



Intracontinental anorogenic alkaline magmatism and carbonatites, associated mineral systems and the mantle plume connection

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ABSTRACT

Intracontinental alkaline complexes and carbonatites commonly occur in rift-related settings, best exemplified by the Cenozoic–present day East African Rift System. Examples in the ancient geological record are more difficult to unravel due to subsequent overprinting during tectono-thermal events unrelated to the original rifting. In the present paper, with the exception of the Bayan Obo carbonatite, I have used examples of anorogenic alkaline complexes and carbonatites that are effectively pristine and unaffected or little influenced by later geological events. In this context, I report on two alkaline complexes and associated mineralisation in Namibia (Brandberg and Erongo) related to the opening of the South Atlantic and the Paraná–Etendeka large igneous province. With regard to carbonatites and associated mineralisation, I discuss the exceptionally well-preserved Kruidfontein and Goudini in South Africa, Mount Weld and Gifford Creek ferrocarbonatite complex in Western Australia, the controversial Bayan Obo in northern China and the carbonatites of the Mianning–Dechang belt in SW China. It is proposed that all these igneous systems are formed by processes of partial melting in metasomatised lithospheric mantle, induced by upwelling mantle plumes. Furthermore, it is contended that these alkaline and carbonatite igneous systems are generally formed in the distal sectors of a mantle plume due to lateral transfer of volatiles, resulting in low degrees of partial melting in the metasomatised sub-continental lithospheric mantle.

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“One thing I have learned in a long life: that all our science, measured against reality, is primitive and childlike — and yet it is the most precious thing we have.”

[Albert Einstein]

1. Introduction

Alkaline magmatism is generally associated, in space and time, with extensional tectonics and more specifically with rifting events. The causes and mechanisms of rifting are complex and are beyond the scope of this contribution, however the interested reader should consult Burke and Dewey (1973), Fitton and Upton (1987), White and McKenzie (1989), Wilson (1989), Kampunzu and Lubala (1991), Storey et al. (1992), Olsen (1995), Condie (1997), Şengör and Natal'in (2001) and Fourel et al. (2013).

Magma genesis in rift environments is thought to be related to the ascent of mantle melts and the effects of their interaction with lithosphere materials. The main products of rift magmatism are fissure-fed tholeiitic flood basalts (large igneous provinces — LIPs), mafic–ultramafic intrusions, dyke swarms and intracontinental alkaline provinces

(Pirajno, 2000). The latter is characterised by anorogenic magmatism, which includes alkaline ring complexes, carbonatites, kimberlites, lamproites, monogenetic and central-vent alkaline volcanoes. In terms of tectonic environments, three main scenarios can be considered in which alkaline complexes and associated igneous rocks are emplaced (Bonin, 1986; Kinnaird and Bowden, 1991). They are: 1) doming of the crust, followed by rifting; 2) “hot spot” traces with time–space migrating magmatism; and 3) alkaline complexes located along crustal-scale shear zones.

In this paper, I direct my attention to the mineral systems that are related to intracontinental alkaline magmatism and associated carbonatites and examine their links with rifting processes and mantle plumes. Rift structures act as major conduits for both magmas and hydrothermal fluids and it is reasonable to assume that ore deposits are formed in which metal associations are related to both magma and interaction of the fluids with rift-related volcano-sedimentary sequences. Based on geochemical and stable isotope studies (e.g. Nelson et al., 1988; Ngwenya, 1994; Bell and Simonetti, 2010; Kogarko et al., 2010), it appears that in the first instance the origin of the mineralising fluids is essentially magmatic, although there are many cases in which the mineralisation was formed through processes of hydrothermal activity related to the influx of meteoric waters and their mixing with high-temperature fluids of magmatic origin (e.g. Kinnaird and Bowden, 1991; Lowenstern, 2001; Bell et al., 2013). The ore-forming processes are always accompanied by alteration of the

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host rocks, including alkali metasomatism (fensitisation), greisenisation, propylitisation, silicification and phyllosilicate alteration (Pirajno, 2013a).

Many, if not most of the structures associated with alkaline complexes are ancient zones of lithospheric structural weakness, which have been repeatedly reactivated during tectono-thermal events. These structures may extend deep into the crust and to the subcontinental lithospheric mantle (SCLM). Thus, several alkaline provinces are localised along crustal scale shear zones, along the margins of cratonic blocks, within Proterozoic mobile belts, and along reactivated zones of faulting and rifting (Begg et al., 2009).

Most primary alkaline magmas are mafic, rather than ultramafic, and in many cases, prolonged lithospheric and/or crustal residence time results in a wide range of rock types, including felsic derivatives far removed from compositions in equilibrium with mantle olivine. The rock types considered here range from mafic to felsic and usually contain feldspathoids (e.g. nepheline) and/or alkali-pyroxene and alkali-amphibole. The silica-oversaturated rock types (generally quartz-bearing) range from monzonite to syenite, alkali-syenite and alkali-granite, and their volcanic equivalents. Nomenclature for the silica-undersaturated rock types (generally feldspathoid-bearing) can be confusing and plagued by locality-based names. The terminology published by Le Maitre (2002) provides a sufficient guide to their classification.

The relationship between alkaline magmatism, thermal anomalies in the mantle and extensional tectonics has been mentioned above. Whether rifting is a passive or active process, the end result is thinning the crust and upper lithosphere, causing lithospheric mantle melts and the asthenosphere to rise. This may result in:

- decompression melting of the upwelled mantle;
- transfer of volatiles into the SCLM and the lower crustal regions;
- melting of the lower crust due to the addition of heat, either through conduction or through intrusion, coupled with the possible effects of introduced volatiles.

The amount of magma generated depends on the temperature of the upwelled mantle and the amount of lithospheric thinning, which are also related to the rate of extension. High rates of extension (e.g. continental break-up) are often associated with voluminous production of basaltic magmas. Lower rates of extension (e.g. intra-continental rift/graben structures) result in lower degrees of partial melting, and more alkalic and silica-undersaturated magmas. In intra-continental settings, the range of resulting rock types is often very wide and not restricted to silica-undersaturated compositions. It is quite common for alkaline complexes to consist of both silica-undersaturated and silica-oversaturated rocks, as is the case for many of the non-carbonatitic complexes of southern Africa (Woolley, 2001; Woolley and Kjarsgaard, 2008). The relationship between these has long been a point of extensive debate. The two types of rocks are separated by a thermal divide (Ab–Or join) that, in theory, cannot be crossed via closed system fractionation of magmas (e.g. Rayleigh fractionation). Hence, two separate magma types have usually been invoked to explain the presence of both silica-undersaturated and silica-oversaturated rocks in alkaline complexes, with the silica-oversaturated rocks generally regarded as products of crustal, rather than mantle or SCLM, melting. Isotopic evidence usually shows the silica-undersaturated rocks to be of mantle or SCLM origin, but does not always show the silica-oversaturated rocks to be of crustal derivation (Nelson et al., 1988; Begg et al., 2009).

In terms of mineral exploration, the recognition of the different magma types may be very important. As a generalisation, mineralisation tends to be affiliated with the silica-oversaturated, rather than the silica-undersaturated rock types. Furthermore, it is those silica-oversaturated rocks that show significant, or dominant, crustal origins which are most commonly and strongly mineralised (Pirajno, 2009 and references

therein). On the other hand, silica-undersaturated magmas can host critical elements, such as rare earth elements (REE), P, U, Th, and F.

In this contribution a selection of diverse intraplate alkaline magmatic systems and carbonatites from Namibia (Brandberg, Erongo), South Africa (Kruidfontein, Goudini), Western Australia (Mount Weld and Gifford complex) and China (Bayan Obo and the carbonatites of the Mianning–Dechang belt) are examined. Except for the Mianning–Dechang belt, I have been involved in various degrees with all of the above mentioned complexes. These alkaline intrusions and carbonatites host REE, P, U, and F mineralisation ranging from non-economic to sub-economic in the present time, to economic. In the present article the emphasis is on carbonatites and related mineralisation, due to their ever increasing economic importance, as explained in the pages ahead.

2. Anorogenic alkaline plutonic and volcano-plutonic complexes

Anorogenic alkaline complexes are generally characterised by unusual concentrations of non-hydrous volatiles (CO₂, B, F), REE and LILE incompatible elements (P, Zr, Ba, Nb, U, Th, Sn, Ta and W) (Bailey et al., 1974; Bailey, 1985; Bailey and Macdonald, 1987; Edgar, 1987; Owen-Smith et al., 2013). These elements may form their own minerals, or they may enter the lattice of rock-forming minerals (e.g. micas), that have crystallised in the subsolidus range of pressure and temperature. The processes that lead to the enrichment of these elements are linked with the activity of residual alkali-rich fluids. Consequently, their concentration, either in rock-forming minerals or in their own mineral species, is largely dependent on the original magma composition and the nature of the source regions where partial melting occurred.

From the stand point of economic mineralisation, important rock associations of alkaline magmatism are: kimberlites, lamproites, mafic ultrapotassic rocks, syenite–pyroxenite–ijolite–carbonatite complexes, carbonatite plugs and dykes, alkali basalts, and peralkaline granite–syenite–gabbro assemblages (Mutschler et al., 1985; Pirajno, 2013a). These rocks occur as dykes, pipes, hypabyssal and subvolcanic intrusions and as central volcano-plutonic structures, many of which form ring complexes. Most important for economic mineral deposits are kimberlites and lamproites (diamonds), carbonatite and volcano-plutonic complexes. Carbonatites are especially significant because they constitute a major source of Nb, P, Zr, Ti, Th, Sr and REE. Most economic mineral deposits (excluding diamonds in kimberlites and ultrapotassic mafic rocks) occur associated with central vent-type complexes of which two main groups are recognised (Bowden, 1985): 1) alkali-granite–syenite (saturated–oversaturated); and 2) ijolite–carbonatite (undersaturated). They are schematically illustrated in Fig. 1. The idealised cross-sections shown in Fig. 1 also illustrate the importance of the level of erosion, and hence age, in terms of the preservation potential of the contained mineral deposits.

In this section, I discuss two alkaline plutonic and volcano-plutonic complexes, Brandberg and Erongo, respectively and both in Namibia. The reason is two-fold: one is that they do provide good examples of anorogenic intracontinental magmatism, in the specific case related to the Paran –Etendeka flood volcanism and the Gondwana continental breakup, and the other reason is because both complexes have been the focus of my study during my work time in Namibia. The distribution and setting of alkaline complexes and carbonatites in south-western Africa is shown in Fig. 2.

2.1. Brandberg

The Brandberg alkaline complex (Hodgson, 1973; Diehl, 1989, 1990; Bowden et al., 1990; Pirajno, 1994; Watkins et al., 1994; Schmitt et al., 2000; Trumbull et al., 2004), forms a conspicuous circular topographic feature approximately 26 km across rising to 2573 m above sea level and 2000 m above the surrounding desert (Fig. 3). The complex is intruded into metasediments of the Damara Orogen and overlying

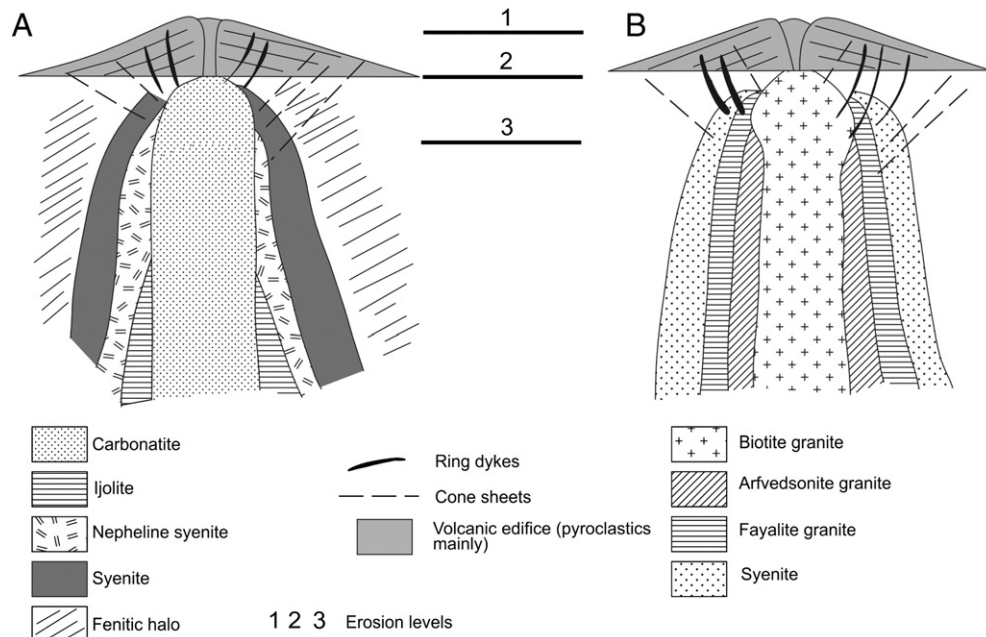


Fig. 1. Idealised cross-sections of central-vent type alkaline complexes; (A) ijolite-carbonatite and (B) alkali-granite-syenite. After Bowden (1985).

volcanic and sedimentary rocks of the 132–131 Ma Etendeka volcanic sequence (Renne et al., 1996), which is part of the Paraná–Etendeka large igneous province (Ernst and Buchan, 2001) (Fig. 2). The Brandberg complex is an A-type metaluminous to peralkaline granitic intrusion (Trumbull et al., 2004) comprising a main massif of biotite–hornblende granite rimmed by an almost continuous contact breccia zone which, according to Diehl (1989, 1990), is surrounded from the margins towards the interior by annular continuous to discontinuous outcrops of fayalite–hedenbergite granite, edenite–augite granite, edenite–biotite granite and biotite–hornblende granite. In addition a number of small ring structures have also been recognised, whereas on the

southwestern margin of the complex is the peralkaline Amis complex (Fig. 4). Roof pendants of Etendeka volcanics are present on the eastern side of the complex (Fig. 4). The hornblende–biotite granite of the main massif, as reported by Schmitt et al. (2000) consists of plagioclase (36–38 vol.%), rimmed by perthitic K-feldspar (23–28 vol.%), quartz (19–24 vol.%), and granophyric intergrowths of K-feldspar and quartz. Albite is present as an interstitial phase. Edenitic hornblende with relic cores of augite and biotite make up a total of 10–15 vol.%. Accessory minerals include Ti-rich magnetite, ilmenite, apatite, zircon, allanite, titanite, chevkinite and thorite. Alteration minerals include chlorite, epidote, sericite and hematite. Schmitt et al. (2000) disputed some of

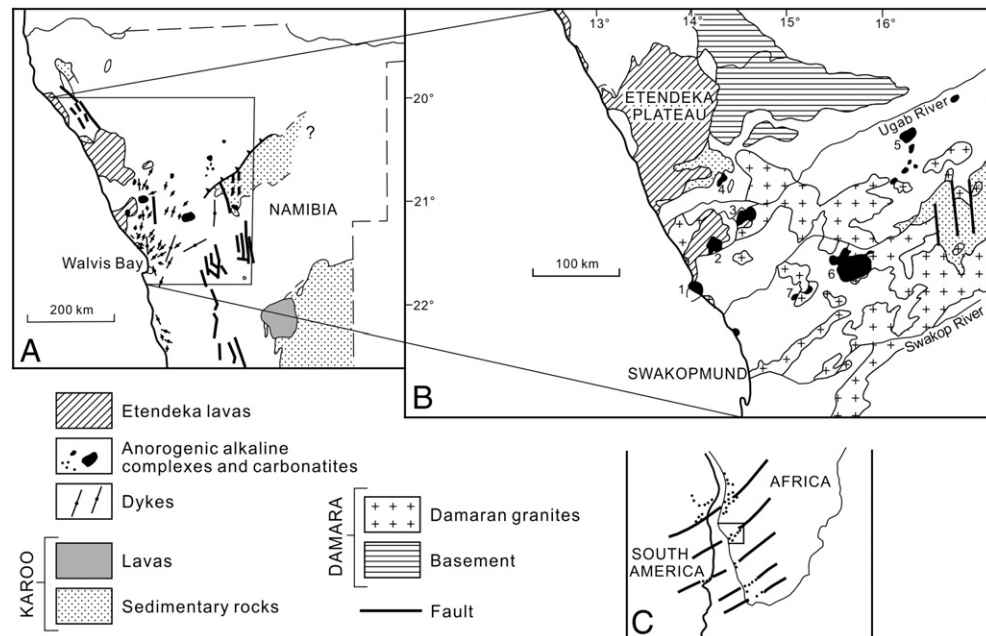


Fig. 2. (A) Schematic illustration showing distribution of Etendeka and Karoo continental flood basalts and associated dyke swarms in north-central Namibia; (B) distribution of Etendeka lavas and associated anorogenic ring-type alkaline complexes and carbonatites; selected complexes named in this figure are 1) Cape Cross, 2) Messum, 3) Brandberg, 4) Doros, 5) Paresis, 6) Erongo and 7) Spitzkop; (C) relationship of alkaline complexes in southwestern Africa and the east coast of South America to transform directions. After Marsh (1973).



Fig. 3. The Brandberg mountain seen from the south.

the findings by Diehl (1990) and found no fayalite in their samples and what had been described hornblende–pyroxene granite is a biotite–hornblende granite. In the central areas of the main equigranular hornblende–biotite granite massif are dykes of leucocratic pegmatitic and aplitic granite, with no preferred orientation. A small round body of monzonite outcrops in the Naib River on the western side of the complex and is likely associated with ring fractures (Fig. 4). Northwest- and north-trending fractures cut across the Brandberg massif (Fig. 4). To the north–northeast are peripheral porphyritic trachydacite dykes, whose age relationship to the main massif is not clear, but Schmitt et al. (2000) suggested that these dykes could be either coeval or late intrusions. In spite of its small exposed size, the Amis peralkaline intrusion in the southwest of the main massif is important in terms of its mineralisation potential (see Section 3.1).

The Amis peralkaline granite intrudes the Brandberg hornblende–biotite granite (Fig. 4). The rock types of the Amis intrusion include arfvedsonite- and aegirine-bearing granitoids, which show pegmatitic, aplitic and fine-grained textural varieties. Quartz, albite and microcline are the felsic mineral phases. Another variety of peralkaline granite, reported by Schmitt et al. (2000) is a fine-grained spotted rock consisting of radiating clusters of aegirine, together with quartz, albite and astrophyllite. The Amis peralkaline granite is enriched in REE minerals, as discussed in Section 3.1. According to Bowden et al. (1990) and based on previous work by Diehl (1989), the Amis intrusion was emplaced within a ring structure and is characterised by banding, sheeted units or mineral layering (Fig. 5), consisting of various proportions of arfvedsonite–aegirine–astrophyllite granite, arfvedsonite–aegirine granite, arfvedsonite granite, hematitic alkali granite and tourmalinised granite, reaching a maximum thickness of 360 m. The hematitic granite is thought to be the result of post-magmatic processes characterised by quartz–hematite intergrowths that replace the arfvedsonite.

Watkins et al. (1994) reported a range of K–Ar ages on mineral separates from Brandberg of 135 ± 1.5 Ma to 125.4 ± 1.3 Ma. These ages span a period of 10 million years. More reliable $^{40}\text{Ar}/^{39}\text{Ar}$ age data were provided by Schmitt et al. (2000), shown in Table 1, ranging from ~133 to ~130 Ma. These authors clearly demonstrated that the Brandberg complex was emplaced in a relatively short period of time during the main phase of the Etendeka volcanism. Trumbull et al. (2004) examined the Nd and O isotope systematics of the Namibian anorogenic intrusions, including the Brandberg intrusion. They reported a good correlation of Nd and O isotope ratios for all the Namibian

anorogenic intrusions, mafic and silicic, which are close to those of the Tristan da Cunha hot spot, implying a mantle plume link.

2.2. Erongo

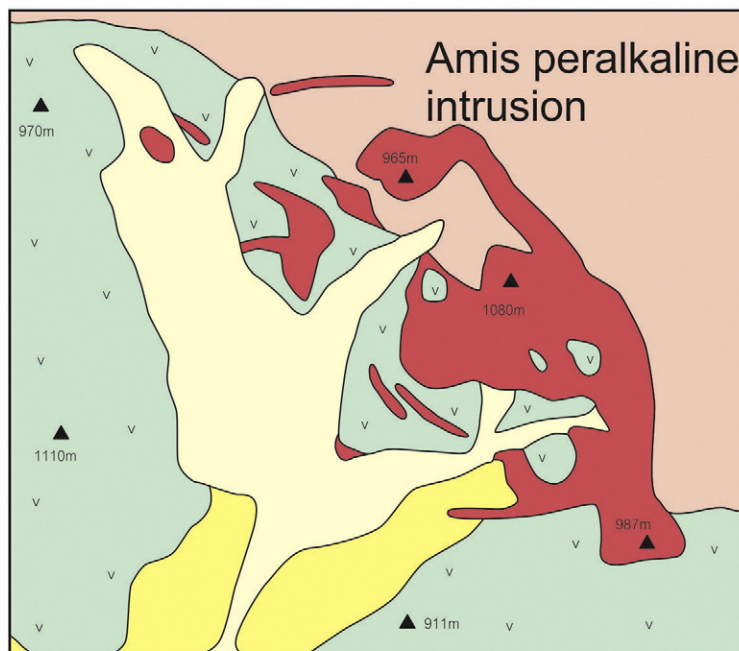
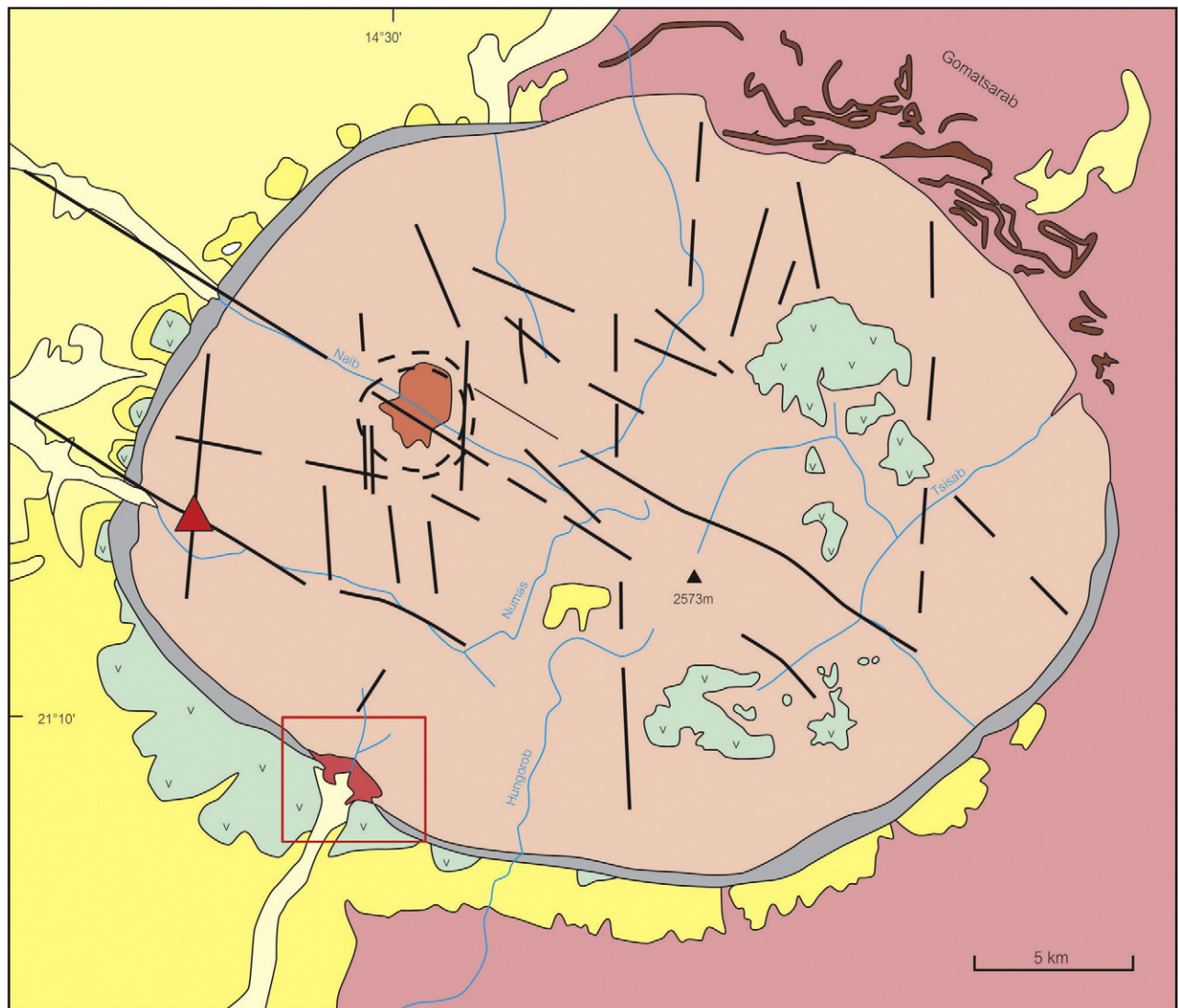
The Erongo volcano–plutonic complex is the largest (about 30 km in diameter, but reaching 50 if the outlying gabbro cone sheet is included) of the Namibian anorogenic complexes (Pirajno, 1990, 1994; Pirajno et al., 2000) (Fig. 2). Dated rock units of the Erongo complex are shown in Table 1. The age data clearly show that the Erongo magmatism had a span of about 2 Myr or less and was likely formed in a single event of short duration contemporaneously with the Etendeka flood basalts, which ceased at about 130 Ma (Wigand et al., 2004).

