times. These form oceanic plateaux and chains of volcanic islands on the sea floor, and continental flood basalts on land. Seismic data indicate that layered mafic-ultramafic intrusions constitute a large proportions of a LIP, and what we see in present-day outcrops of

LIP may only be the "tip of the iceberg". This implies that outcropping fossil magma chambers (e.g. layered complexes and sill complexes), probably represent the roots of what must have been ancient LIP, whereas the great dyke swarms, represent the feeders.

Thus, mantle plume-related intraplate magmas directly or indirectly are involved in the making of a wide range ore deposits. These include magmatic ore deposits that form in magma chambers or in feeders or in lava channels, and hydrothermal ore systems in a variety of tectonic settings, from anorogenic igneous complexes, including carbonatites and kimberlites, that complement and accompany many LIPs, to those that form by the circulation of fluids in giant hydrothermal systems in rift zones, themselves originated by the impact of mantle plumes onto the base of the lithosphere. Indeed, the introduction to the crust via mafic-ultramafic melts, of siderophile and chalcophile elements can be directly linked to mantle upwellings to the lithosphere. Further concentration and/or selective uptake of these elements takes place during hydrothermal convection in the crust, with the heat supplied by mantle-derived magmas.

As a colourful and perhaps fitting finale to the mantle plume connection, I invite the reader to reflect on the real possibility that the speciation of hominids, which culminated with the advent of *Homo sapiens*, is after all the result of the tectonic and related environmental forcing, around 3.0–2.0 Ma, that affected the Afar region (*Kimbel*, 1995; *Redfield* et al., 2003), thereby providing the ultimate link to a mantle plume.

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