The Erongo complex is a well-preserved caldera structure, surrounded by a granitic ring dyke and a mafic cone sheet (Fig. 6). The Erongo magmatism is characterised by three phases, from oldest to youngest (Pirajno, 1990): 1) an early mafic phase, which consists of a succession of tholeiitic lavas, dykes and the above-mentioned cone sheet; 2) a felsic phase of pyroclastic flows; and 3) emplacement of mafic alkaline plugs. The early mafic phase is the result of voluminous effusions of tholeiitic lavas, forming a succession up to 300 m thick, whose areal extent may have been much greater than the present-day distribution. This is suggested by the presence of the Krantzberg lava outlier to the northeast, which is spatially associated with greisen style mineralisation. In the upper parts of the lava pile, the composition becomes trachytic and the lavas become intercalated with the earliest pyroclastic units, suggesting that these uppermost trachytic lava flows heralded a change in the composition of the magma towards more felsic compositions.

The felsic phase is dominated by voluminous pyroclastic deposits, which based on stratigraphic, petrographic and geochemical evidence can be subdivided into three distinct sequences. From oldest to youngest, they are: Erongorus ash-flow tuff (EAFt), Ombu ash-flow tuff (OAFt) and rheomorphic rhyolitic rocks (RHEOR). The EAFt rocks represent the first explosive event of the Erongo complex, characterised by having only juvenile material, no lithics, abundant vesicles and lithophysae. The thickness of the EAFt sequence is estimated at between 200 and 370 m, and the composition ranges from andesitic in the lower levels to rhyodacitic at the top. The OAFt is volumetrically the major rock type of the complex, reaching a thickness of at least 500 m. In the east, northeast and southeast the OAFt rests on the tholeiitic lava pile, whereas in the central and western areas it overlies the

Fig. 4. Simplified geology of the Brandberg alkaline complex and the Amis peralkaline complex; position of the fossil hot spring system along the Numas River is also shown (location shown in Fig. 2).

This figure is slightly modified from Schmitt et al. (2000).



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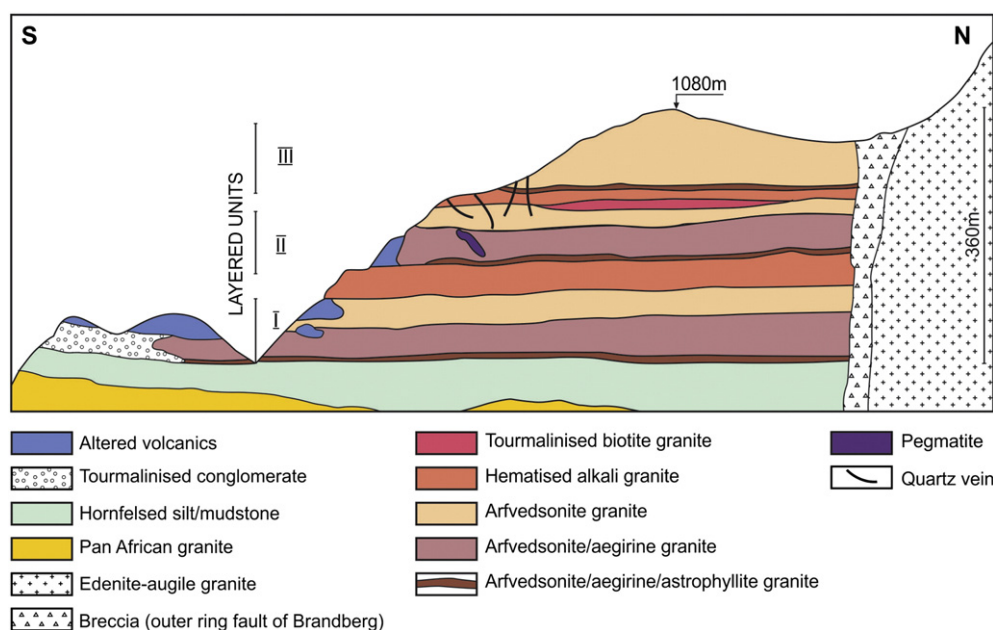


Fig. 5. Simplified section across the Amis intrusion in the southwest of the Brandberg massif. Modified after Bowden et al. (1990).

EAF. OAF rocks are characterised by having abundant crystal and lithic fragments and have compositions ranging from dacitic to rhyodacitic. RHEOR rocks form isolated outliers that generally overlie EAF and OAF units, have a rhyolitic composition and are characterised by flow banding. Tourmaline is present as fine overgrowths and along microfractures. The presence of tourmaline in the RHEOR rocks and their chemistry suggest that they may be genetically related to the tourmaline-rich Erongo Granite.

The Ombu Granodiorite and the Erongo Granite are the subvolcanic intrusives of the complex. The Ombu Granodiorite is a coarse to fine-grained monzogranite to granodiorite rock, charged with numerous xenoliths of the same nature as the lithic fragments in the OAF. The Ombu Granodiorite locally exhibits hydrothermal alteration, of which two trends can be recognised: potassic (K-feldspar + biotite) and phyllic-propylitic (sericite + chlorite + tourmaline). The Erongo

Granite forms plutons distributed between the ring dyke and the caldera structure. The main rock type is a coarse-grained biotite granite, containing abundant topaz and tourmaline, minor apatite and fluorite. A prominent feature of the Erongo Granite is the presence of disseminated quartz-tourmaline nests, up to 30 cm in diameter, veins and stringers, all of which tend to increase in intensity and to coalesce towards the upper parts of the plutons. In addition to tourmaline and quartz, the nests also contain topaz, fluorite, apatite and cassiterite.

The last magmatic event is manifested in the northeast of the complex, and consists of a cluster of small (up to 100 m in diameter) undersaturated mafic plugs, which comprise basanites and tephrites. This area is coincident with a positive gravity anomaly. Trumbull et al. (2003) studied the petrology and isotope systematics of these rocks and suggested a link with thinned lithosphere and decompression melting of a mantle plume.

Table 1
Age data from Brandberg, Erongo, Gifford Creek, Bayan Obo and Maoniuping.

Igneous Cx	Rock/mineral dated	Method	Age (Ma)	Reference
Brandberg	Peralkaline granite/biotite	$^{40}\text{Ar}/^{39}\text{Ar}$	133.0 ± 0.8	Schmitt et al. (2000)
	Biotite	$^{40}\text{Ar}/^{39}\text{Ar}$	132.6 ± 1.8	Schmitt et al. (2000)
	Arfvedsonite	$^{40}\text{Ar}/^{39}\text{Ar}$	132 ± 0.7	Schmitt et al. (2000)
	Arfvedsonite	$^{40}\text{Ar}/^{39}\text{Ar}$	130.9 ± 0.7	Schmitt et al. (2000)
	Amis granite/astrophyllite	$^{40}\text{Ar}/^{39}\text{Ar}$	130.5 ± 0.7	Schmitt et al. (2000)
Erongo	OAF/zircon	U–Pb SHRIMP	135.0 ± 3.2	Pirajno et al. (2000)
	Ombu Granodiorite/biotite	$^{40}\text{Ar}/^{39}\text{Ar}$	132.6 ± 1.0	Wigand et al. (2004)
	Mafic alkaline plugs/biotite	$^{40}\text{Ar}/^{39}\text{Ar}$	132.0 ± 1.0 and 130.8 ± 1.0	Wigand et al. (2004)
	Erongorus rhyodacite	U–Pb SHRIMP	131.8 ± 1.0	Wigand et al. (2004)
	Ombu granodiorite	U–Pb SHRIMP	132.3 ± 1.9	Wigand et al. (2004)
	Erongo Granite	U–Pb SHRIMP	130.3 ± 1.4	Wigand et al. (2004)
	Ekuta Rhyolite	U–Pb SHRIMP	132.6 ± 4.2	Wigand et al. (2004)
	Ferrocarnatite/apatite	U–Pb LA-CP-MS	1075 ± 35	Pirajno et al. (2014)
Gifford Creek ferrocarnatite cx	Monazite	U–Pb–Th LA-CP-MS	1050 ± 24	Pirajno et al. (2014)
	Whole rock, carbonatite dykes	Sm–Nd isochron	1354 ± 59	Yang et al. (2011a,b)
Bayan Obo	Whole rock, dolomitic marble	Sm–Nd isochron	1341 ± 160	Yang et al. (2011a,b)
	Carbonatite dyke	SHRIMP U–Pb	1418 ± 29	Fan et al. (2014)
	Mafic dyke	Sm–Nd isochron	1227 ± 60	Yang et al. (2011a,b)
	Pyrite	Re–Os	439 ± 86	Liu et al. (2004)
	Biotite	K–Ar	31.8 ± 0.7 to 40.3 ± 0.7	Yuan et al. (1995)
Maoniuping REE deposit; Mianning–Dechang belt	Magnesian-arfvedsonite	K–Ar	31.7 ± 0.7	Yuan et al. (1995)
	Biotite in stringer zone		27.8 ± 0.5	Yuan et al. (1995)

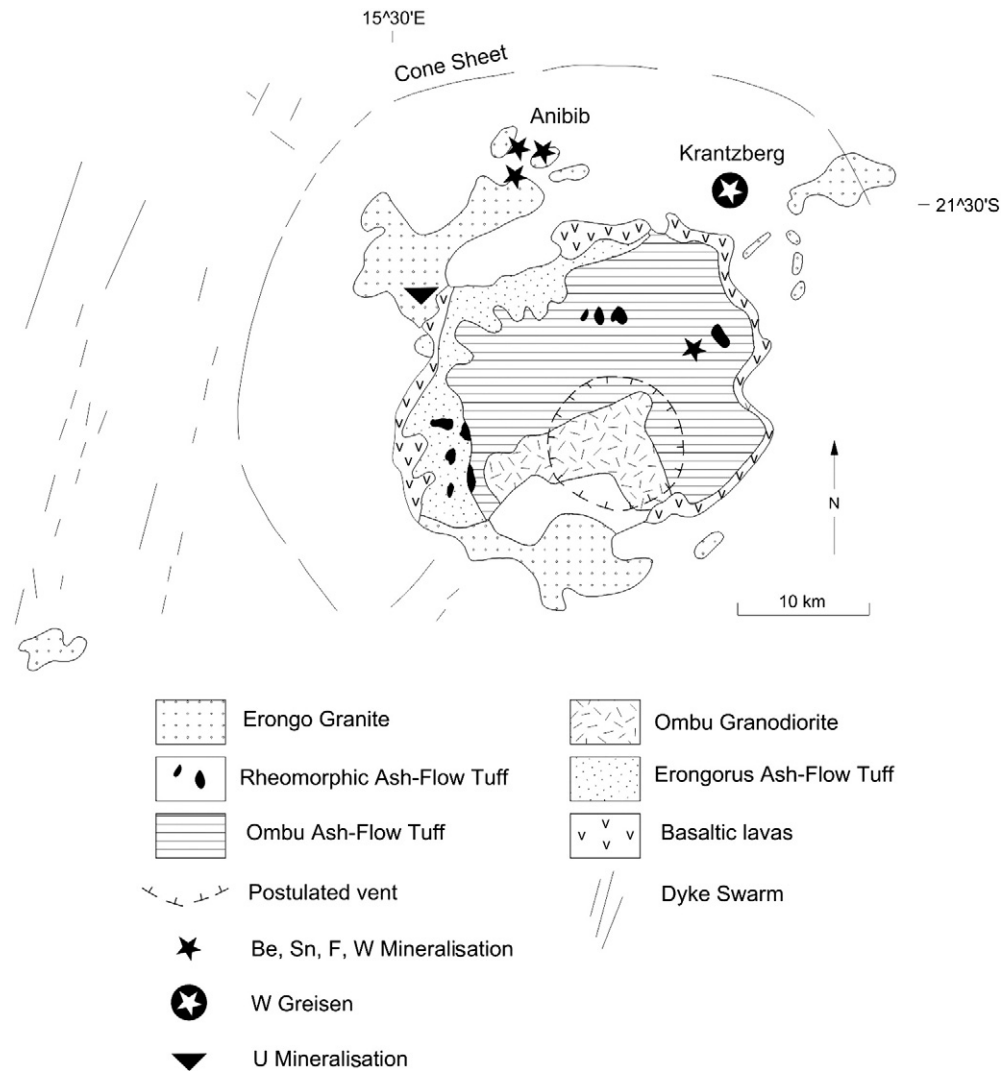


Fig. 6. Simplified geology of the Erongo volcano-plutonic complex (location shown in Fig. 2) and distribution of mineral occurrences. After Pirajno et al. (2000).

The Ombu Granodiorite occupies a vent area, now forming the Ombu plains. OAFt rocks may have come from a shallow magma chamber, from which central eruptions ensued. The eruption column was probably sub-Plinian in character, with a greater gas-thrust component than the EAFT columns. Nevertheless, the absence of air-fall pyroclastics suggests that the OAFt columns were only high enough to cause gravitational collapse resulting in pyroclastic flows. The pyroclastic flows accumulated rapidly and welded onto one another forming thick cooling units. Degassing of the cooling units resulted in fumarolic activity with alteration and localised mineralisation (mainly fluorite disseminations).

The emplacement of the Erongo Granite took place sometime after the last OAFt eruption. It is postulated that below the present surface the Erongo Granite forms a continuous mass above and around subsided blocks of volcanic and basement rocks. The Erongo Granite is important from the metallogenic viewpoint because it acted both as heat engine and as source for hydrothermal fluids, which were H₂O-poor, but rich in alkalis, B, and F. Widespread metasomatism occurred with local greisen-type mineralisation containing W, Sn, F, and Be, hosted in the roof zones of the granite and in the surrounding country rocks (Pirajno and Schlögl, 1987). The quartz tourmaline orbicules have been interpreted to represent the initial stages of the development of a magmatic hydrothermal system, resulting from the localised separation of a B-rich fluid in the roof zones of the pluton (Sinclair and

Richardson, 1992; Slack, 1996). The Erongo mineralization, discussed below, is thought to have originated from the evolution of this magmatic hydrothermal system.

The rheomorphic rhyolitic rocks (RHEOR), representing the third volcanic event, may be the effusive equivalent of the Erongo Granite. If this interpretation is correct, then the eruption of the RHEOR units must have occurred from a ring fracture outside the central volcanic edifice. This ring fracture, at the present level of exposure is occupied by the Erongo Granite. Collapse of the volcanic edifice at about this time led to the formation of the caldera-like structure and the tilting of the rock units towards the centre of the Complex. The final stage in the Erongo's geological history was the emplacement of the undersaturated mafic plugs that intrude the pyroclastic sequences in the northeastern areas. It is possible that this may represent a resurgence of alkaline mafic magmatism, due to re-activation of deep-seated fractures.

3. Mineralisation in anorogenic alkaline complexes

The mineralising fluids associated with alkaline complexes develop during the late stages of crystallisation and differentiation of the melts. These late-stage melts have an alkaline to peralkaline chemistry. Depending on the permeability of the country rocks (porosity, fractures, faults, shear zones), they are expelled from the cooling magma inducing

metasomatic changes from a few hundred metres to several kilometres away from the complex (Pirajno, 2009).

Hydrothermal alteration related to alkaline complexes is dominated by alkali metasomatism, which in the specific case of alkaline complexes is also known as fenitisation, with the altered rocks being called fenites (the name derives from the Fen carbonatite in Norway; Pirajno, 2013a and references therein). Fenites are formed through progressive metasomatism of the country rocks surrounding the alkaline complexes. In general terms, it can be said that K-metasomatism (K-feldspathisation) is associated with carbonatites and undersaturated magmas, whereas Na-metasomatism (albitisation) is linked with saturated–oversaturated magmas (Pirajno, 2013a).

The origin of mineral deposits of economic significance, therefore, is related to the development, composition and evolution of these late stage fluids, their interaction (and disequilibrium) with the wallrocks, and meteoric waters. Mineral deposit types that are formed in this way can be considered in terms of the nature of the magmas involved. Thus, we may distinguish mineralisation associated with: 1) saturated–oversaturated complexes; and 2) undersaturated complexes. The saturated–oversaturated complexes tend to be enriched in Sn, W, U, Mo, Bi, As, Y, Li, Ta, Zr, La, Ce and Nb (pyrochlore) and the undersaturated complexes in Sr, Ba, Th, Nb, Ti, Fe (magnetite) and P (apatite) (Pirajno, 1994, 2013a). Some of the most common ore minerals, which occur in mineral deposits of alkaline rocks are listed in Table 2.

3.1. Mineralisation associated with the Brandberg alkaline complex

The banded (or layered) arfvedsonite aegirine granite is of economic importance because of zircon, uraniferous and REE-bearing pyrochlore disseminations (von Knorring et al., 1985; von Knorring, 1986). Analyses of grab samples of this material gave values of approximately 1% Zr, 0.3% Nb₂O₆ and 0.2% Y₂O₃. Near the contact with the country rocks pegmatitic phases contain pyrochlore, fluorite and monazite, in addition to zircon–arfvedsonite and stanniferous aegirine (Bowden et al., 1990). Schmitt et al. (2000) reported a variety of peralkaline granite consisting of spotted mesocratic segregations with albite, quartz

Table 2
Common ore minerals of alkaline and carbonatite complexes.

Mineral name	General formula/composition
Allanite	Ca, Al, Ce, Di, La silicate
Apatite	Ca ₅ (F,Cl,PO ₄) ₃ F
Arsenopyrite	FeAsS
Bastnaesite	[(REE)(CO ₃)F]
Britholite	complex Ce, P, Ca silicate–phosphate
Baddeleyite	(Zr,Hf)O ₂
Barite	BaSO ₄
Bismuthinite	Bi ₂ S ₃
Cassiterite	SnO ₂
Chalcocopyrite	CuFeS ₂
Columbite–tantalite	Solid solution series of niobate–columbite– and tantalate–tantalite
Florensite	Ce phosphate
Galena	PbS
Ilmenite	FeTiO ₃
Magnetite	Fe ₃ O ₄
Molybdenite	MoS ₂
Monazite	Ce, La, Dy phosphate
Pyrochlore	Complex niobate, including Ca, Ca, Th, F, Ti
Rutile	TiO ₂
Scheelite–powellite	(Ca,Mo)WO ₄
Sphalerite	ZnS
Thorite	ThSiO ₄
Topaz	Al(F,OH) ₂ AlSiO ₄
Uraninite	(La,Y,Pb,Th)O ₂
Wolframite	(Fe,Mn)WO ₄
Xenotime	YPO ₄
Yttrifluorite	Y-bearing fluorite
Zircon	ZrSiO ₄

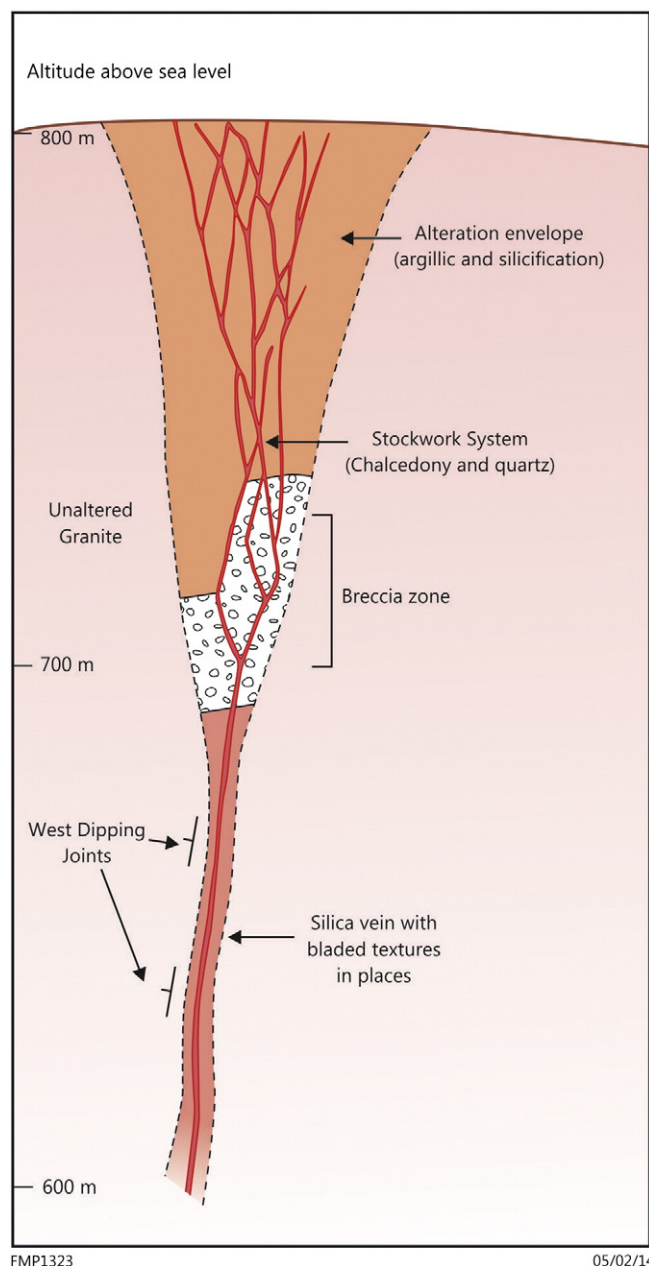


Fig. 7. Epithermal vein system exposed in the Numas Valley, this vein system remains untested for possible REE mineralisation (see text). After Pirajno (1987).

and radiating clusters of aegirine, highly enriched in pyrochlore, thorite, zircon and Nb–Zn-bearing ilmenite.

The mineral layering and mineralization in the Amis intrusion are believed to be related to the influx of metasomatic peralkaline fluids (Bowden et al., 1990).

On the western margin in the Numa Valley of the Brandberg massif, Pirajno (1987) described the occurrence of a west–northwest-trending fracture-controlled hot spring-related vein system (Fig. 4). The vein system has a vertical extent of about 200 m (Fig. 7) and consists of a chalcedonic silica vein associated with bladed textures in the lower parts of the system. Further up the single vein ramifies into a stockwork of quartz and chalcedony with limonitic and hematitic infilling open spaces. Opaline quartz and sulphide disseminations are locally present. The stockwork zone was interpreted as a hydrothermal breccia resulting from the escape of volatiles during boiling. The wall rocks around the stockwork show argillic alteration, locally overprinted by

pervasive silicification. Also, fluidisation textures were noted in the altered wall rocks, indicative of volatile escape during boiling. With regard to this epithermal vein, it is worthy of note to point out that REE mineralisation could be associated with this vein, since fluids could have leached, transported and deposited minerals such as xenotime. This type of epithermal vein system, hosting abundant disseminated xenotime, is found in the northwest of Western Australia (unpublished my own field observations at John Gault and Browns Ranges prospects; see also Cook et al., 2013). Equally worthy of note is that the causative source magmas, likely of alkaline or superalkaline nature, are not exposed at these prospects.

3.2. Mineralisation associated with the Erongo volcano-plutonic complex

The Erongo volcano-plutonic complex contains deposits and occurrences of W–F–Sn, W, Be, Sn, F, and U, all of which are spatially and genetically related to the Erongo Granite. The distribution of the principal deposits is shown in Fig. 6. This mineralisation was described by Pirajno and Schlögl (1987) and Pirajno (1990) and is summarised below.

Be and Sn mineralisation, hosted by a quartz–albite rock containing numerous tourmaline veins and nests, is present in the north of the complex. This mineralisation is associated with quartz–muscovite (greisen) alteration containing minor cassiterite disseminations. Drusy

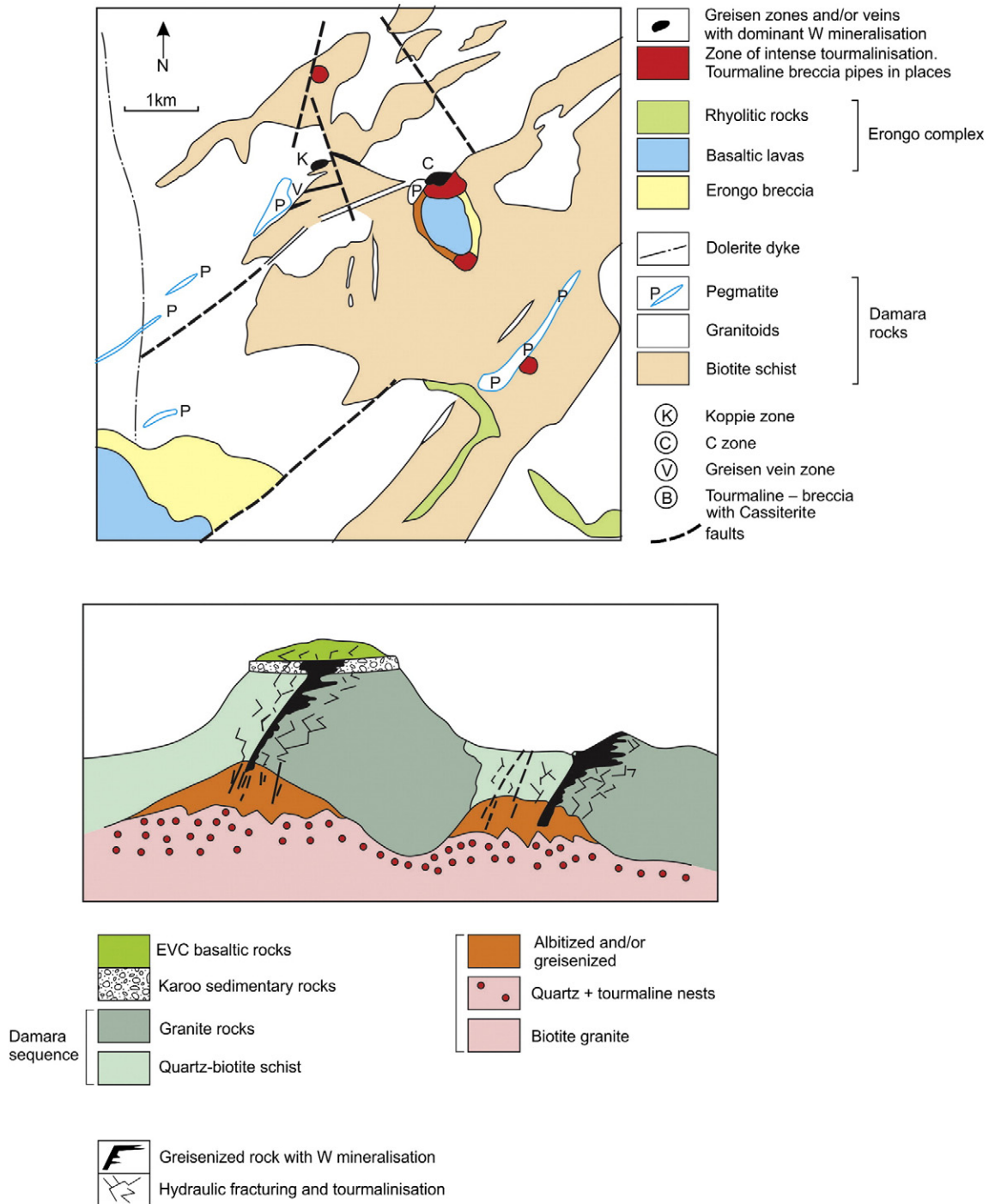


Fig. 8. (Top) Simplified geological map of the Krantzberg area to the northeast of the Erongo complex (see Fig. 6 for location); (bottom) not-to-scale idealised cross-section of the Krantzberg area, showing geometry and position of the mineralised zones. After Pirajno and Schlögl (1987).

pegmatitic pods with crystals of beryl and tourmaline are also present. The quartz–albite rock is probably an alteration zone in a high level (cupola) facies of the Erongo Granite. Rocks around the mineralised locality are brecciated and are cut by veins of tourmaline, attesting to hydraulic fracturing induced by B-rich fluids. Fractures in basement granitoids show intense hydraulic breccias, tourmalinisation, albitic alteration and localised cassiterite, wolframite and fluorite mineralisation.

Uranium mineralisation is present in the Erongo Granite, particularly in the northwest and southwestern areas. Secondary U minerals are associated with jointing in areas of primary U enrichment. It is probable that this mineralisation is due to meteoric waters that have leached U from the fresh Erongo Granite rock in areas of high permeability due to close-spaced jointing.

The old Krantzberg W–F–Sn deposit was a major tungsten producer in Namibia. Mining operations ceased in 1979 and subsequent exploration did not define economically feasible grades and tonnages. Mineralisation at Krantzberg consists of W (ferberite), F (fluorite), minor Sn (cassiterite), Mo, Fe, and Cu sulphides. This mineralisation is hosted in replacement-type greisen rocks and quartz–tourmaline breccias. The style of the Krantzberg deposit is shown in Fig. 8. The ‘Krantzberg’ is a prominent hill of rocks of the Swakop Group (Damara Orogen) that was protected from erosion by an overlying cap of clastic sedimentary rocks and an outlier of Erongo’s basaltic rocks (Figs. 6 and 8). There are three main zones of alteration–mineralisation, all formed by the pervasive greisenisation of Neoproterozoic Damaran granites. The Krantzberg greisen alteration is interpreted to be the result of acid fluids that emanated from the subsurface apical portions of

fractionated Erongo Granite cupolas. The mineralising fluids were also B- and F-rich and induced extensive selective replacement of pre-existing granite lithologies by quartz, sericite, topaz, and tourmaline. Late-stage veins contain ore minerals such as ferberite, scheelite, and minor sulphides. A tentative mineral paragenesis for the Krantzberg W deposit is shown in Fig. 9. Breccia pipes on the southeastern slopes of the Krantzberg contain fragments of volcanic rocks cemented by quartz and tourmaline. Cassiterite occurs as fine disseminations or as patches along the periphery of the breccia bodies. These breccias are interpreted as a type of hydraulic fracturing due to streaming of B-rich fluids.

An idealised cross-sectional view of the Erongo complex with associated hydrothermal occurrences is shown in Fig. 10.

4. Carbonatites

The increasing importance of critical and rare earth elements in modern technology (e.g. solar panels, wind turbines, fibre optics, television screens, light bulbs), makes it appropriate to briefly examine how magmas that commonly contain these elements, namely carbonatites, form. The origin of carbonatite magma is controversial, and at the heart of the controversy is the relationship between carbonatites and the silica-undersaturated rocks with which they are commonly associated. An understanding of how carbonatites form and their association with REE is a primary concern in order to target new deposits, improve exploration methods and the re-evaluation of

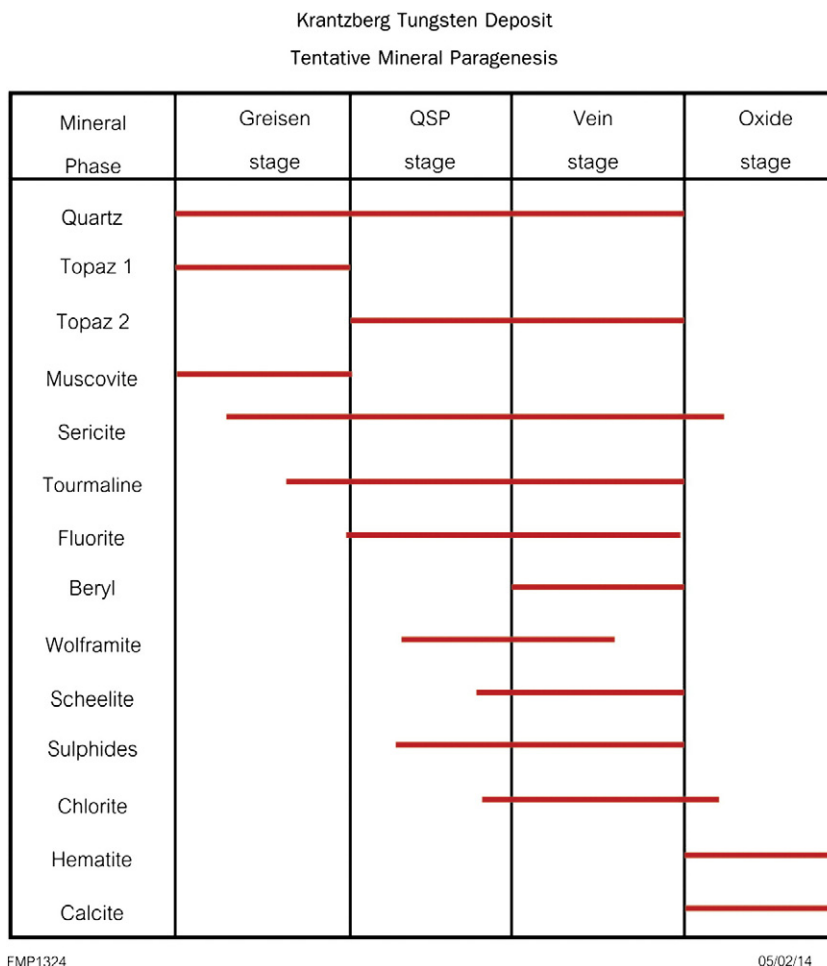


Fig. 9. Tentative mineral paragenesis at Krantzberg. After Pirajno and Schlögl (1987).

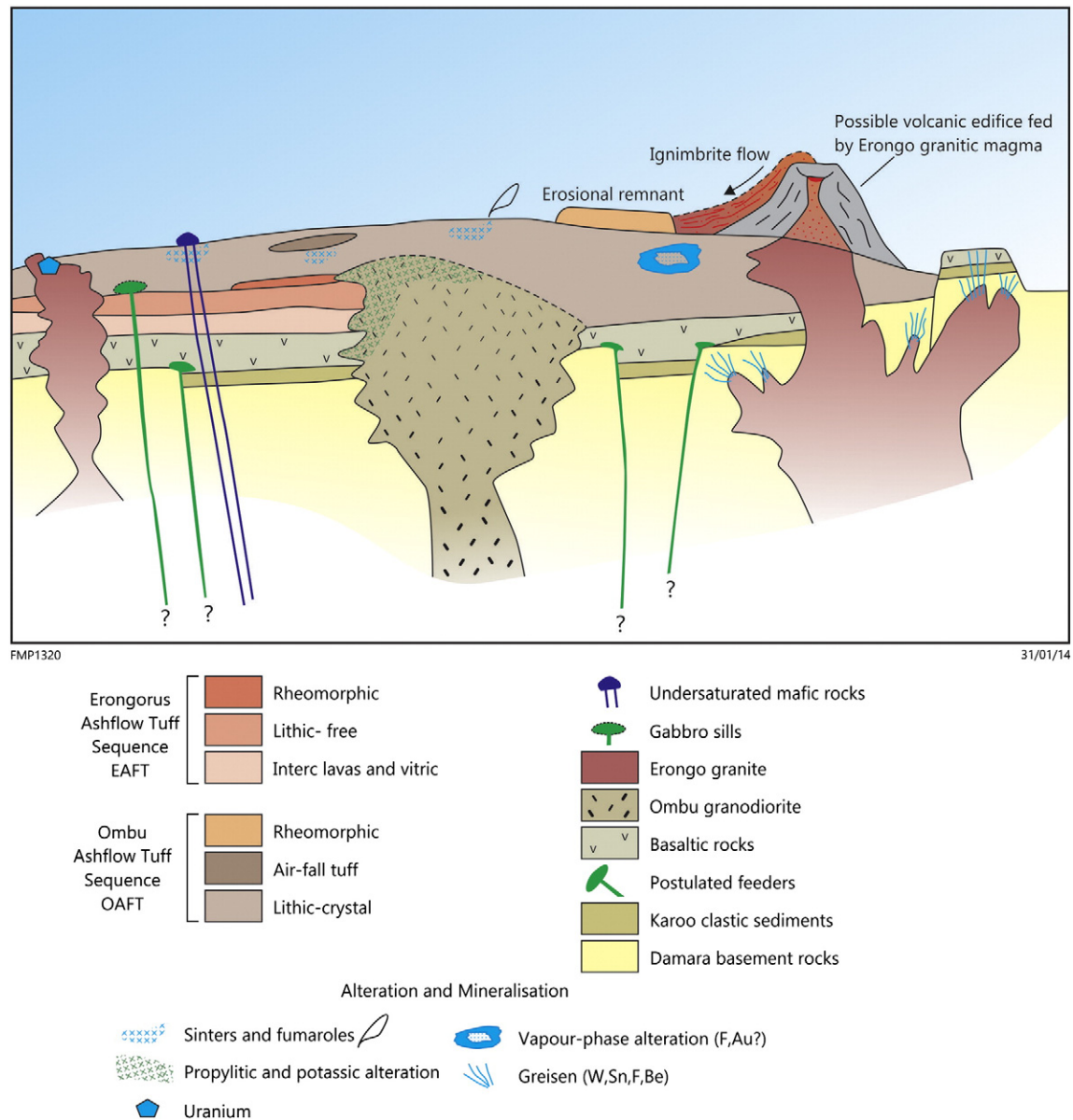


Fig. 10. Idealised cross-section (not to scale) of the Erongo volcano-plutonic complex and associated hydrothermal systems.

sub-economic deposits. Theories that attempt to explain the origin of carbonatite are summarised below:

- Carbonatites evolve via crystal fractionation, at crustal pressures, of a mantle-derived alkali-rich silicate melt (Watkinson and Wyllie, 1971). Silica-undersaturated extrusive mafic rocks are considered to represent primary mantle-derived liquids and their formation at mantle depths requires significant amounts of CO₂. These silica-undersaturated liquids are thought to be the parental magmas for ultramafic–alkaline–carbonatite complexes. Extensive fractionation would lead to the crystallisation of silicates, while carbonates (calcite, dolomite) would crystallise from the residual liquids after removal of alkalis by fenitisation processes. The origin of carbonatite liquids by partial melting of a carbonated lithospheric mantle or mantle source is supported by experimental data, which indicate that carbonate melts can be generated from carbonated peridotites at depths greater than 70 km (Gudfinnsson and Presnall, 2005; Veksler and Lentz, 2006), with near-solidus liquids dominated by calcite and magnesite (dolomitic) and with the concentration of other components being related to the bulk composition of the mantle source.
- Carbonatites form from contact metasomatism between carbonate wall rocks and intrusions, resulting in the development of skarns, release of CO₂, decarbonisation and flux melting of crustal carbonates (Lentz, 1999).
- Carbonatites are derived via immiscible separation of a liquid from an originally homogeneous carbonated silicate melt, the stage of separation determining the composition of both magmas (e.g. Le Bas, 1989). Silicate–carbonate liquid immiscibility may occur from peralkaline nephelinitic magmas. This liquid immiscibility has been investigated in several experiments, although it has to be cautioned that experimental systems may not be totally applicable to natural magmatic systems. Nevertheless, evidence of liquid immiscibility is provided by textural evidence detected under the microscope (see below) and in melt inclusions, furnishing adequate evidence that carbonatite genesis may be linked to silicate–carbonate unmixing (Veksler and Lentz, 2006).
- A two-stage process, as envisaged by Nielsen and Veksler (2002), whereby Na-rich minerals are formed within a nephelinitic–phonolitic volcanic plumbing system by the activity of magmatic–hydrothermal fluids. Subsequently the consolidated fluids are re-

melted and re-mobilised by the heat energy provided by a following pulse of ascending silicate magma. This model links the origin of natrocarbonatite with fenitising fluids, which would exsolve from carbonatitic liquids.

In most cases, the recognition of true carbonatites can be ascertained by the presence of key silicate minerals, such as phlogopite, acmite–diopside, forsteritic olivine, nepheline and monticellite, or accessory phases such as apatite, pyrochlore, and perovskite, in addition to dominant carbonate minerals (ankerite, calcite, dolomite). Igneous series that are generally associated with carbonatites include (Woolley and Kempe, 1989; Woolley, 2003): 1) melilite–nepheline–phonolite–trachyte; 2) nepheline–phonolite–trachyte (ijolite); 3) basanite–trachyte–alkali gabbro; 4) phonolite–trachyte–nepheline, sodalite, and cancrinite syenites; 5) trachyte and syenite; and 6) kimberlite.

In this section, I discuss the Kruidfontein and Goudini carbonatites (South Africa), Mount Weld and Gifford Creek carbonatites (Western Australia), Bayan Obo (Inner Mongolia, China) and the Mianing–Dechang carbonatite belt in southwest China. The above covers a significant range of carbonatite systems and associated mineralisation.

4.1. Kruidfontein and Goudini (South Africa)

The Neoproterozoic Kruidfontein and Goudini carbonatite complexes are well-preserved volcanic structures, which are part of several alkaline ring-type systems and kimberlites that are located in and around the Bushveld Igneous Complex in the Limpopo province

(formerly Transvaal) of northern South Africa (Fig. 11). More specifically, the Kruidfontein and Goudini are included in the 1.4–1.2 Ga Pilanesberg Alkaline Province (Verwoerd, 2006).

The Kruidfontein Carbonatite Complex (KCC) is a caldera-type structure of approximately 5 km diameter. The Sm–Nd age of the complex is 1246 ± 26 Ma (Verwoerd, 2006). A simplified geological map of the KCC is shown in Fig. 12. The volcanic history of the complex consists of two main eruptive episodes: An earlier episode dominated by undersaturated silicate lavas and pyroclastics, and a later episode characterised by carbonatitic volcanism. Thus, the KCC consists of an outer ring, approximately defining the caldera structure, containing variably altered and metasomatised silicate lavas and pyroclastics that were probably erupted from a central vent. Phonolitic and nephelinitic rocks are recognised within this outer zone. An inner ring contains carbonatitic pyroclastic products, including bedded tuffs, ash-flow tuffs and vent breccias. The bedded tuffs are interpreted as air-fall, water-lain and surge deposits (Clarke, 1989; Clarke and Le Bas, 1990; Pirajno et al., 1995). Carbonatite plugs and dykes intrude the central portions of the inner ring. Clarke (1989) classified the KCC carbonatites into three types: sövite (coarse-grained), alvikite (medium- to fine-grained) and fluorite-rich carbonatite.

The fresh KCC carbonatites, as seen in drillcores, are generally fine-grained, grey to brown in colour, and composed of ferroan calcite (approx. 60–70%), fluorite (approx. 10–30%), hematite, pyrite and chalcopryrite (approx. 5–10%) (Clarke, 1989; Pirajno et al., 1995). A selection of key features of the Kruidfontein carbonatite rocks is shown in Fig. 13 and photomicrographs of selected thin sections in

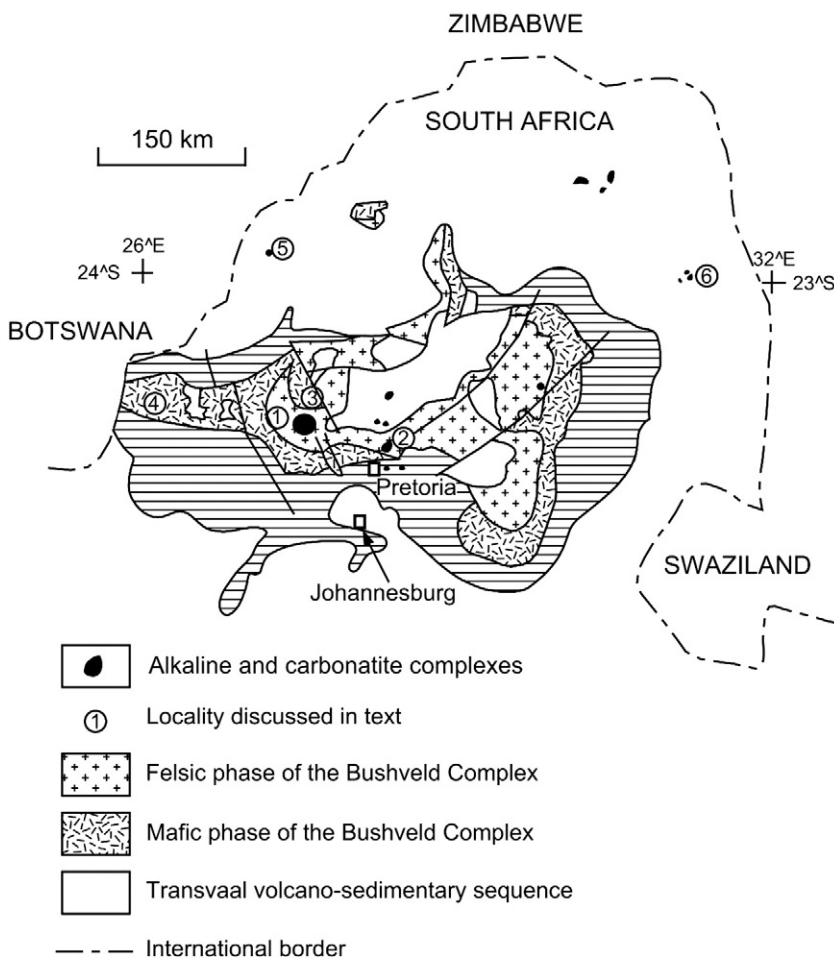


Fig. 11. Distribution of alkaline ring-complexes and carbonatites in the Limpopo province (former northern Transvaal), South Africa; numbered localities, 1 Pilanesberg, 2 Pienars River, 3 Kruidfontein, 4 Goudini, 5 Glenover, 6 Palabora. After Pirajno (2009) and references therein.

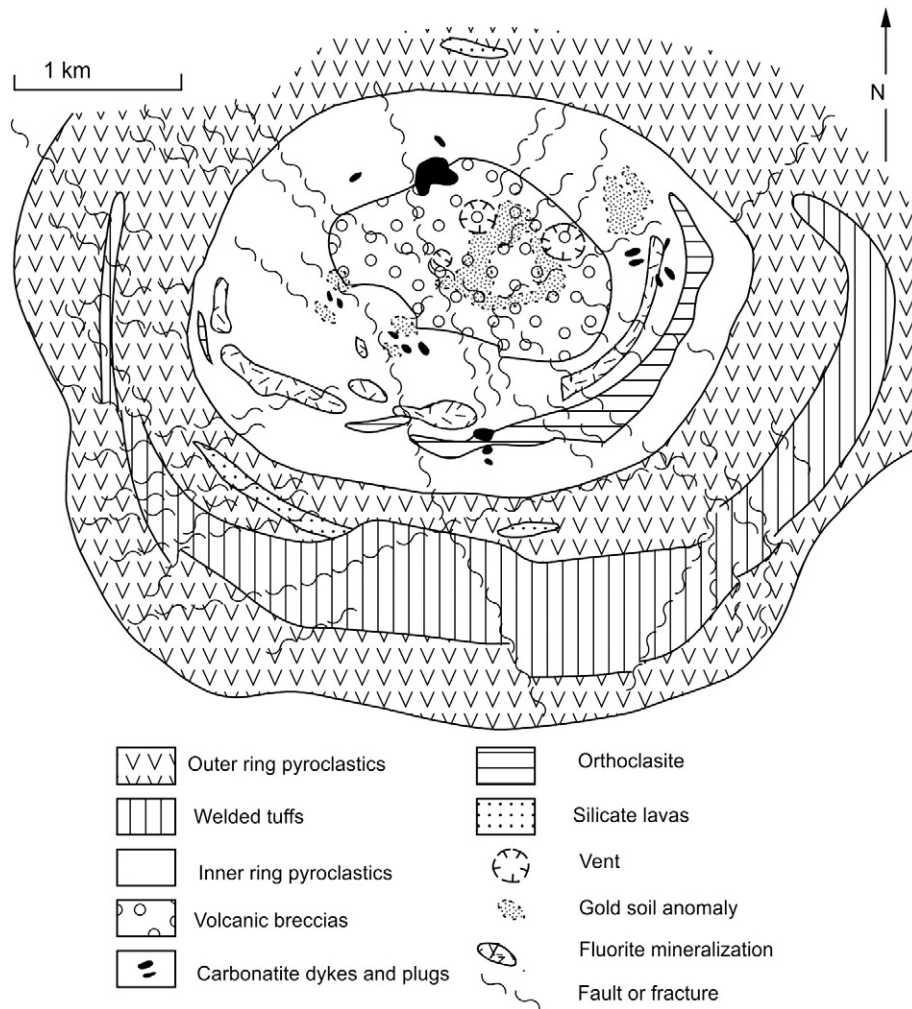


Fig. 12. Kruidfontein geology map; simplified geological map of the Kruidfontein Carbonatite Complex. Modified after Clarke, 1989 and shown in Pirajno et al. (1995).

Fig. 14. An important aspect of the KCC is the presence of an almost monomineralic K-feldspar-rich rock (orthoclase of Clarke, 1989). This is interpreted as the product of K-metasomatism of pre-existing silicate rocks, which resulted in the development of the characteristically brick-red, orthoclase and/or microcline rock.

The Goudini Carbonatite Complex (GCC) (Fig. 15), extensively studied by Verwoerd (1993, 2006, 2008), has a Sm–Nd age of 1190 ± 80 Ma (Nelson et al., 1988) and consists of a maar diatreme surrounded by volcanic breccia and volcanoclastic sediments infilling the palaeocrater. The latter comprise rhythmically layered black tuffite, intercalated with layers of carbonatitic ash-fall which, as Verwoerd (2008) suggested, was deposited contemporaneously with the volcanoclastics. The ash-fall beds were later altered to a form of calcrete that was earlier assumed to be limestone. To the northwest of the maar is what is referred to as metabasite, possibly a metasomatic replacement of accretionary lapilli tuff (Verwoerd, 2008). Amygdaloidal green lava occurs in the northern part of the GCC, which is characterised by spheroids as well as amygdaloids filled with calcite (Fig. 16). I concur with Verwoerd's conclusion that the calcite spheroids result from liquid immiscibility, a feature apparently common with carbonatite–silicate melts, as also dramatically shown by several phases of immiscible carbonate–silicate liquids in the Gifford Creek ferrocarbonatite complex, discussed below. This green lava also contains acicular aegirine, nepheline, magnetite and accessory apatite and was interpreted by Verwoerd (2008) as a phonolitic nephelinite. Northwest-trending sövite dykes cut through the central part of the complex. A typical feature of Goudini is the presence of strongly fenitised

norite country rocks, studied in detail by Verwoerd (1966). This norite (mainly Ca-plagioclase + orthopyroxene), a chilled floor phase of the Bushveld Igneous Complex (Eales and Cawthorn, 1996), was affected in the vicinity of the GCC by metasomatic alteration (fenitisation), mainly along fractures. These fractures were filled with sodic amphiboles, carbonates (mostly dolomite) and chlorite. Closer to the carbonatite the norite is strongly brecciated and its groundmass is altered to green aegirine, cancrinite and calcite (Fig. 16). Apatite is also present as clusters or disseminations.

4.2. Mount Weld and Gifford Creek (Western Australia)

Mount Weld is a 2025 ± 10 Ma (Re–Os isochron age; Graham et al., 2004) carbonatite plug in the Eastern Goldfield Province of the Yilgarn Craton. Mount Weld is part of a larger regional alkaline province characterised by mafic–ultramafic rocks, kimberlites and other carbonatites (Duncan and Willett, 1990; Graham et al., 2004; Hoatson et al., 2011). The primary carbonatite plug or pipe is about 3 to 4 km in diameter and dominantly comprises sövite, dolomitic sövite and beforite. The plug has a nearly cylindrical shape with an approximate 0.5 km wide zone of fenitic alteration. A lateritic regolith blanket from 10 to 70 m thick overlies the carbonatite plug and constitutes the supergene apatite-dominant ore (Fig. 17 and see below). It was initially reported that the plug is surrounded by a ring of glimmerite (ferriphlogopite). However, recent work (Pirajno, unpublished data)

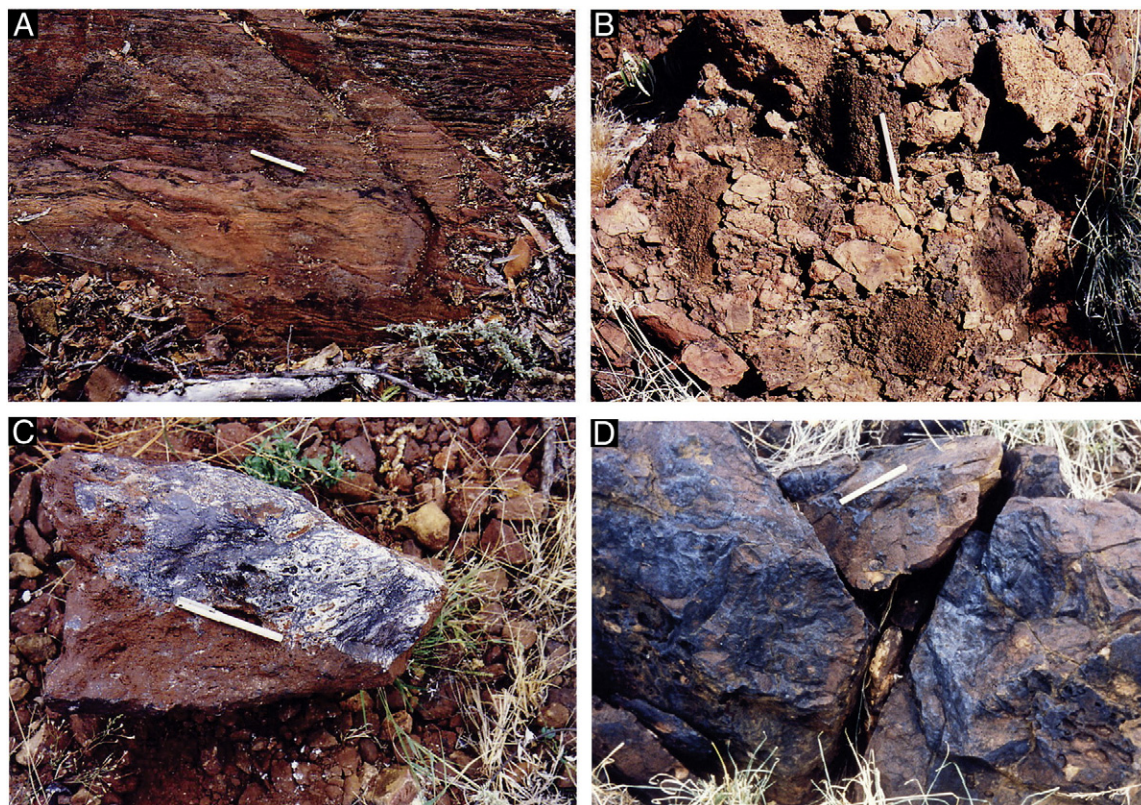


Fig. 13. Kruidfontein photos of outcrops; A) bedded tuff; B) carbonatite breccia; C) sinterous material; D) ferruginous gold-bearing rock.

based on field observations, petrographic and XRD and EDX analyses has revealed a somewhat different story, as elaborated here.

Mount Weld is probably a maar-type diatreme, formed by processes of phreatomagmatic activity, comprising a carbonatite–glimmerite breccia, agglutinated pyroclastic ash and a vent facies. As mentioned above, the primary carbonatite consists mostly of sövite and dolomitic sövite, commonly exhibiting distinct granular to globular textures, interpreted as a fluidised vent-facies. Various amounts of disseminated magnetite, ferriphlogopite, blue magnesio-arfvedsonite, accessory fluoroapatite, pyrochlore, churchite and monazite are present in the carbonatite rocks. Based on textural relationships a tentative paragenetic sequence is as follows: calcite I → pyrochlore → ferriphlogopite → calcite II → calcite + arfvedsonite → pyrite–magnetite. The carbonatite–glimmerite breccia material is likely a remnant of hydroclastic to pyroclastic fragmentation due to explosive interaction of a two-phase system containing juvenile fragments and xenoliths of country rocks. Amongst these fragments are xenoliths of cumulus apatite, which may represent an early liquidus that crystallised prior to the emplacement of the main carbonatite. Fenitic alteration consists of ferriphlogopite (glimmerite) occurring as irregular patches, potassian magnesio-arfvedsonite (detected by EDX analysis), typically with a flow-banded texture and locally replaced by sepioclhorite sheafs and aenigmatite crystals. There are several generations of veinlets, fracture fillings of sövite, arfvedsonite, ferriphlogopite and local disseminations of pyrite and chalcopyrite. Many of the fragments show plastic-like deformation, possibly resulting from an agglutinated carbonatite pyroclastic. A selection of photomicrographs from thin sections of core samples is shown in Fig. 18. Much work remains to be done in order to unravel the sub-volcanic and evolutionary history of the Mount Weld maar. At the time of writing work is still in progress.

The Gifford Creek Ferrocarbonatite Complex (GFC) is located within the Gascoyne Province of the Capricorn Orogen in Western Australia (Martin et al., 2006; Sheppard et al., 2007, 2010), between the Archean Yilgarn and Pilbara cratons (Fig. 19). The Province consists of a suite of

Neoarchaean to Palaeoproterozoic granite gneisses and metasedimentary rocks (Johnson et al., 2011a,b) that are interpreted as basement to various Proterozoic rocks, including granitic rocks of the 1680–1620 Ma Durlacher Supersuite and the metasedimentary rocks of the 1760–1680 Ma Pooranoo Metamorphics.

Rocks of the GFC were first studied and reported by Pearson (1996), Pearson et al. (1996) and Pearson and Taylor (1996). The Gifford Creek rocks comprise ferrocarbonatite dykes, veins and sills that are surrounded by variably fenitised country rocks. Due to the mineralogical preponderance of Fe-rich carbonates and Fe-rich silicates, in some cases almost exclusively, Pirajno and González-Álvarez (2013) renamed it Gifford Creek Ferrocarbonatite Complex (GFC). One of the main features of the GFC is the widespread occurrence of ironstone (mainly magnetite, hematite and goethite) veins that are spatially associated with the ferrocarbonatite intrusions. These are surrounded by relatively narrow haloes of fenitic alteration interpreted to have formed during post-magmatic alteration processes. Age determinations from the ferrocarbonatite and fenitic alteration rocks (see Table 1; Pirajno et al., 2014) fit well with the age range of the Warakurna large igneous province (Wingate et al., 2004). A link with a mantle plume that may have generated the Warakurna LIP is considered possible, as discussed in Section 6.

The ferrocarbonatite sills and dykes form a NW-trending belt in the south-western sector of the GFC, parallel to, and to the north of the Lyons River Fault (Fig. 19). On the northern side of this NW-trending belt are swarms of carbonatite and ironstone veins, distributed in a series of sinuous trends, terminating along a WNW-trending belt of carbonatite–ironstone veins. This WNW-trending belt is along a poorly defined lineament, which Pearson et al. (1996) called the Bald Hill Lineament (Fig. 19).

Petrographic, XRD and SEM investigations (Pirajno et al., 2014), show that the ferrocarbonatites are dominantly composed of ankerite–dolomite, magnetite, Fe–Ti oxides, arfvedsonite–riebeckite, with variable proportions of perovskite, pyrochlore, columbite,

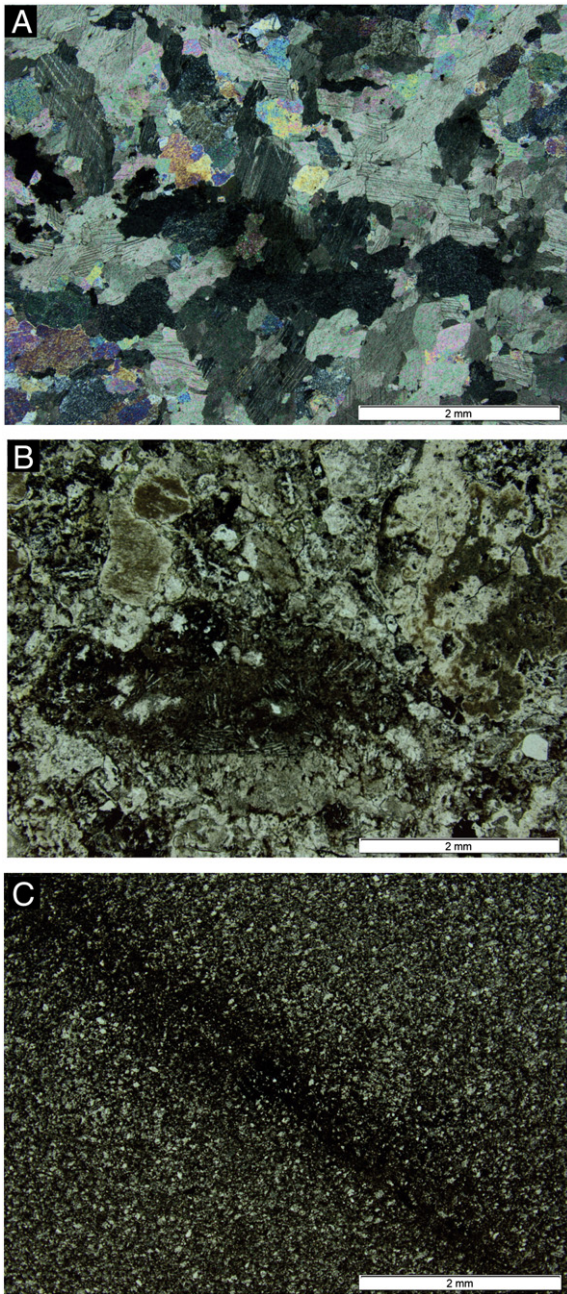


Fig. 14. Kruidfontein photomicrographs; A) calcite grains of primary carbonatite; B) volcanic breccia, partially replaced by carbonate; C) fine-grained carbonate assemblage with cross-cutting veinlets of Fe oxides, probably the initial ferrous alteration.

phlogopite and lesser calcite, and apatite, barite and monazite. The WNW-trending sills and dykes north of the Lyons River Fault are characterised by a carbonate-rich matrix, containing fluoroapatite, celestite, bastnaesite, phlogopite, barite and monazite. Field and petrographic work shows that these rocks are generally ankeritic carbonatites (>50 vol.% ankerite–dolomite), and locally silicocarbonatite (arfvedsonite-rich). In one sample of ferrocarbonatite, millimetre scale banding exhibits features that are interpreted as due to a quench-textured silicate–carbonate immiscibility (Fig. 20). Immiscibility of two liquids, carbonate–silicate, has been experimentally investigated and observed in a number of alkaline complexes in East Africa and Italy (e.g. Kjarsgaard and Peterson, 1991; Solovova et al., 2005; Stoppa et al., 2005).

The fenitic haloes in the country rocks surrounding the carbonatites are mostly characterised by the presence of feldspars and/or alkali-amphiboles and magnetite. Alkali amphibole has compositions ranging

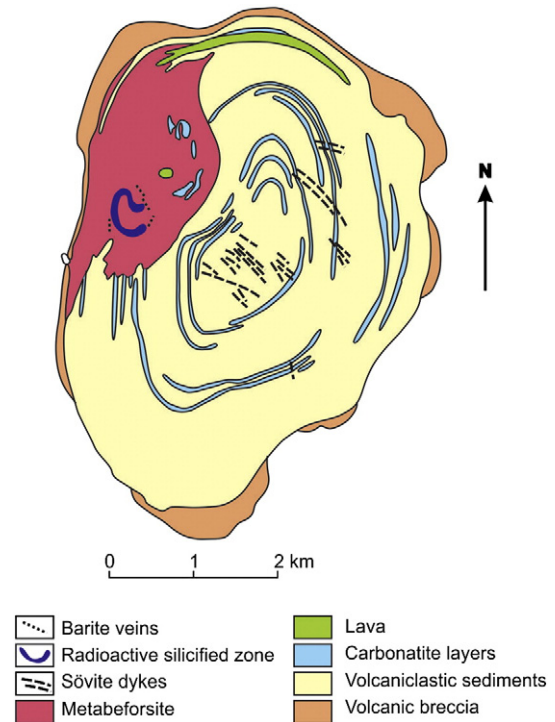


Fig. 15. Simplified geology of the Goudini complex. Modified after Verwoerd (2008).

from potassian magnesio-arfvedsonite to magnesio-riebeckite. The amphiboles in the fenitic rocks are classified as potassium magnesio-fluor-arfvedsonite (Pearson and Taylor, 1996). Sodic pyroxene is generally aegirine with a high $\text{NaFe}^{3+}\text{Si}_2\text{O}_6$ component and with TiO_2 contents of up to 4 wt.%. Mica is metasomatic and is Al-deficient fluorphlogopite (F contents from 1.5 to 4.5 wt.%; Pearson and Taylor, 1996).

The above mentioned minerals are the main components of fenitic alteration and occur in varying amounts, presumably due to, not only the composition of the alkaline fluids, but also to the primary mineralogy of the rocks affected. This results in a great variety of textures and dominant mineralogy. Furthermore, petrological analysis shows that the fenitisation process was highly dynamic and multiphase. For example, there are at least two generations of aegirine–augite/aegirine: coarse grained, and very fine-grained granular aggregates, overprinted by acicular riebeckite (Fig. 21A). In turn, acicular riebeckite is overprinted by phlogopite mica. In the same rock isolated euhedral pyrochlore (Fig. 21B), slightly pleochroic lamprophyllite crystals and stellate aggregates of green-blue riebeckite may also occur. Fluidal textures are common in these fenites and generally consist of bands or laminae of alternating fine-grained and closely packed aggregates of K-feldspar and plagioclase (oligoclase–albite compositions) and of pale green biotite. The feldspar bands exhibit a granular to fluidal-like texture in places (Fig. 21C). In other cases, assemblages of euhedral nepheline, forming linear granular aggregates, occur within masses or polycrystalline aggregates of biotite, with distinct green pleochroism and abundant acicular riebeckite (Pirajno et al., 2014) (Fig. 21D).

The emplacement of the GFC intrusions occurred along the Lyons River Fault, a major trans-crustal structure, which a seismic transect shows that it reaches into the subcontinental lithospheric mantle (Johnson et al., 2013). Movements along the Lyons River Fault could have formed a pull-apart basin (see below) into which the carbonatite melts migrated possibly from the distal sector of the mantle plume responsible for the formation of the Warakurna LIP.

The apatite and monazite ages (Table 1) suggest the possibility of two separate events (Fig. 22). The first at ~1075 Ma was responsible for the emplacement of dykes and sills along the Lyons River fault for

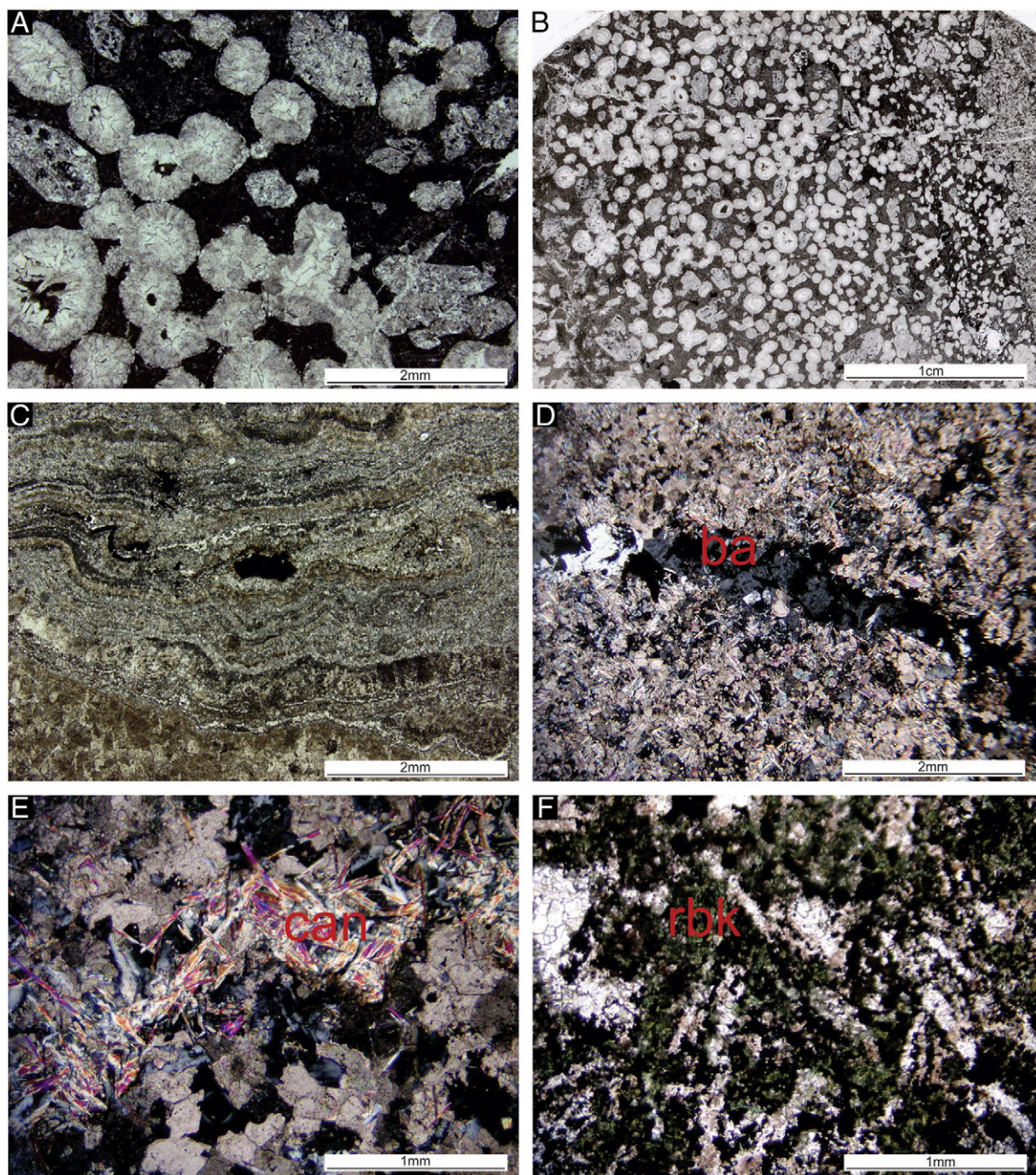


Fig. 16. Goudini photomicrographs; A) and B) carbonate spherules in nephelinite, reflecting a good example of silicate–carbonate liquid immiscibility (see text); C) flow-banded carbonatite; D) barite vein (ba) in carbonatite; E) cancrinite (can) overprinting calcite; F) fragment of volcanic rocks with matrix replaced by riebeckite (rbk); A and B plane polarised light; C, D, E and F crossed polars.

which, as mentioned above, seismic profiling indicate that it cuts through the crust and reaching into the sub-continental lithospheric mantle (Johnson et al., 2012). Carbonatite melts will have formed in a metasomatised lithospheric mantle, perhaps due to the heat induced by the lateral flow of the Warakurna mantle plume (Pirajno and Hoatson, 2012), resulting in the emplacement of the NW-trending ferrocarbonatites dykes and sills (Fig. 22A). The Lyons River fault system was activated time and again during tectonic events that affected the Capricorn Orogen. It was suggested in Pirajno et al. (2014) that in one of these events, at about 1050 Ma, a small pull-apart structure was formed on the sites where the ferrocarbonatites had been intruded. This stimulated the re-activation of the carbonatite system, widening the fenitic halo in the country rocks and producing a sinuous carbonatite

vein system which eventually was locally altered to the ironstone veins (Fig. 22B).

4.3. Bayan Obo (Inner Mongolia, China)

Bayan Obo, located on the northern margin of the North China Craton (NCC), is an extremely complex, hydrothermal, stratabound mineral system hosted in carbonate rocks of the Mesoproterozoic Bayan Obo Group (Fig. 23) (Pirajno, 2013b and references therein). This Group is an ~1850 m thick package of clastic and carbonate rocks deposited in graben structures that formed in the Archaean basement. The Bayan Obo Group is deformed as a series of northwest-dipping thrust slices and duplex structures. Granitic rocks of Carboniferous–

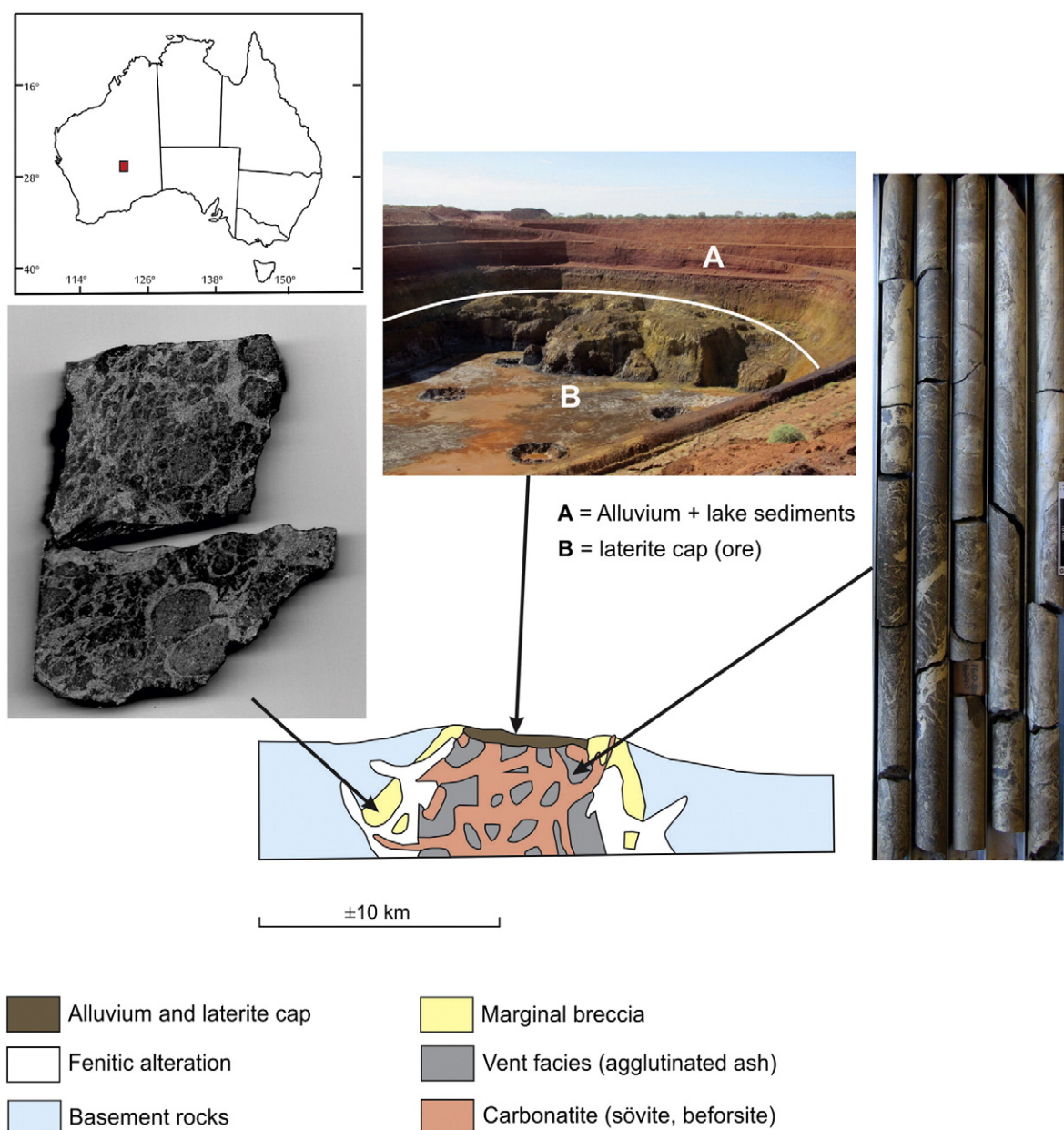


Fig. 17. Interpreted and idealised cross-section of the Mount Weld carbonatite, showing examples of glimmerite-bearing marginal breccia, ore-bearing lateritic cap and glimmerite bearing vent facies rocks.

Permian age intruded the basement and the sedimentary rocks. Other igneous rocks in the area include at least two suites of carbonatite dykes (Table 1), with fenitised (Na-amphibole–albite–phlogopite) country rocks.

The Bayan Obo Group consists of a number of sedimentary units, which consists of a basal conglomerate and sandstone (Y1 unit), overlain by a succession of carbonaceous shale, arkosic sandstone and quartzite (Y2 to Y7 units), followed by limestone and dolomite beds (Y8 unit, host of the REE mineralisation), black shale (Y9), and capped by white and grey limestone (Y14 to Y18 units). This shale is extensively brecciated and altered by potassic metasomatism in the core of the syncline where the REE orebodies are located. The structural setting of the Bayan Obo area is very complex and this complexity has hindered a clearer understanding of the Bayan Obo mineral system, which has been debated since its discovery, resulting in more than 100 publications. The complex deformation is the result of collisional processes that affected the northern margin of the North China Craton, with tectonic transport from northwest to southeast (Drew et al, 1990).

Drew et al. (1990) made the interesting suggestion that the entire deformed package is now a giant roof pendant (6–8 km across) in the Carboniferous–Permian granitic rocks, which were emplaced during the collision event.

The Bayan Obo carbonatite contains, apart from dolomite, benstonite $[(Ca,Sr)_6Ba_6Mg(CO_3)_{13}]$, norsethite $[BaMg(CO_3)_2]$, barytocalcite $(BaCa(CO_3)_2)$ and strontianite $(SrCO_3)$. Sulphide gangue minerals are pyrite, pyrrhotite, sphalerite, galena and chalcocopyrite. In addition, more than 100 carbonatite dykes have been recognised in the general area (Fan et al., 2014) (Fig. 23).

The age of the Bayan Obo mineralisation is also controversial and uncertain. The K–Ar and $^{39}Ar/^{40}Ar$ isotopic systems provide a range of ages from 1800–1290 Ma, 900–800 Ma, 555–474 Ma and 430–420 Ma (Chao et al., 1997; Yang et al., 2014) to 311–306 Ma for early stage granites (mainly granodiorite) and 258–234 Ma for late stage biotite K-feldspar granites (Fan et al., 2004a and references therein). More precise age data, ranging from whole-rock to individual minerals, given in Table 1, add to the ongoing uncertainties. Chao et al. (1997)

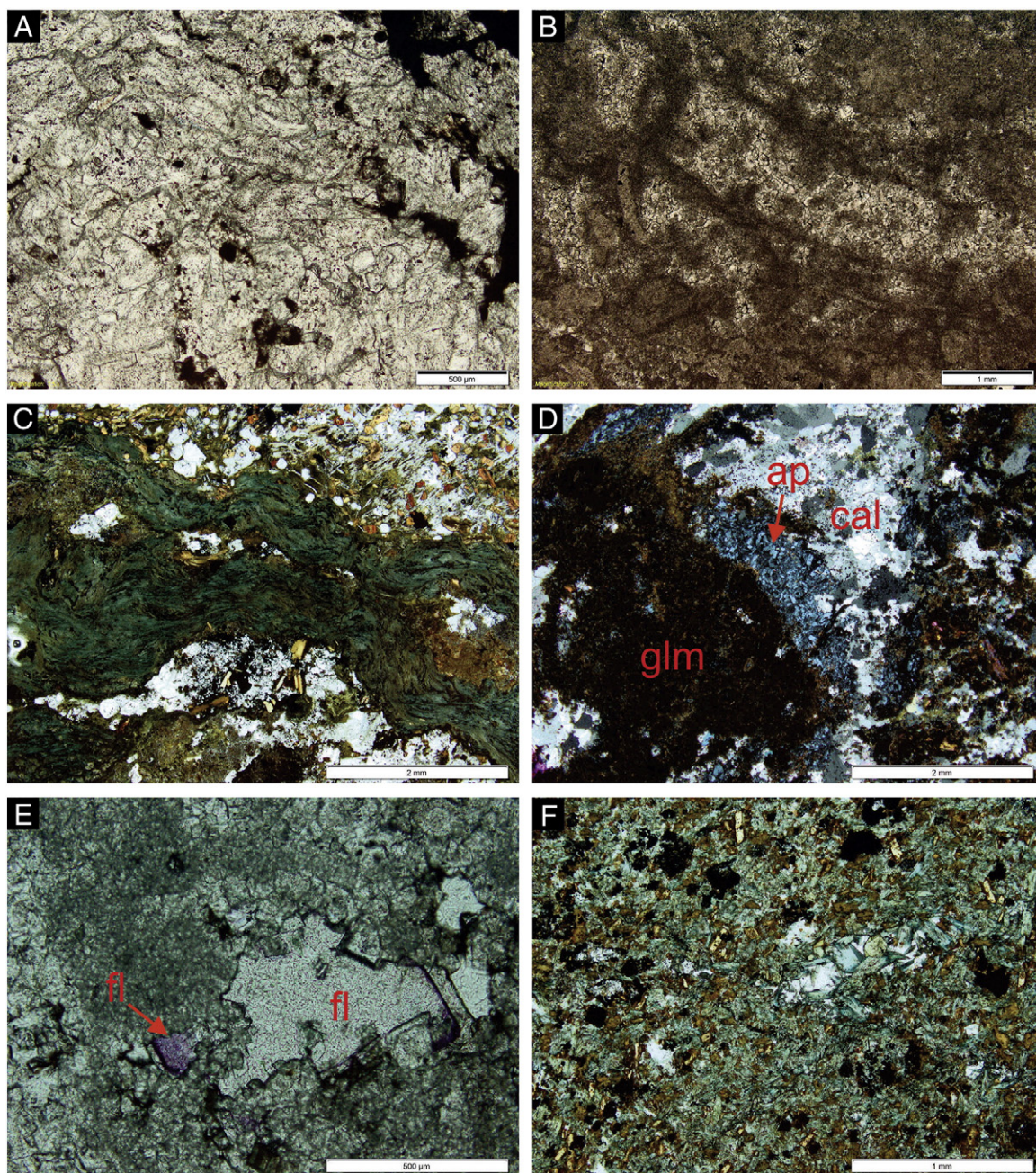


Fig. 18. Mt Weld photomicrographs of A) and B) shard-like texture in carbonatite pyroclastic; C) magnesio-arfvedsonite with fluidal texture; D) fragment of cumulus apatite (ap), calcite (cal) and glimmerite (glm); E) fluorite (fl) vug infill in carbonate aggregate; F) intergranular assemblage of phlogopite (brown grains), carbonate (light grey) and arfvedsonite (bluish-white in central part); A, B, C, E and F in plane polarised light, D in polarised light.

distinguished three periods of regional metamorphism and metasomatism: 1) 890 Ma, with recrystallisation of the Y8 carbonate to marble; 2) 425–395 which would have formed metasomatic–metamorphic sodic amphibole; and 3) Hercynian event at ca 300 Ma. Dating of monazite, using ^{232}Th – ^{208}Pb chronology, yielded ages ranging from 555 to 420 Ma, with peaks at 430–420 Ma (Wang et al., 2004). Given the wide range of ages it is likely that the Bayan Obo mineral system is polychronous and polygenic, that is the product of multiple tectono-thermal events, beginning at ca. 1800 Ma, successively overprinting later ones. Moreover, the variability of the age data led Yang et al. (2011b) to suggest that different types of carbonatites were emplaced sequentially, accounting for the large errors in the age determinations.

Fan et al. (2004a,b) working on fluid inclusions recognised three types: two-phase aqueous liquid–vapour, two to three-phase, almost pure, CO_2 and three-phase liquid–vapour–solid. In the latter, daughter minerals are REE-carbonates (cebaite and bastnaesite), halite, sylvite, barite and calcite, identified by SEM and EDX analyses. Homogenisation temperatures of 480–420 °C and 250–220 °C were obtained from heating–cooling experiments on the L–V–S inclusion fluids. Salinities ranging from 1 to 15 wt.% NaCl equivalent were obtained by Smith and Henderson (2000).

Thus, the carbonatite model seems feasible and could be considered within the framework of intraplate and rift-related A-type magmatism (Pirajno, 2000, 2013b). Considering the regional context, the Bayan

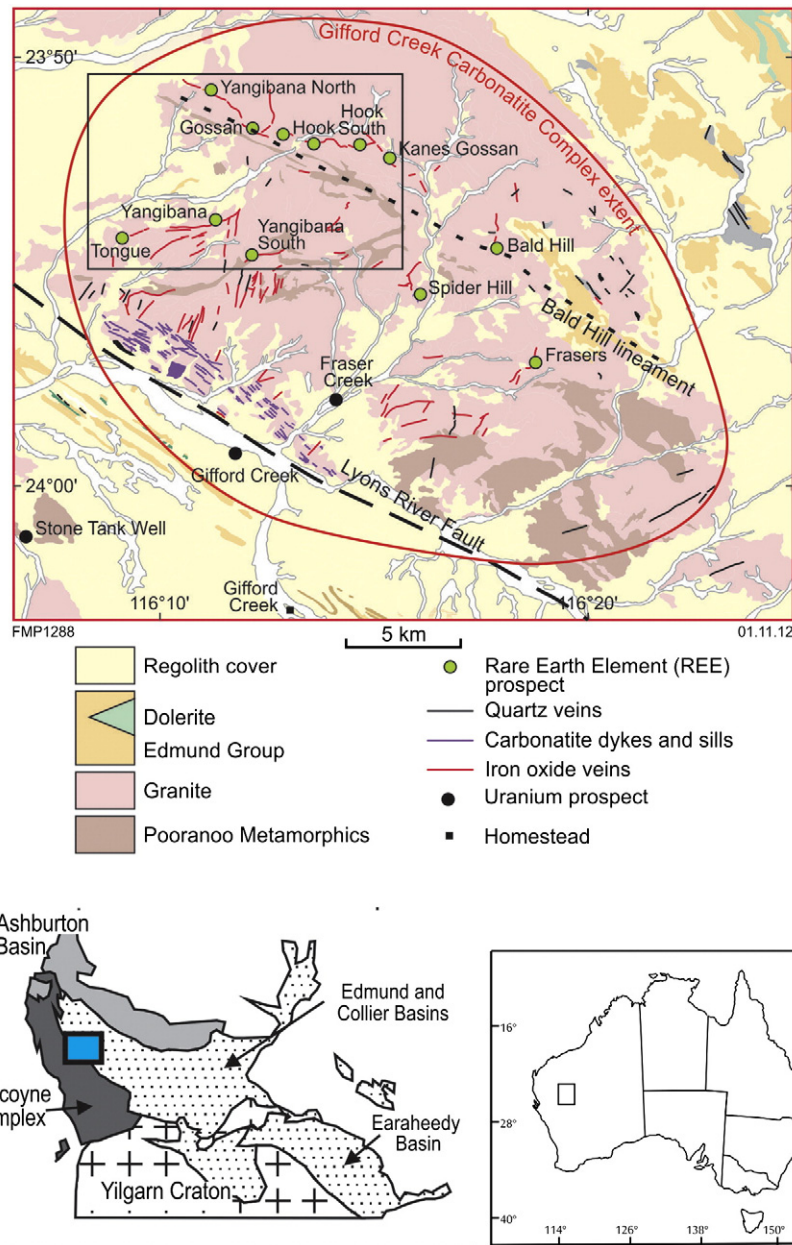


Fig. 19. Simplified geological map of Gifford Creek Ferrocyanatite Complex, showing ironstone veins, dykes and NW-trending belt of ferrocyanatite dykes and sills; also shown are the main REE prospects; position of the Gifford carbonatite complex in the Gascoyne province is shown below the main figure.

Obo sedimentary package, the rift setting, subsequent collision and extensional events that affected the NCC and the accompanying magmatism, it is possible that some of these magmas could have been linked to the mantle plume that originated the rifts and caused the continental volcanism (Xiong'er LIP; Pirajno and Chen, 2005) and dyke swarms in the NCC at ~1780 Ma (Peng et al., 2005, 2008). The complex paragenesis of the ores (Fig. 24) and the wide range of ages probably reflect overprinting of the subsequent thermal, tectonic and metamorphic events.

As mentioned above, the Bayan Obo mineral system is located on the rifted margin of the NCC. The NCC was subjected to rifting events that affected both its northern and southern margins, beginning about 1800 Ma. There is reasonable evidence that this rifting was associated with the emplacement of continental flood basalts and giant dyke swarms (Peng et al., 2005, 2008). In the north, the Taihang Mountain

mafic dyke swarm and the Xiong'er Volcanic Province belong to this ca 1800 Ma thermal event that affected the NCC and have been linked to the breakup of the Columbia supercontinent (Wang et al., 2004; Peng et al., 2005). Peng et al. (2005) working on the geochronology of dyke swarms in the NCC, identified four dyke suites with SHRIMP U–Pb ages of 2147 ± 5 Ma, 1929 ± 8 Ma, 1834 ± 5 Ma and 1778 ± 3 Ma. Peng et al. (2005) suggested that the older group of dyke suites could relate to the amalgamation of the NCC to the Columbia supercontinent, whereas the younger suites are related to the breakup of Columbia and a mantle plume event at 1830–1750 Ma (Hou et al., 2008). This mantle plume would have been implicated in a triple-junction rifting at the southern margin of the NCC at ca 1.78 Ga, leading to its breakup from the Dharwar Craton in India (Peng et al., 2005). In this scenario, the Bayan Obo Group was deposited in a rift system with graben structures on the northern margin of the NCC (Fig. 25), estimated to

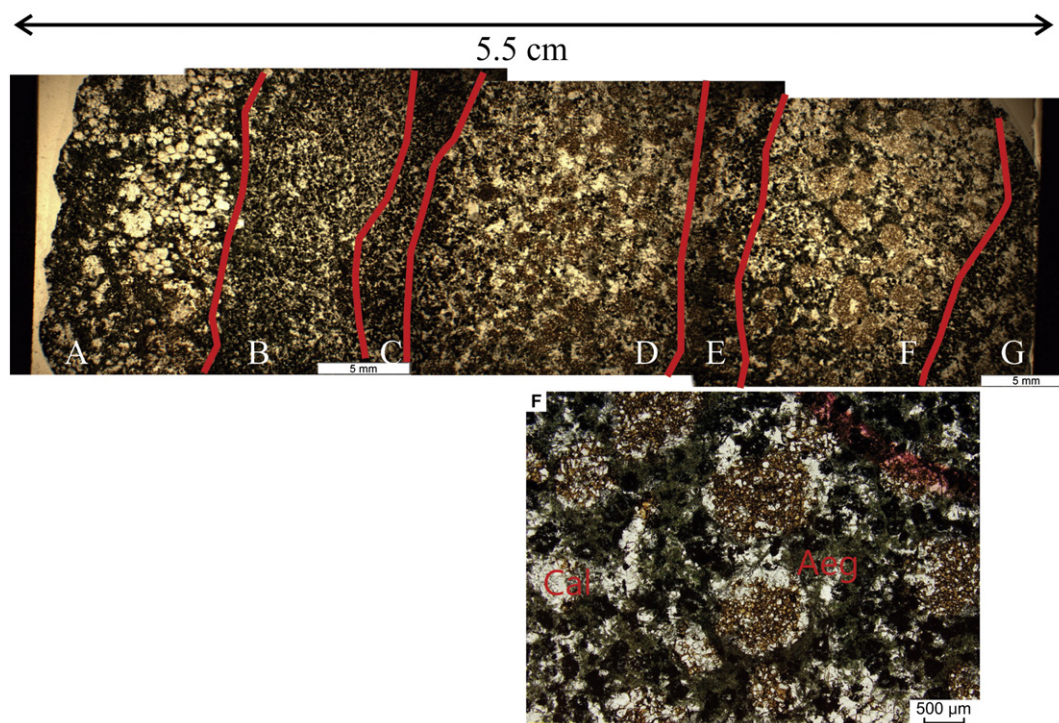


Fig. 20. Gifford Creek sample displaying banded carbonatite at thin section scale. Photomicrographs; A) Calcite–arfvedsonite; B) Fe-carbonate–arfvedsonite; C) Dominantly Fe carbonate and Fe oxides; D) Fe carbonate–calcite–arfvedsonite; E) Dominantly arfvedsonite; F) Fe-carbonate–calcite–arfvedsonite; G) globular Fe-carbonate–calcite–arfvedsonite. A photomicrograph of (F) is shown below, in which partially oxidised Fe carbonate globules are set in a matrix of acicular aegirine and calcite, this texture is interpreted as due to immiscible carbonate–silicate liquids (see text for more details).

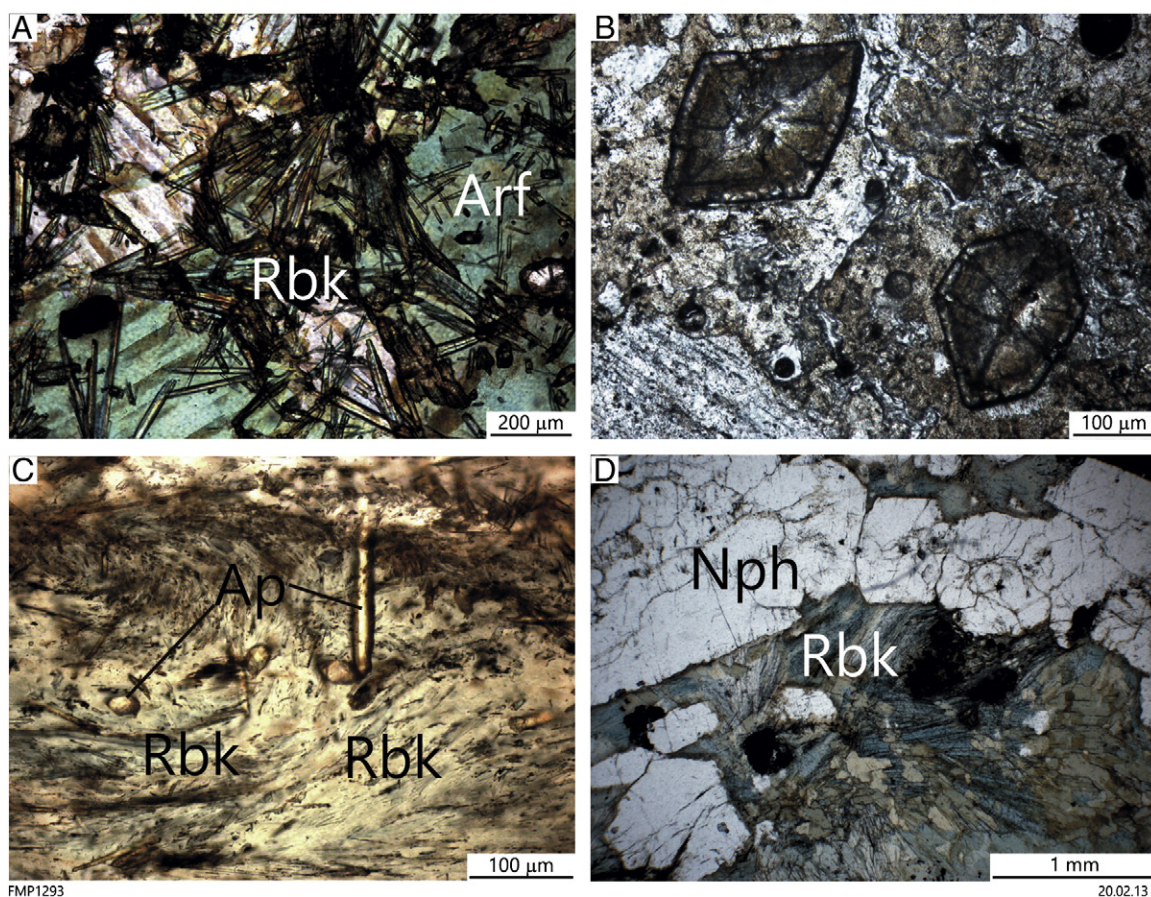


Fig. 21. Photomicrographs in plane polarised light of fenitic wallrocks at Gifford Creek; A) acicular riebeckite (Rbk), overprinting arfvedsonite (Arf); B) zoned pyrochlore crystals; C) acicular riebeckite and apatite (Ap) crystals in a fluidal texture; D) nepheline (Nph) crystals and acicular riebeckite.

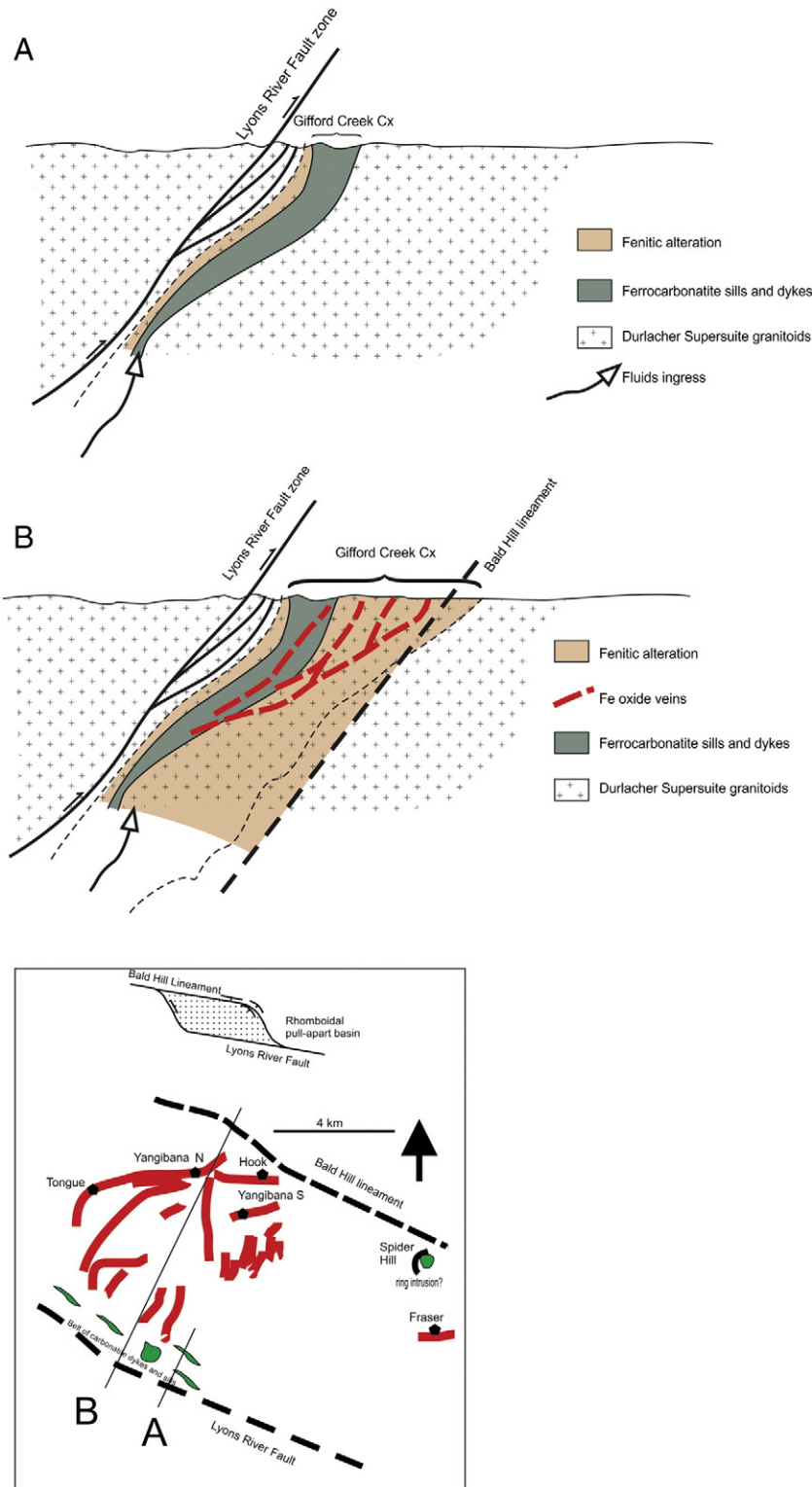


Fig. 22. Suggested model for the emplacement of the Gifford Creek carbonatites in a pull-apart basin formed by strike-slip along the Lyons River Fault, the inward dips on the northern side of the pull-apart structure do not correspond with the presently observed dips, probably due to later re-activation. Based on age data the model consists of two-stage relating to the emplacement of ferrocarbonatite sills and dykes at ~1075 Ma along Lyons River fault (A), followed by strike-slip movements, formation of a pull-apart structure and emplacement of carbonatite–ironstone veins swarm at ~1050 Ma (B); bottom panel shows a schematic spatial distribution of the 1075 Ma ferrocarbonatites dykes and sill and the adjacent 1050 Ma Fe oxides veins swarm emplaced in a pull-apart structure formed during re-activation of the Lyons River Fault; lines A and B in this panel schematically represent the two cross-sections above.

After Pirajno et al. (2014).

have occurred between 1800 and 1350 Ma. Mafic and carbonatite dykes in the Bayan Obo region, have Sm–Nd isochron ages ranging from ca 1418 Ma to ca 1227 Ma (Table 1), confirming mid-late Neoproterozoic

ripping processes within the NCC and have been linked to a mantle plume that led to the breakup of the Columbia supercontinent (Rogers and Santosh, 2002; Hou et al., 2008; Santosh and Zhao, 2009).

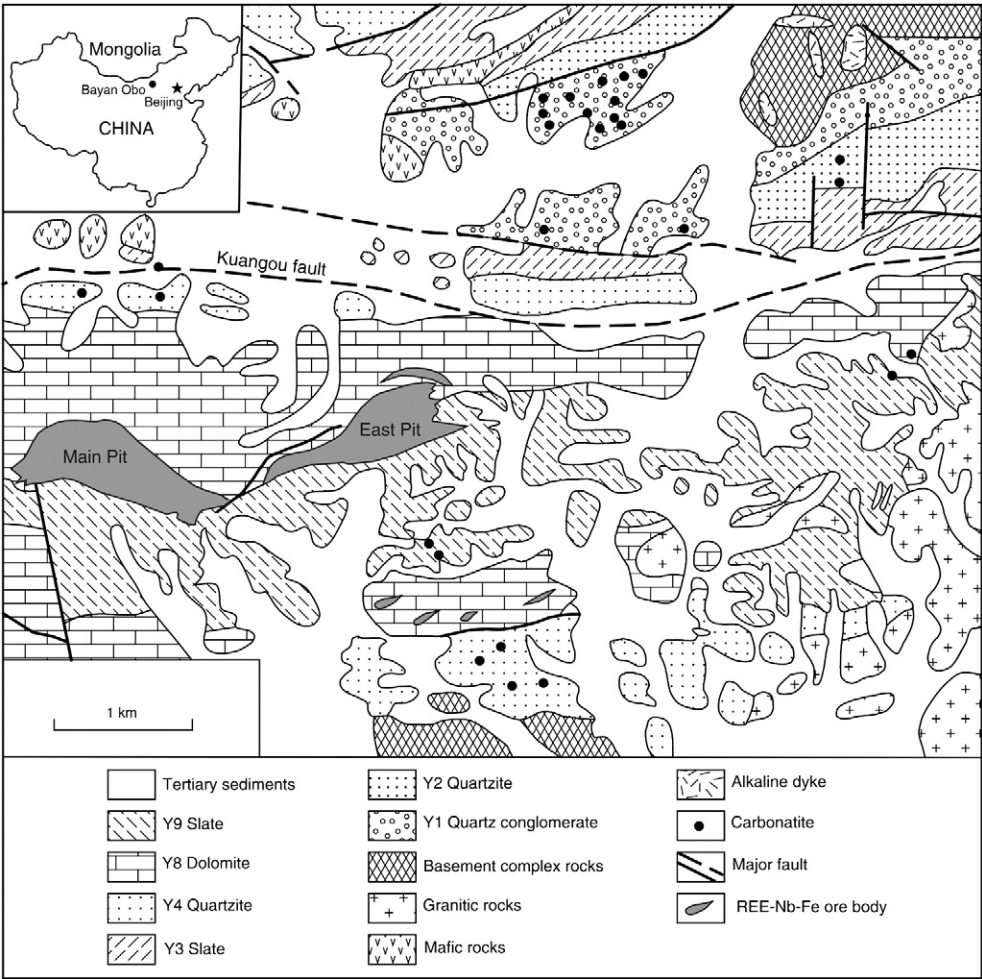


Fig. 23. Simplified geological map the Bayan Obo area, distribution of ore zones and carbonatite dykes, shown as dots because their length (ca <60 m) is too small to be represented at this scale.
After Fan et al. (2004b).

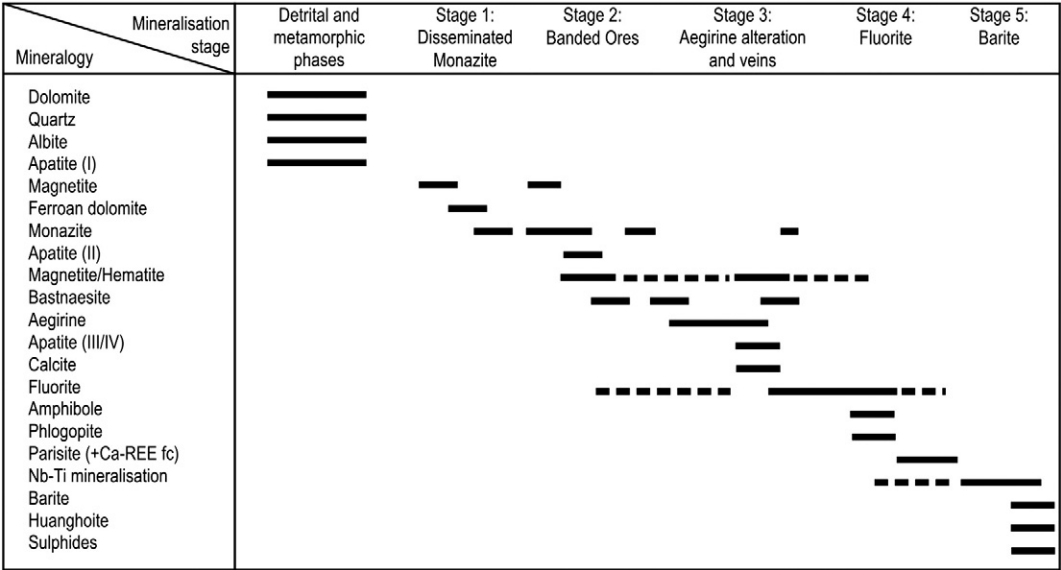


Fig. 24. Simplified paragenesis and mineralisation stages at Bayan Obo.
After Smith et al. (2000).

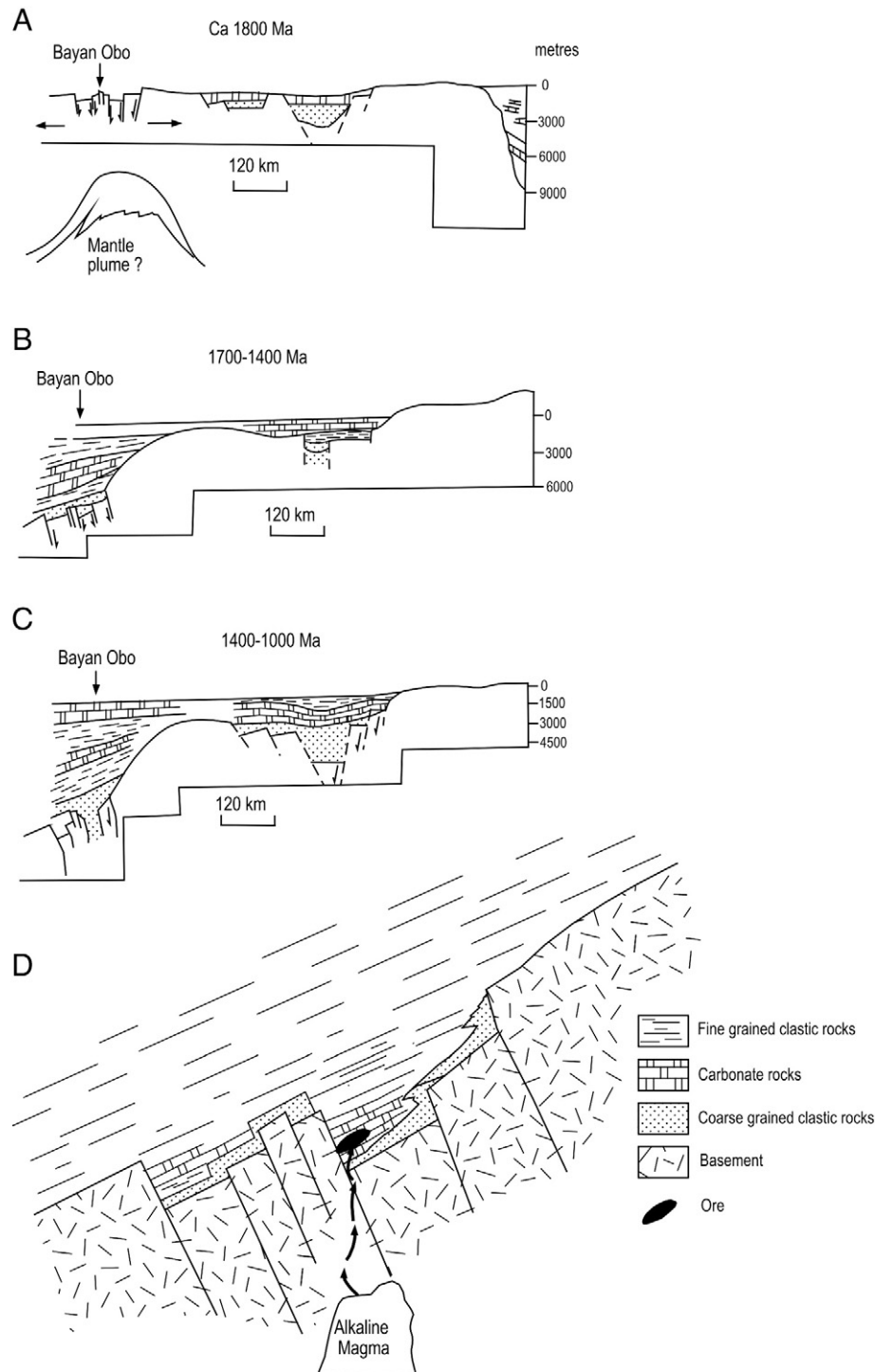


Fig. 25. Evolution of the Bayan Obo rift, between ca 1800 Ma and ca 1000 Ma on the northern margin of the North China Craton. At ca 1800 Ma a major thermal event occurred in the region, possibly due to a mantle plume; details in text. According to [Drew et al. \(1990\)](#).

4.4. Mianning–Dechang belt (Tibet–Sichuan, China)

The 270 km long and 15 km wide Cenozoic Mianning–Dechang (MD) belt in eastern Tibet and western Sichuan (also referred to as Himalayan MD belt) is located on the western margin of the Yangtze Craton and is characterised by numerous syenite and carbonatite intrusions, spatially associated with the prominent strike-slip faults in the region (Red River, Ailaoshan, Xiaojiang faults) and within the

prominent north-trending Panxi rift ([Fig. 26](#)). The Himalayan MD REE belt has been studied by [Yang et al. \(2000, 2001\)](#), [Hou et al. \(2009\)](#) and [Xie et al. \(2006, 2009, 2014\)](#). The synopsis given below is taken from these authors.

The MD carbonatite–alkaline complexes appear to be related to the highly evolved CO_2 -rich silicate magma. Fluid inclusion data indicate that gangue minerals such as fluorite, quartz and barite from the MD complexes host abundant melt and melt/fluid inclusions with

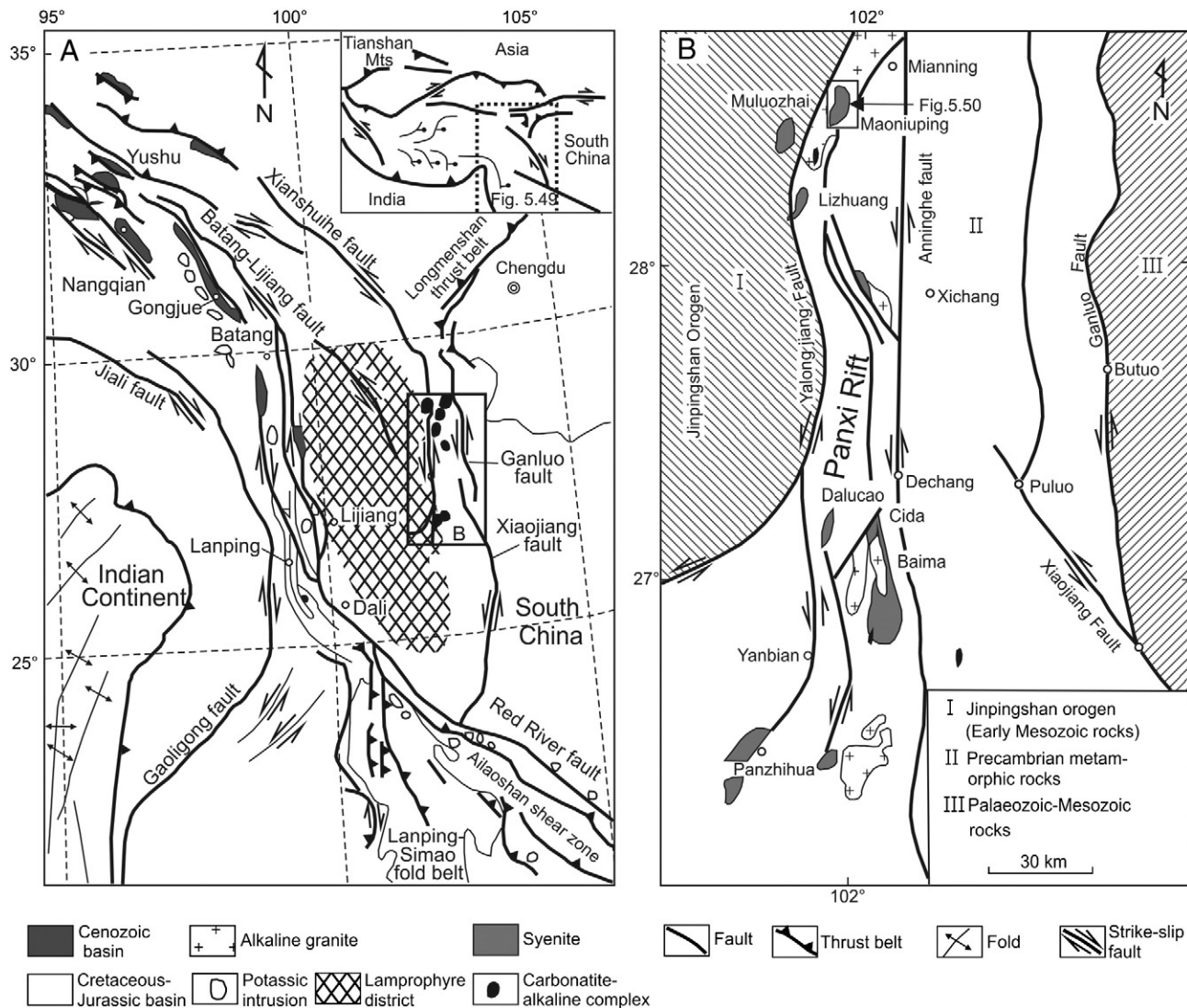


Fig. 26. Distribution of strike-slip fault-controlled Cenozoic Mianning–Dechang (MD) belt of alkaline intrusions and carbonatites in eastern Tibet and the north-trending Panxi rift (A), detailed map of the area of REE mineralised carbonatites (B); inset in A shows regional position of the MD region. After Hou et al. (2009).

abnormally high homogenisation temperatures (up to 750 °C) and enriched in BaSO_4 , K_2SO_4 , CaSO_4 , CaCO_3 , CaF_2 and CO_2 (Xie et al., 2009). Xu et al. (2003) determined C, O and Sr isotopic compositions for the Maoniuping REE deposit, showing that the $\delta^{13}\text{C}_{\text{PDB}}$ and $\delta^{18}\text{O}_{\text{SMOW}}$ values range from -6.6% to -7.0% and 6.4% to 7.4% , respectively, are well within the range of primary igneous carbonatites. Similarly, the initial Sr and Nd isotopic compositions (e.g. $^{87}\text{Sr}/^{86}\text{Sr} = 0.706074$ to 0.706149 ; $\epsilon_{\text{Nd}} = -4.2$ to -3.7) fall within the range of both carbonatites and syenitic rocks (Xu et al., 2003). All these features are typical of carbonatites and peralkaline magmas.

In the MD belt a liquid immiscibility origin for carbonatite–syenite suite is advocated by Hou et al. (2009), who cited the following points to reinforce their view: (1) the close temporal and spatial relationship between carbonatites and alkalic syenites; (2) low contents of MgO ($<0.73\%$) and FeO ($<1.20\%$) in the carbonatites, distinguishing them from primary magnesiocarbonatites derived directly from partial melting of a mantle source; and (3) similar REE patterns of syenites and carbonatites with higher total REE amount and lower Ce/Yb ratios, which coincide with experimental results on REE partitioning between immiscible carbonatitic and silicate melts. In addition, F- and REE-rich fluids can be separated from carbonatite magma by liquid immiscibility.

Himalayan magmatic activity (40 to 24 Ma, with a peak of activity at 35 Ma) in the region forms a nearly continuous potassic igneous province, including from west to east: (1) a 1000-km-long belt, of potassic igneous rocks, associated with early-middle Cenozoic pull-apart basins and developed along the Nanqian thrust, the Batang–Lijiang fault and the Red River shear zone; (2) a 50,000 km² shoshonitic (calc-alkaline) lamprophyre district; and (3) the MD belt of carbonatite–alkaline complexes (Fig. 26). The carbonatite–alkaline complexes in western Sichuan intrude Proterozoic crystalline basement and overlying Palaeozoic–Mesozoic volcano-sedimentary sequence, forming a narrow, north-trending REE mineralised corridor, bordered by north–south trending strike-slip faults (Fig. 26). There are two districts of carbonatite–alkaline complexes: 1) a northern district, which comprises Maoniuping, Lizhuang and Muluoizhai; and 2) a southern district controlled by the Dalucao strike-slip fault (Fig. 26B). The northern district has 40–30 Ma syenitic intrusions with carbonatite dykes and sills, whereas the southern area consists of syenite stocks, carbonatite sills, and REE-bearing breccia-pipes, with SHRIMP U–Pb zircon ages for carbonatite and nordmarkite of 12.99 ± 0.94 Ma and 14.53 ± 0.31 Ma, respectively. Fluid inclusions (Hou et al., 2009) include aqueous–vapour, aqueous–liquid CO_2 , aqueous–liquid CO_2 –daughter

mineral, aqueous–vapour CO_2 –daughter mineral and pure- CO_2 inclusions. Melt inclusions (Xie et al., 2009) are found mainly in calcite and fluorite from the carbonatites, and lesser amounts in gangue quartz. Melt/fluid inclusions are abundantly present in gangue fluorite and contain several solids, occupying 60–95 vol.% of the inclusion. Raman spectroscopy and SEM/EDS analyses show that these solids mainly consist of bastnaesite, gypsum, barite and fluorite in bastnaesite-hosted inclusions; fluorite and celestine in fluorite-hosted inclusions; quartz, calcite and gypsum in quartz-hosted inclusions; and barite, celestine and strontianite in calcite-hosted inclusions (Yang et al., 2001). Thermometric measurements of fluid inclusions revealed the complexity of carbonatite and REE mineral fluids, containing K_2SO_4 – Na_2SO_4 – CO_2 – H_2O , instead of the more common NaCl – H_2O – CO_2 (Hou et al., 2009). Homogenisation temperatures (T_h) measured from fluid inclusions in quartz, calcite, fluorite, barite and bastnaesite indicate average values ranging from ca. 410 to 142 °C. Stable isotope systematic (O–D and C) show variable values that are difficult to interpret, but suggesting mixing of high- T hydrothermal fluids with crust-derived fluids, as well as meteoric fluids introduced in the final stages in the evolution of the carbonatite–REE mineral system.

The Himalayan MD belt, is associated with the Indian–Asian continental collision at 65–55 Ma and a series of Cenozoic strike-slip faults, reactivated fault systems (Guo et al., 2005; Xia et al., 2012) and large-scale shear zones, such as the Red River shear zone, along the eastern margin of the Tibetan Plateau and Eocene–Oligocene pull-apart basins and alkaline igneous rocks formed by these strike-slip faults at about 35 Ma (Hou et al., 2003).

5. Mineral systems associated with carbonatites and fenitisation

Carbonatites are the main host of Nb–Ta, Fe-apatite, Nb–REE and fluorite mineralisation, both magmatic and hydrothermal (Mariano, 1989). Of these, the largest include Bayan Obo (China), Mountain Pass (USA) and Palabora (also spelled Phalaborwa, South Africa) (see Pirajno, 2009 for overviews and references). Carbonatite-related Fe-apatite, Nb–Ta, Nb–REE deposits (e.g. Tomtor, Karasug, Beloziminskoe) are numerous in the Siberian Craton and surrounding orogenic belts (Seltmann et al., 2010). The Mountain Pass carbonatite typically contains bastnaesite–parisite as primary magmatic minerals, albeit in a late-stage carbonatite (Castor, 2008). Hydrothermal REE minerals are more common, forming polycrystalline aggregates, veinlets and interstitial fillings, and are usually associated with barite, fluorite and monazite in varying amounts, which may or may not be economic. Bastnaesite is one of the more important REE mineral species, together with ancylite, parisite, synchysite and britholite. Most hydrothermal carbonatite-hosted REE deposits are small and uneconomic, although it is possible that a Bayan Obo style mineral system could be found in rift-related tectonic settings (Pirajno, 2009). Weathering of carbonatite complexes results in dissolution of carbonate minerals, leaving residual and/or supergene monazite, apatite and REE minerals, such as bastnaesite, parisite and florencite, ilmenite, rutile, monazite and synchysite in the carbonatite rocks (Lottermoser, 1990). Apatite is one of the principal accessory minerals in carbonatite and alkaline complexes and in some cases, apatite can exceed >50 vol.% (Mariano, 1989).

Carbonatites are also a major resource of Nb, which is mostly contained in minerals such as pyrochlore and ferrocolumbite (Table 2), in addition to minerals in which Nb proxies for other elements (e.g. perovskite, titanite). Pyrochlore too precipitates from hydrothermal fluids and may be associated with bastnaesite and fluorite. Other ore minerals of economic importance include titanite, which can be quite abundant in carbonatite and alkaline complexes, whereas rutile and brookite can be found as disseminations in fenitic aureoles, associated with carbonate, fluorite and low-temperature feldspar (Mariano, 1989). The element Ti can reach economic abundances in the mineral anatase, as a weathering product of perovskite. Fluorite too forms a

common accessory in many carbonatite and alkaline complexes. At Bayan Obo, fluorite associated with bastnaesite, is a major ore mineral (see below). It must be borne in mind that F in hydrothermal fluids can move for long distances and fluorite can be precipitated in host rocks that have no relationship with carbonatites or a carbonatite source. Other elements that can be important as economically viable ores are Ba and V. Barite can crystallise from carbonate melts or from low-temperature hydrothermal fluids, as is the case at the Mountain Pass carbonatite (Castor, 2008). Vanadium can be hosted in minerals such as aegirine and magnetite in carbonatites and alkaline ultramafic rocks. Copper is the major ore element of the Palabora carbonatite complex, where it may have resulted from the circulation of hydrothermal fluids (see Pirajno, 2009 for an overview and references therein). Thorium and U are enriched in carbonatites and alkaline rocks, with Th usually being more abundant than U. Thorium- and U-bearing minerals of carbonatites include perovskite, pyrochlore, thorite and REE mineral species (Table 2). Chakhmouradian and Wall (2012) and Chakhmouradian et al. (2014) reviewed and carried out extensive research on the role in modern society of critical metals (including REE and Y).

At *Kruidfontein*, evidence of hot spring activity is indicated by the occurrence of siliceous and Fe-rich sinter material in the inner ring (Fig. 13). Fluorite mineralisation is present and was probably formed during a late-stage hydrothermal event (Crocker et al., 1988; Clarke, 1989). At least three types of fluorite mineralisation are present: (1) primary fluorite, forming disseminated grains in the sövite rocks, and reaching average grades of about 20% CaF_2 ; (2) secondary fluorite mineralisation is present as replacement bodies in the metabasite rocks, where it may reach concentrations of up to 60% CaF_2 ; and (3) quartz–fluorite veins and fissure fillings. Ore grade and anomalous Au occurs in the weathered carbonatitic pyroclastics within the inner ring of the KCC close to vent areas and is bound by north-trending and east–northeast-trending faults (Fig. 12). In an attempt to explain the KCC gold mineralisation Pirajno et al. (1995) considered two genetic models: 1) a supergene origin and 2) a hypogene-hydrothermal origin for the mineralising solutions.

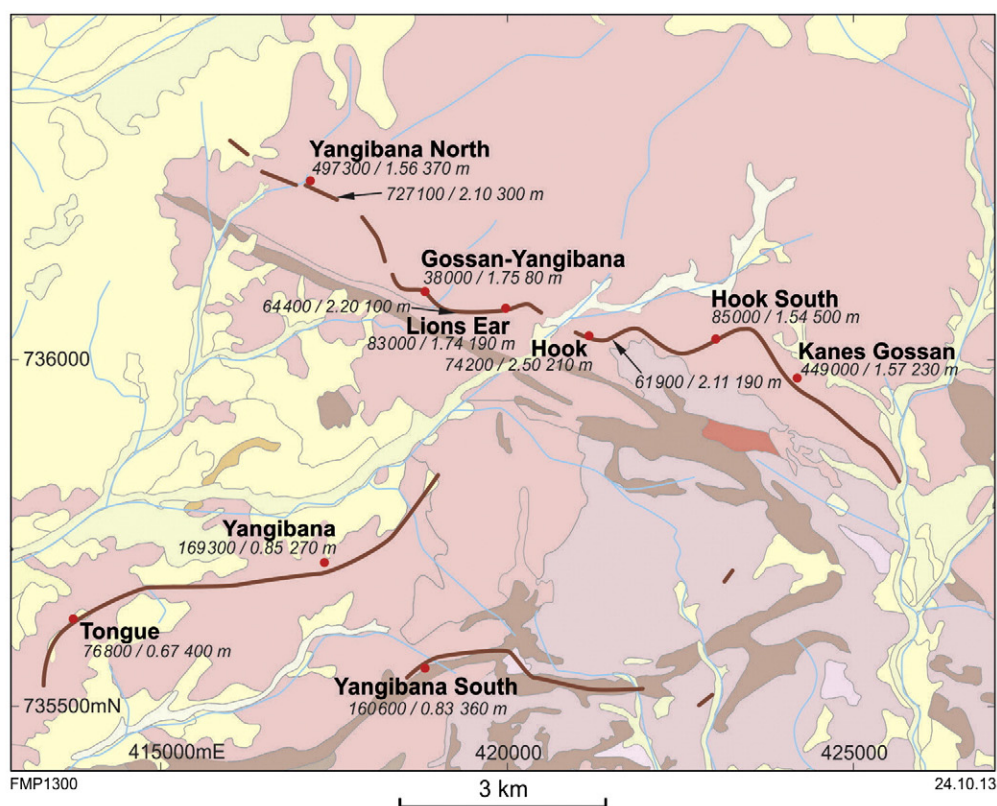
In the supergene model, the oxidised zone is interpreted as a lateritic profile. This consists of a surficial ferruginous zone (up to 25 m), a Fe- and Mn-rich oxide zone (25–50 m), mottled clay zone and ferruginous saprolite (50–77 m) and saprolite to about 150 m. Gold, Mn, and Y are concentrated in the upper 80 m of the profile. Lateritic supergene enrichment implies secondary accumulation of gold from a precursor similar to that of the underlying fresh carbonatites. Such enrichment, together with any vertical or lateral dispersion, is by both physical and chemical mechanisms. Enrichments at similar depths are a feature of lateritic regolith in semi-arid environments, as exemplified by the Yilgarn Craton of Western Australia (Butt, 1989; Anand, 2011). This redistribution of gold deep in the profile is considered to be an arid phase modification, perhaps due to mobilisation and subsequent complexing by chloride ions. Therefore, it is possible that gold was mobilised as a thiosulphate complex resulting in some redistribution and local concentration of gold in the upper 80 m.

An alternative hypothesis is that the Au mineralisation was formed through the action of hydrothermal solutions. In this model the hydrothermal fluids, activated by cooling magma at depth, ascended along faults, leaching out gold from a postulated Au-rich protolith at depth. Fluid–rock interaction resulted in Fe-rich alteration (Fig. 13D), starting from microfractures (Fig. 14C) and gold deposition close to the palaeosurface. Progressive decrease of Fe enrichment eastward from a fault zone is seen to occur in the altered pyroclastics. Perhaps the fluids vented as hot springs depositing barren silica at the surface, as suggested by the presence of sinter-like deposits in the area (Fig. 13C). The gold may have been transported as sulphide complexes and precipitated in the carbonatitic pyroclastic unit which was, at a later stage, subjected to weathering and lateritisation. These later supergene processes could have further enhanced the gold mineralisation.

At Goudini the northwest-trending dykes host fluorite, barite and monazite (Verwoerd, 2008). Barite veins and disseminations occur in sövitic rocks (Fig. 16), whereas Y and heavy rare-earths and thorium are found in carbonate minerals. Fluorite and barite are hosted by, or intergrown with flow-banded dolomitic carbonatite (Verwoerd, 1993, 2008).

The laterite cap of the Mount Weld carbonatite hosts world-class resources of Nb, Ta and P. Total indicated and inferred REO + Y resources, as reported in Hoatson et al. (2011), are 17.5 Mt grading 8.1%. The Mount Weld carbonatite complex contains residual apatite at the base of the regolith, magnetite, vermiculite and Y and lanthanide compounds, derived from the primary pyrochlore. The lateritic regolith

profile (Duncan and Willett, 1990; Lottermoser, 1990; Jaireth and Hoatson, 2014) consists of a residual zone overlying the fresh carbonatite, in turn overlain by a supergene-enriched zone. This lateritic regolith is overlain by alluvial and lake sediments (Fig. 17). The residual zone at the base of the profile is a karst-like feature developed on the fresh carbonatite and contains relic primary minerals such as apatite, magnetite, ilmenite, pyrochlore and monazite (Jaireth and Hoatson, 2014). The overlying supergene zone contains phosphates, clays, crandallite-group minerals, and Fe and Mn oxides. The presence of secondary monazite in polycrystalline aggregates locally replacing apatite reaches concentrations of up to 45% combined lanthanide oxides. In general grades are quite variable ranging from 1.5% to 6%



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Fig. 27. Sketch map showing distribution of the mineral prospects, with defined resources within current tenements owned by Hastings Rare Metals Ltd.; typical outcrops of ironstone veins are shown below A) unnamed location; B) Tongue prospect. The position of the present figure is shown by the square in Fig. 19. After Pirajno and González-Álvarez (2013).

Nb₂O₅ (Duncan and Willett, 1990; Jaireth and Hoatson, 2014). The unanswered question is the age of the lateritic regolith, which could range from Permian to Early Cenozoic. During the Permian a weathering crust would have been removed by glaciers, thereby exposing a clean carbonatite surface on which the present regolith was formed, followed by the deposition of lacustrine sediments in the Late Cretaceous to Cenozoic.

The Gifford Creek Ferrocarbonatite Complex is at the time of writing being explored by mining concerns for their REE and U potential. In the past exploration efforts were directed at the Fe oxide veins, largely because of their high visibility due to their gossanous appearance (Fig. 27). An overview of the mineralisation at Gifford Creek is provided by Flint and Abeyasinghe (2000). Numerous sinuous ferrocarbonatite veins associated with ironstone veins and pods (Figs. 19 and 27), enclosed or surrounded by zones of fenitisation of the country rocks, are located to the north of the ferrocarbonatite belt. The above-mentioned ironstone veins, associated with carbonatites, are one of the main features of the Complex (Pirajno and González-Álvarez, 2013). The veins are characterised by having sinuous or arcuate trends and locally occur as pods of massive Fe oxides and are generally between a few centimetres to about 10 m thick. Veins of hematite–magnetite, in places with supergene alteration to massive goethite, are locally associated with quartz veins. The presence of quartz is attributed to later hydrothermal activity, distinct from the effects of Fe-oxide rich and fenitising fluids. Attitude, field relationship, and petrographic data suggest that the ironstone veins post-date the emplacement of the ferrocarbonatite intrusions. Flint and Abeyasinghe (2000) recorded rare earth elements and uranium mineralisation in the ferrocarbonatite and ironstone veins and quoted estimated combined resources of 3.5 Mt, averaging 1.7% REE oxides. Recent drilling by Hasting Rare Metals Ltd. at Yangibana reported low grades of the Heavy Rare Earth Oxides (HREOs), but of particular interest is the high Nd grade, which averages around 0.4 wt.% Nd₂O₃, or 24% of the Total Rare Earth Oxides (TREO; Hastings Rare Metals Ltd., 2012). The distribution of the REE prospects with defined resources (Pirajno and González-Álvarez, 2013) is shown in Fig. 27.

The Bayan Obo REE–Nb–Fe deposit in Inner Mongolia contains a resource of 57.4 Mt with grades ranging from 5.17 to 6.19% REE₂O₃, Nb reserves estimated at about 2.2 Mt with average grades from 0.126 to 0.141% Nb₂O₃ and 1.5 Bt of Fe at an average grade of 35% (Drew et al., 1990). Bayan Obo constitutes the largest rare earth element (REE) resource so far discovered, representing about 95% of China's total REE resources (Zhu, 2007) and about 60–70% of the world's total. This unusual deposit was discovered in 1927 by one Ding Daoheng and has been, and still is, the object of numerous studies of which, in addition to the above, are those cited in Section 4.3. I have drawn from these works for the account that follows. There are three main orebodies: Main, East and West, all characterised by an Fe-rich zone, dolomite, aegirine, riebeckite, biotite and fluorite zones. The ore material is hosted by fine- and coarse-grained dolomitic marble of the Y8 unit, which is banded and consists of recrystallised dolomite, apatite, magnetite, with minor pyrochlore and Na-amphibole. Ore minerals include monazite, bastnaesite, huanghoite [(Ba,Ce,La,Nd)CO₃]₂F, parasite, and cebaite [Ba₂Ce₂(CO₃)₅F₂] for the REE ores, columbite, pyrochlore, fergusonite (YNbO₄), and aeschynite [(Ce,Ca,Fe,Th)(TiNb)₂(O,OH)₆] for the Nb ores and magnetite, hematite, ilmenite, martite and goethite for the Fe ores. Barite, fluorite, apatite, cerianite, aegirine, arfvedsonite, riebeckite, albite and microcline are common gangue minerals. Thus, the genesis of the Bayan Obo mineral system, like its age, remains controversial. Ore genesis models included synsedimentary and exhalative deposition or epigenetic hydrothermal linked to granite magmatism or A-type alkaline magmatism and/or carbonatite magma. The main question hinges on the origin of the ore-hosting Y8 dolomitic marble, with ideas ranging from sedimentary, hydrothermal, skarn to carbonatite-related. The skarn hypothesis has problems because of the age gap between the granitic rocks (~311–306 and 258–234 Ma) and the age of the monazite

(555–420 Ma; Wang et al., 1994). The presence of carbonatite and albitite dykes that cut the sedimentary package a few km north and south of the orebodies and the composition of the dolomite support the carbonatite hypothesis. Yang and Le Bas (2004) showed that the Y8 dolomitic marble in the ore zone has a geochemical composition (MnO and SrO contents higher than 0.15 wt.%) similar to that of the carbonatite dykes. Therefore, the ore-hosting dolomitic marble has carbonatitic affinities and may not be part of the Y8 sedimentary carbonate sequence (Yang and Le Bas, 2004). Le Bas et al. (2007), largely on the basis of REE abundances and major elements data, proposed that the various carbonate dykes and ore-hosting dolomitic rocks, represent intrusive carbonatite (carbonatite, ferrocarbonatite and magnesio-carbonatite), which are indisputably different from sedimentary carbonate rocks.

The hydrothermal system that formed Bayan Obo is responsible for widespread alteration over an area of ~50 km² in both Archaean migmatites and sedimentary rocks. Drew et al. (1990) distinguished footwall and hanging wall alteration. Calcite–barite–magnetite–albite–monazite veins, up to 3 m wide, cut the footwall migmatite rocks forming alteration haloes containing aegirine, Na amphibole, albite and K-feldspar. Thinner veins and stockworks cut the Y7 and Y6 units and have the same mineralogy as those that cut the migmatites. The Y8 unit is pervasively altered as the clastic rocks and migmatites described above. Sodic amphibole, micas (annite, biotite and phlogopite) and Fe oxides are the dominant minerals. Magnetite and hematite are disseminated in the host carbonate rocks, locally forming lenses up to 20 cm thick along bedding planes in economically viable concentrations. In the hanging wall, the Y9 shale acted as an impermeable cap rock focusing the fluids derived from the reactions that replaced the Y8 dolomite, along the contact between Y8 and Y9, resulting in the emplacement of the REE–Nb–Fe orebodies. The base of the Y9 shale is brecciated and cut by small veins and stockworks, which exhibit a zonation of alteration minerals, outward from fluorite to biotite to K-feldspar. K-feldspar is dominant alteration phase in the matrix of the brecciated material. The paragenesis of the Bayan Obo ore is also complex with 11 stages proposed by Chao et al. (1997), from deposition to metamorphism and mineralisation, to granite intrusions. This was simplified by Smith and Henderson (2000), Smith et al. (2000) and Smith (2007), with a paragenetic sequence, shown in Fig. 24, comprising five stages of mineral deposition. A panoramic view of the Bayan Obo open pit, together with a type hand sample and corresponding photomicrograph, is shown in Fig. 28.

Carbonatite–syenite complexes and associated REE mineralisation in the Mianning–Dechang belt are estimated to contain in excess of 3 Mt of REE ore in the Maoniuping and Dalucao deposits. Hou et al. (2009) classified the Maoniuping REE deposit as world-class, containing a resource of 1.2 Mt REO at an average grade of 2.89% REO; with more recent estimates indicating a total resource of 3.17 Mt REO (Xie et al., 2014). The REE mineralisation is temporally and genetically associated with the Maoniuping complex, which consists of syenitic and carbonatite intrusions, porphyritic alkaline aegirine–augite-bearing granitic dykes and alkaline pegmatite dyke swarms. The Maoniuping carbonatite–alkaline complex is controlled by a strike-slip fault, is about 1.4 km long and 260 to 350 m wide and is intruded into Mesozoic granitic and rhyolitic rocks. The Maoniuping complex is surrounded by a 1800 m long and up to 600 m wide halo of fenitic alteration, mainly represented by albite, aegirine–augite, arfvedsonite and Mg-rich biotite replacing country rocks. This fenitic halo hosts REE-bearing albite veins cutting through the Mesozoic country rock and a zone of aegirine and aegirine–augite developed within the carbonatite and pegmatitic rocks. The fenitised rocks contain disseminated bastnaesite, but the bulk of the REE mineralisation is related to late stages of fenitisation, typically characterised by veins containing calcite, fluorite, barite, celestite, aegirine–augite and bastnaesite. The Maoniuping vein system has a north–northeast strike and is about 2.65 km long and is made up of 1 to 30 cm-thick veinlets, stockworks and stringers. Stockworks generally occur at the margins of the vein zones and consists of

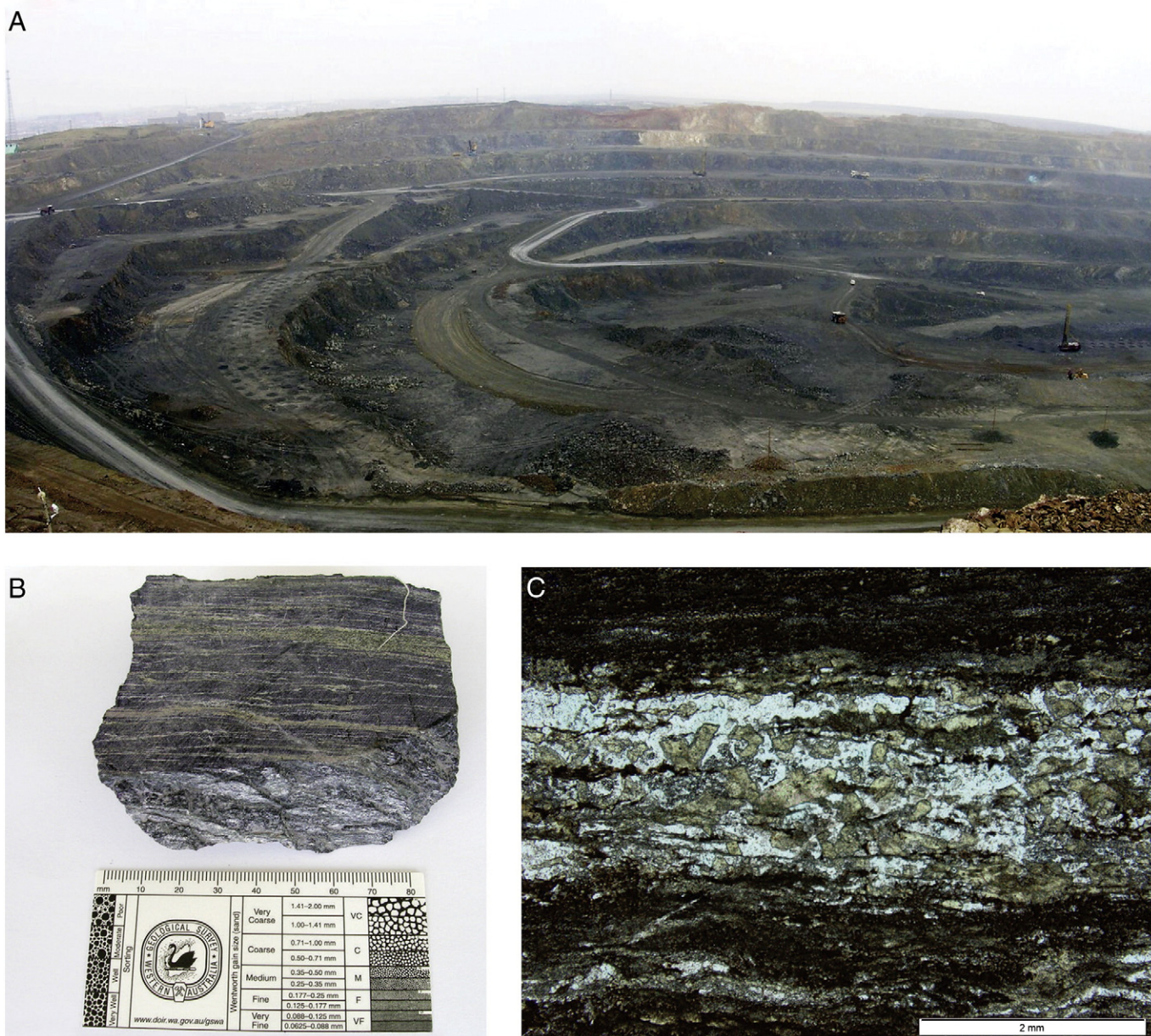


Fig. 28. A) View of the Bayan Obo pit (photo taken in 2008), B) hand specimen of banded fluorite, aegirine, bastnaesite + nepheline; C) photomicrograph showing bastnaesite crystals in centre, flanked by fluorite and granular aegirine.

bastnaesite-bearing barite + aegirine-augite, fluorite + orthoclase and calcite + fluorite. In the north of the complex large (15 to 50 cm long) bastnaesite crystals are present within a fluorite-quartz-barite-bastnaesite zone in the centre of the system to an outer barite-aegirine-augite zone. In addition, microscopy studies by Xie et al. (2006), revealed the presence of unusual minerals including zinccooperite (Cu_2Zn) and Sn-bearing native copper. Hou et al. (2009) reported that at the time of their study, 71 orebodies averaging from 1 to 9.05% REO had been defined by drilling programmes. Four main ore types have been recognised: 1) Veinlets or pockets, in the northern part of the complex and consisting of pegmatitic barite + fluorite + aegirine-augite + bastnaesite and orthoclase + aegirine-augite + fluorite + bastnaesite. 2) Carbonatitic ore as disseminations in dykes and/or sills, and forming thick lenses in the northern parts of the deposit, containing coarse-grained calcite, barite (10–20%), fluorite (5–10%), and minor bastnaesite (0.5–5%), locally with varying amounts of pyrite, sphalerite and galena. 3) Brecciated REE ore occurs in the southern and middle segments of the deposit,

grading laterally into stringer and parallel vein zones; the breccia is clast-supported, with the clasts consisting largely of angular fragments of nordmarkite, in a matrix of fluorite + barite + calcite + bastnaesite assemblages. 4) Stringer ores form stockwork zones between the main orebody and surrounding wall-rocks (nordmarkite and fenitised granite pluton).

The age of the Maoniuping REE deposit reported by Yuan et al. (1995) and determined by K–Ar dating of biotite and magnesioarfvedsonite, yielded ages ranging from 27 to 40 Ma (see details in Table 1).

A genetic model proposed by Hou et al. (2009) is shown in Fig. 29, in which the authors envisage three depth levels (probably end-members) of carbonatite emplacement and related REE mineralisation. At the lower level (~2.5–3 km), are high-T, REE- and sulphate-bearing NaCl–KCl– CaCl_2 fluids, derived from the carbonatitic-syenitic magmas with no interaction with external fluids. Filling and replacement of carbonatite by calcite + fluorite + barite + bastnaesite assemblages represent mineralisation along fractures and as dissemination within

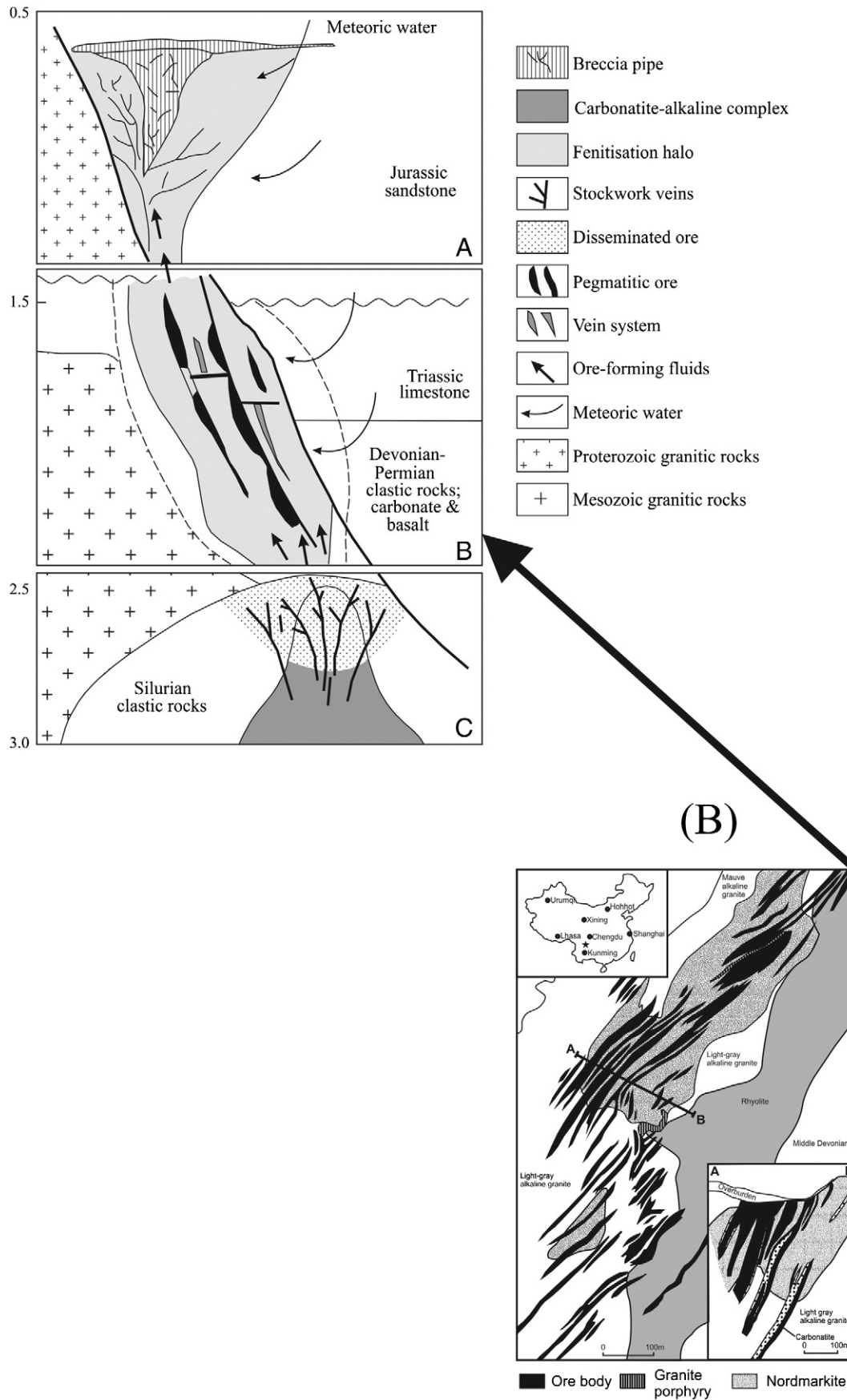


Fig. 29. Genetic model of REE mineralisation associated with carbonatite-alkaline intrusion in the Mianning–Dechang belt (explanation for this figure in text). After Hou et al. (2009).

and above the carbonatite–syenite complex (panel C of Fig. 29). At intermediate levels, multi-phase hydrothermal activity resulted in the formation of pegmatitic and complex vein systems developed along a mesh of fractures developed in a transpressional regime and characterised by bastnaesite-bearing assemblages (panel B in Fig. 29 and geology and cross-section of the Maoniuping REE deposit). At high levels, extensive boiling of the ore forming fluids, resulting in hydrothermal breccia pipes, accompanied by REE mineral precipitation (panel A in Fig. 29). Influx of meteoric waters into the hydrothermal system caused precipitation of lesser amounts of fine-grained bastnaesite, quartz and Fe oxides.

6. The mantle plume connection: discussion

The precise origin of alkaline and peralkaline magmas is still debated, however there is good evidence to indicate that they form through small degrees of partial melting and that they originate in the SCLM (Groves et al., 2010 and references therein). Mantle metasomatism was advocated by Bailey (1987), who speculated that rifting may localise mantle degassing (especially C and H) which would metasomatise the SCLM and cause partial melting. He pointed out that large quantities of CO₂, F and Cl are emitted from modern rifts, such as the East African Rift System and the spectacular CO₂ active venting at the Mefite d'Ansanto site, just west of the Vulture alkaline–carbonatite volcanic system in southern Italy (Rogie et al., 2000; Stoppa et al., 2009; Bell et al., 2013). Thus, rifts are the sites of upwelling asthenospheric melts rich in Cl, F, C, P, as well as HFSE (Nb–Ta, Zr–Hf, Th). These melts penetrate the lithosphere, metasomatising it and causing partial melting and developing thermal anomalies. It is of interest to note that carbonated melts have very low viscosities and, based on fluid-flow modelling, these melts are thought to rise to the surface at speeds of up to 65 m s^{−1} (Genge et al., 1995; Ernst and Bell, 2010).

A study by Bell and Simonetti (1996), reinforced in Bell (2001) and further taken up in Bell and Simonetti (2010), based on Nd, Pb and Sr isotopic systematics from Oldoinyo Lengai, a 2200 m high nephelinite–phonolite–carbonatite volcano in the East African Rift (northern Tanzania) is supportive of a mantle plume source. Other views in the case of Oldoinyo Lengai suggest that its volatiles are indistinguishable from those of a mid-ocean ridge (MORB) source (e.g. Fischer et al., 2009). Bell and Simonetti (1996) envisaged that carbonatites and associated nephelinitic magmas are derived from a two-stage process, linked to a mantle plume. In the first stage, melts are generated directly from an upwelling mantle plume. Volatiles released from the plume itself cause intense metasomatism of the overlying lithosphere. In the second stage, low degrees of partial melting in the metasomatised subcontinental lithosphere produce the carbonatite–nephelinite volcanic activity that characterises the Gregory Rift in the East African Rift System (Pirajno, 2000, 2009 and references therein).

The upwelling of a mantle plume beneath a subcontinental lithosphere leads to the formation of a “mushroom-shaped” head of hot mantle material, which could reach up to 1500 km across (Campbell and Griffiths, 1990; Griffiths and Campbell, 1990). The lithosphere above this head is dynamically lifted, resulting in doming of the crust and subsequent rifting processes. However, it must be pointed out that a mantle plume impacting the lithosphere does not need, and most probably will not have a symmetrical geometry, as depicted in many simple experimental laboratory models. An irregular surface topography due to heterogeneities of the overlying plates as well as tectonic processes, caused by the upwelling of mantle plumes is tackled by Burov and Gerya (2014), who used 3D numerical modelling of mantle lithosphere interaction. Tomographic images show that most plumes are not simple vertical structures. Rather, lithosphere heterogeneities and deep fracture systems all contribute to what could be a very uneven uplift, with local subsidences. All these features need to be taken into consideration, resulting in a range of possible structures. For example, the case of crustal doming associated with the Emeishan flood

volcanism is elegantly proven by detailed sedimentological studies (He et al., 2003). A lack of obvious uplift appears to be the case for the Warakurna large igneous province (LIP) (Wingate et al., 2004), which extends from a central region, westward for some 2000 km in western and central Australia. It contains sill complexes, lavas of bimodal composition, A-type granites and a large layered intrusion. The shape of the LIP and its nature suggest that a mantle plume flowed westward along a crustal and lithospheric discontinuity (Morris and Pirajno, 2005). On the other hand, the apparent lack of crustal doming for the Siberian flood volcanism is by no means well established. In fact, for the Siberian Traps, a huge rift system is present to the west and northwest of the flood lavas, which form a field that extends east and southeast. Thus, the lava field, as seen today, does not coincide with the central zone of the uplift and rifting, which could also be where the plume head impacted (Reichow et al., 2002; Griffin et al., 2005). Cox's concept (Cox, 1989) is a brilliant deduction that well explains many of the essential features of rifting and continental flood volcanism. The important point here, I believe, is that where sea floor spreading develops following rifting, the original uplifted rift shoulders become separated, so that on one side of the rift there is only part of a dome structure, or in other words a dome flank, this appears to be the case of the SW African coastline relating to the uplift of the Paraná–Etendeka LIP. Rift flank uplift is generally asymmetric and uneven, as was probably the case for the Deccan region and as well explained for the Arabian Peninsula (Daradich et al., 2003).

Crustal uplifts are also a common feature of many volcanic rifted margins and these have been attributed to magmatic underplating linked to plume activity (Pirajno, 2000; Nikishin et al., 2002; Ziegler and Cloetingh, 2004). Major crustal uplifts may also be caused by mantle upwelling associated with convective removal of the mantle lithosphere, as may be the case for the Tibetan Plateau. The removal of the lower mantle lithosphere causes the influx of hot mantle asthenospheric material, resulting in a rise of thermal gradients, partial melting of the lithosphere, and uplift of the crust above the buoyant asthenospheric blob. Molnar et al. (1993) suggested this mechanism to explain the uplift of the Tibetan Plateau and the associated potassic volcanism. This may have been the case for the Mianning–Dechang carbonatite belt, discussed in this paper.

Doming of the crust can result in rupture and rifting, with development of triple junctions, as observed in present day Afar region of East Africa (Afar triangle), attributed to a mantle plume impinging beneath the continental lithosphere (Wolfenden et al., 2004; Yirgu et al., 2006; Begg et al., 2009; Beccaluva et al., 2011; Chang and van Der Lee, 2011; Natali et al., 2011). The mantle plume model for the development of the Afar triple junction is well supported by seismic tomography (Grand et al., 1997; van der Hilst et al., 1997; Zhao, 2004, 2007, 2009; Montelli et al., 2006). Several lines of evidence (including seismic tomography) indicate that many giant continental rift systems are related to lithosphere thinning and the following irregular crustal uplifts, ultimately linked to upwelling mantle plumes. The Cenozoic East African Rift System (EARS), including the incipient oceanic arm of the Red Sea, extends for about 5000 km, with a series of graben, half-graben structures, strong uplifts, and igneous activity that includes carbonatites, peralkaline, alkaline and bimodal volcanism, development of lakes and basins with thick accumulations of clastic, chemical and evaporite sediments, as well as active discharge of hydrothermal fluids (Degens and Kulbicki, 1973; Tiercelin et al., 1989; Pflumio et al., 1994; Schlüter, 1997; Renaut et al., 2002; Yirgu et al., 2006).

The breakup of supercontinents, such as Rodinia and Gondwana, is thought to be caused by the impact of mantle plumes, which led to the formation of giant rift systems that eventually led to separation of continental blocks (Şengör, 2001; Conrad and Gurnis, 2003; Santosh et al., 2009; Pirajno and Hoatson, 2012; Xia et al., 2012; Pirajno and Santosh, 2014, in review). In Permo-Triassic times a giant Gondwana intracontinental rift system developed between India and Australia–Antarctica (Harrowfield et al., 2005). This rift system

eventually developed into an ocean resulting in the separation of India from Australia and Antarctica. During the early Cretaceous breakup between India and Australia, flood basalts were erupted between 132 Ma and 119–118 Ma forming the Broken Ridge–Ninety East, Wallaby and Naturaliste LIPs, emplaced on oceanic crust and Kerguelen Plateau (emplaced on a rifted microcontinent), the Rajmahal continental basalts and lamprophyric rocks in India (Coffin et al., 2002; Duncan, 2002; Benard et al., 2010). The Kerguelen plume generated submarine and continental flood magmatism within a part of the Gondwana rift that could be considered as a triple junction between India, Antarctica and Australia, and as such, recording the breakup between Australia and Greater India and Australia and Antarctica (Coffin et al., 2002; Duncan, 2002; Benard et al., 2010; Pirajno and Hoatson, 2012). In the context of this contribution, it is considered possible that the breakup of that part of Gondwana in what is present day South Atlantic, may be related to the St Helena and Tristan da Cunha hotspots (Toyoda et al., 1994; Gibson et al., 2006; Ernst and Bell, 2010), although there are other views (Kamanetsky et al., 2001; Rocha-Júnior et al., 2012, 2013), which do not consider these hotspots to be responsible for the

eruption of the Paraná–Etendeka LIP, and associated alkaline complexes and carbonatites. With respect to the latter, it is interesting to note that Regelous et al. (2009) suggested that the breakup of the South Atlantic occurred as a result of lithospheric delamination, which would have resulted in the eruption of the Paraná–Etendeka LIP, with the Tristan da Cunha volcanic system being related to a modern-day eruptive event following the South Atlantic sea floor spreading. On the other hand, Stica et al. (2014) in their work on the Pelotas Basin situated on a volcanic passive margin, to the west of the Paraná–Etendeka LIP, pointed out that several authors consider the rise of the Tristan da Cunha mantle plume to be associated with the breakup of Gondwana, although other authors (Becker et al., 2012) question the model of the South Atlantic opening as being related to the Paraná–Etendeka LIP.

In any case, these alkaline and carbonatite intrusions were emplaced along major transform and strike-slip fractures (see Fig. 2), on both sides of the rift system, from which South America and southwest Africa drifted. Fig. 30 shows a possible Late-Jurassic–Early Cretaceous reconstruction of Gondwana, with distribution of the hot spots and areas of uplift due to impinging mantle plumes, resulting in alkaline



Fig. 30. A Late Jurassic–Early Cretaceous reconstruction of southern Gondwana showing possible time–space relationship that may have existed between LIPs and hotspots, associated rift systems which led to the breakup of the supercontinent (after Şengör, 2001). Although several other configurations of Gondwana have been published (e.g. Rogers et al., 1995; Segev, 2000; Stampfli et al., 2013; Meert, 2014), the notion of hotspots, rifting, associated alkaline magmatism and breakup remains valid.

and carbonatitic magmatism, final break up and drifting of the land masses (Dalziel et al., 2000; Segev, 2000; Şengör, 2001; Conrad and Gurnis, 2003; Meert and Lieberman, 2008; Yoshida and Santosh, 2011).

Dyke swarms and LIPs ranging in age from about 1.27 to 0.55 Ga of the Rodinia supercontinent heralded the impingement of mantle plumes, which created triple junctions and global rifts systems that would have initiated the breakup of the supercontinent (Dalziel, 1991; Ernst et al., 1996; Ernst and Buchan, 1997; Hanson et al., 1998; Li et al., 1999). These dyke swarms and LIPs are part of a major cluster of intraplate igneous events, which include the 1.27 Ga MacKenzie dyke swarm, the 1.17–1.08 Ga Keeweenaw magmatism of the Mid-Continent rift system both in present-day North America, the 1.1 Ga Umkondo event in the Kalahari craton and the 1.07 Ga Giles event (Warakurna LIP) in central-western Australia (Ernst et al., 2008). The 1.07 Ga Giles event in Australia (Howard et al., 2011), which includes the Warakurna LIP (Wingate et al., 2004) was interpreted by Pirajno and Hoatson (2012) to be related to a mantle plume, which affected zones of lithospheric and crustal weakness between the North Australian and the Western Australian (Yilgarn and Pilbara) cratons. However, this is disputed by Evins et al. (2010) and Smithies et al. (2014) who proposed that the Giles event was triggered by intraplate tectonic processes associated with magmas focussing on a pre-existing lithospheric architecture. In either views, mantle dynamics are involved in the emplacement of the associated Warakurna LIP which covers areas from the west coast of Australia to the centre of the continent extending, approximately along an E–W trend, for a length of about 1500 km, a width of about 800 km, covering an area of at least 1.5×10^6 km². The Warakurna LIP is characterised by mafic–ultramafic layered intrusions and extensive bimodal volcanism in the Musgrave Province, tholeiitic lavas, mafic dykes and dolerite sill complexes in the central and western parts (Capricorn Orogen). The boundary between the Yilgarn and Pilbara cratons, affected by the Warakurna mantle plume event, can be considered as the equivalent of a failed intracontinental rift. The Gifford Creek ferrocarnatite complex with an age of ~1075 Ma, discussed in this contribution, is likely related to the Warakurna mantle plume (Pirajno et al., 2014). If this is correct, the Gifford carbonatites would represent a distal expression of the Giles hotspot, because volatile-rich melts migrate laterally from the hotspot, intersecting the volatile-rich mantle solidus (Ernst and Bell, 2010). Thus, and following Ernst and Bell (2010), carbonatitic and alkaline silicate melts would be restricted to the volatile-rich parts of the lithospheric mantle that are distal from the plume centre. However, it must be pointed out that there are instances in which carbonatite systems can be emplaced early in the magmatic history of a LIP, as seems to be the case for the Bushveld Igneous Complex (Rajesh et al., 2013). But even in this latter case, it remains essential that the lithospheric mantle becomes strongly metasomatised due to the influx of volatiles, in order to produce alkaline and carbonatitic melts.

7. Concluding remarks

Anorogenic and intraplate alkaline magmatism, which includes silica-oversaturated and silica-undersaturated complexes and carbonatites is the spotlight of increasing attention due to its potential to host economically viable critical and rare metals.

Mineralisation in alkaline and carbonatite rocks is formed through at least three principal ore-forming processes, namely: 1) primary magmatic, in which the ore minerals are precipitated from residual fluids in the subsolidus range of temperature and pressure; 2) hydrothermal, from late-stage fluids evolved from (1) and their mixing with heated meteoric waters; hydrothermal mineralisation commonly overprints the primary magmatic; and 3) supergene, usually developed in the weathering profiles of carbonatite in tropical climates and under conditions of high Eh and low pH.

The products of intraplate and rift-related magmatism are volumetrically small when compared to the large granitic batholiths formed at

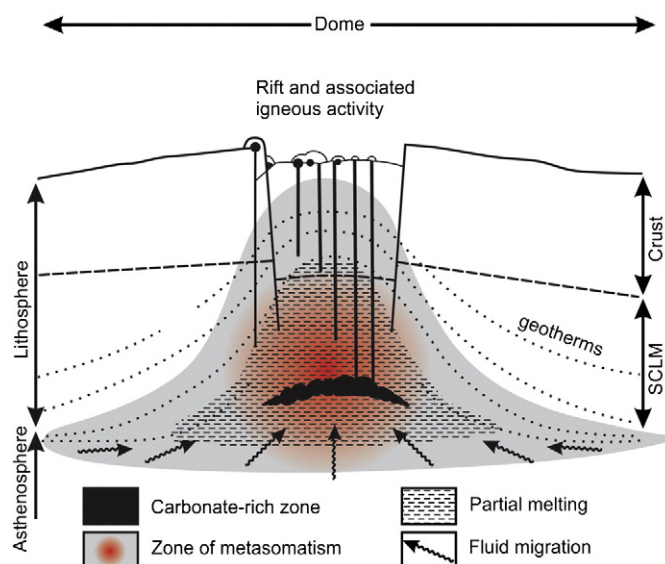


Fig. 31. Model of continental rifting and associated magmatism, according to Woolley (1987) and Martin (2006), showing upward migration of fluids and volatiles (H₂O, F and CO₂ mainly), resulting in partial melting of a metasomatised sub-continental lithospheric mantle, leading to the development of silica-oversaturated and silica-undersaturated complexes and carbonatitic (carbothermal fluid phase) igneous activity.

subduction fronts. Nevertheless, in spite of this, rift-related alkaline magmas and carbonatites provide excellent and most useful insights into mantle and sub-continental lithospheric mantle dynamics.

With this in mind, I have chosen to showcase some of the relevant aspects of intraplate alkaline magmatism, associated carbonatites and related mineral systems. In so doing I have reviewed a selection of anorogenic alkaline plutonic and volcano-plutonic complexes (Brandberg and Erongo in Namibia) and carbonatites (Kruidfontein and Goudini in South Africa, Mount Weld and Gifford Creek in Western Australia, Bayan Obo and Maoniuping in China). The Brandberg and Erongo anorogenic complexes are part of the ~138–130 Ma Paraná–Etendeka LIP, outcropping in present-day South America and southwest Africa and linked to the Tristan da Cunha hotspot. These anorogenic complexes were emplaced along major transform and later strike-slip faults, resulting from the drifting movement of the South American and African plates.

In conclusion, intraplate alkaline magmatism typically with syenitic–nephelinitic melts and associated carbonatites are dominantly emplaced in rift structures formed by the impingement of mantle plumes at the base of the sub-continental mantle lithosphere. This is abundantly shown in the geological record as well as present-day intracontinental rifts (e.g. East African Rift System). The model put forward by Woolley (1987) and re-enforced by Martin (2006) shown in Fig. 31 is considered quite pertinent for the topic treated in this contribution.

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References

- Anand, R.R., 2011. Weathering history, landscape evolution and implications for exploration. In: Butt, C.R.M., Cornelius, M., Scott, K.M., Robertson, I.D.M. (Eds.), *Regolith Expression of Australian Ore Systems*. CRC, LEME, pp. 15–45.
- Bailey, D.K., 1985. Fluids, melts, flowage and styles of eruption in alkaline ultramafic magmatism. *Transaction of the Geological Society of South Africa* 88, 449–457.
- Bailey, D.K., 1987. Mantle metasomatism — perspective and prospect. *Geological Society, London, Special Publication* 30, 1–14.

- Bailey, D.K., Macdonald, R., 1987. Dry peralkaline felsic liquids and carbon dioxide flux through the Kenya rift zone. *Geochemical Society Special Publication* 1, 91–105.
- Bailey, D.K., Cooper, J.P., Knight, J.L., 1974. Anhydrous melting and crystallization of peralkaline obsidians. *Bulletin of Volcanology* 38, 653–665.
- Beccaluva, L., Bianchini, G., Wilson, M. (Eds.), 2011. *Volcanism and Evolution of the African Lithosphere*. The Geological Society of America Special Paper 478 (331 pp.).
- Becker, K., Franke, D., Schnabel, M., Schreckenberger, B., Heyde, I., Krawczyk, C.M., 2012. The crustal structure of the southern Argentine margin. *Geophysical Journal International* 189, 1483–1504.
- Begg, G.C., Griffin, W.L., Natapov, L.M., O'Reilly, S.Y., Grand, S.P., O'Neill, C.J., Hronsky, J.M.A., Poudjom Djomani, Y., Swain, C.J., Deen, T., Bowden, P., 2009. The lithospheric architecture of Africa: seismic tomography, mantle petrology, and tectonic evolution. *Geosphere* 5 (1), 23–50.
- Bell, K., 2001. Carbonatites: relationships to mantle plume activity. *Geological Society of America, Special Paper* 352, 267–290.
- Bell, K., Simonetti, A., 1996. Carbonatite magmatism and plume activity: implications from the Nd, Pb and Sr isotope systematics of Oldoinyo Lengai. *Journal of Petrology* 37, 1321–1339.
- Bell, K., Simonetti, A., 2010. Source of parental melts to carbonatite—critical isotopic constraints. *Mineralogy and Petrology* 98, 77–89.
- Bell, K., Lavecchia, G., Rosatelli, G., 2013. Cenozoic Italian magmatism — isotope constraints for possible plume-related activity. *Journal of South American Earth Sciences* 41, 22–40.
- Benard, F., Callot, J.-P., Vially, R., Schmitz, J., Roest, W., Patriat, M., Loubrieu, B., ExtraPlac Team, 2010. The Kerguelen Plateau: records from a long-living/composite microcontinent. *Marine and Petroleum Geology* 27, 633–649.
- Bonin, B., 1986. *Ring Complex Granites and Anorogenic Magmatism*. North Oxford Academic (189 pp.).
- Bowden, P., 1985. The geochemistry and mineralization of alkaline ring complexes in Africa (a review). *Journal of African Earth Sciences* 3, 17–40.
- Bowden, P., Kinnaird, J.A., Diehl, M., Pirajno, F., 1990. Anorogenic granite evolution in Namibia — a fluid contribution. *Geological Journal* 25, 381–390.
- Burke, K., Dewey, J.F., 1973. Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. *Journal of Geology* 81, 406–433.
- Burov, E., Gerya, T., 2014. Asymmetric three-dimensional topography over mantle plumes. *Nature* 513, 85–89.
- Butt, C.R.M., 1989. Genesis of supergene gold deposits in the lateritic regolith of the Yilgarn Block, Western Australia. *Economic Geology Monograph* 6, 11–29.
- Campbell, I.H., Griffiths, R.W., 1990. Implications of mantle plume structure for the evolution of flood basalts. *Earth and Planetary Science Letters* 99, 79–93.
- Castor, S.B., 2008. The Mountain Pass rare-earth carbonate and associated ultrapotassic rocks, California. *Canadian Mineralogist* 46, 779–806.
- Chakhmouradian, A.R., Wall, F., 2012. Rare earth elements: minerals, mines, magnets (and more). *Elements* 8, 333–340.
- Chakhmouradian, A.R., Smith, M.P., Kynicky, J., 2014. From “strategic” tungsten to “green” neodymium: a century of critical metals at a glance (Editorial). *Ore Geology Reviews* <http://dx.doi.org/10.1016/j.oregeorev.2014.06.008>.
- Chang, S.J., Van Der Lee, S., 2011. Mantle plumes and associated flow beneath Arabia and East Africa. *Earth and Planetary Science Letters* 302, 448–454.
- Chao, E.C.T., Back, J.M., Minkin, J.A., Tatsumoto, M., Wang, J.W., Conrad, J.E., McKee, E.H., Hou, Z.L., Meng, Q.R., Huang, S.G., 1997. The sedimentary carbonate-hosted giant Bayan Obo REE–Fe–Nb ore deposit of Inner Mongolia, China: a cornerstone example for giant polymetallic ore deposits of hydrothermal origin. *United States Geological Survey Bulletin* 2143.
- Clarke, L.B., 1989. The Geology of the Kruidfontein Volcanic Complex, Transvaal, S. Africa. Unpubl. PhD Thesis, University of Leicester, 167 pp.
- Clarke, L.B., Le Bas, M.J., 1990. Magma mixing and metasomatic reaction in silicate-carbonate liquids at the Kruidfontein carbonatitic volcanic complex, Transvaal. *Mineralogical Magazine* 54, 45–56.
- Coffin, M.F., Pringle, M.S., Duncan, R.A., Gladchenko, T.P., Storey, M., Müller, R.D., Ghagan, L.A., 2002. Kerguelen hotspot magma output since 130 Ma. *Journal of Petrology* 26, 1121–1139.
- Condie, K.C., 1997. *Plate Tectonics and Crustal Evolution*, 4th edn. Butterworth Heinemann, Oxford (282 pp.).
- Conrad, C.P., Gurnis, M., 2003. Seismic tomography, surface uplift, and the breakup of Gondwanaland: integrating mantle convection backwards in time. *Geochemistry, Geophysics, Geosystems* 4 (3). <http://dx.doi.org/10.1029/2001GC000299>.
- Cook, N.J., Ciobanu, C.L., O'Reilly, D., Wilson, R., Das, K., Wade, B., 2013. Mineral chemistry of Rare Earth Element (REE) mineralization, Browns Ranges, Western Australia. *Lithos* 172–173, 192–213.
- Cox, K.G., 1989. The role of mantle plumes in the development of continental drainage patterns. *Nature* 342, 873–876.
- Crocker, I.T., Martini, J.E.J., Sonhge, A.P.G., 1988. The fluorite deposits of the Republics of South Africa and Bophuthaswana. *Handbook Geological Survey of South Africa* 11 (172 pp.).
- Dalziel, I.W.D., 1991. Pacific margins of Laurentia and East Antarctica–Australia as a conjugate rift pair: evidence and implications for an Eocambrian supercontinent. *Geology* 19, 598–601.
- Dalziel, I.W.D., Lawver, L.A., Murphy, J.B., 2000. Plumes, orogenesis and supercontinental fragmentation. *Earth and Planetary Science Letters* 178, 1–11.
- Daradich, A., Mitrovica, J.X., Pysklyvec, R.N., Willett, R.N., Forte, A.M., 2003. Mantle flow, dynamic topography and rift flank uplift of Arabia. *Geology* 31, 901–904.
- Degens, E.T., Kulbicki, G., 1973. Hydrothermal origin of metals in some East African lakes. *Mineralium Deposita* 8, 388–404.
- Diehl, M., 1989. *Geology, Mineralogy, Geochemistry and Hydrothermal Alteration of the Brandberg Alkaline Complex, Namibia*. PhD Thesis (unpubl.) J Gutenberg Universität, Mainz, Germany.
- Diehl, M., 1990. *Geology, mineralogy, geochemistry and hydrothermal alteration of the Brandberg alkaline complex, Namibia*. Geological Survey Namibia Memoir 10, 1–55.
- Drew, L.J., Meng, Q., Sun, W., 1990. The Bayan Obo iron–rare earth–niobium deposits, Inner Mongolia, China. *Lithos* 26, 43–65.
- Duncan, R.A., 2002. A time frame for construction of the Kerguelen Plateau and Broken Ridge. *Journal of Petrology* 43, 1109–1119.
- Duncan, R.K., Willett, G.C., 1990. Mount Weld carbonatite. In: Hughes, F.E. (Ed.), *Geology and Mineral Deposits of Australia and Papua New Guinea*. Australasian Institute of Mining and Metallurgy, Monograph 14, pp. 591–597.
- Eales, H.V., Cawthorn, R.G., 1996. The Bushveld Complex. In: Cawthorn, R.G. (Ed.), *Layered Intrusions*. Elsevier, Amsterdam, pp. 181–230.
- Edgar, A.D., 1987. The genesis of alkaline magmas with emphasis on their source region: inferences from experimental studies. In: Fitton, J.G., Upton, B.G.J. (Eds.), *Alkaline Igneous Rocks*. Geological Society Special Publication 30, pp. 29–54.
- Ernst, R.E., Bell, K., 2010. Large igneous provinces (LIPs) and carbonatites. *Mineralogy and Petrology* 98, 55–76.
- Ernst, R.E., Buchan, K.L., 1997. Giant radiating dyke swarms: their use in identifying pre-Mesozoic large igneous provinces and mantle plumes. In: Mahoney, J.J., Coffin, M.F. (Eds.), *Large Igneous Provinces — Continental, Oceanic and Planetary Flood Volcanism*. American Geophysical Union, Monograph 100, pp. 297–334.
- Ernst, R.E., Buchan, K.L., 2001. Large mafic magmatic event through time and links to mantle plume heads. *Geological Society of America, Special Paper* 352, 483–575.
- Ernst, R.E., Buchan, K.L., West, T.D., Palmer, H.C., 1996. Diabase (Dolerite) dyke swarms of the world. *Geological Survey of Canada Open File* 32411st edition.
- Ernst, R.E., Wingate, M.T.D., Buchan, K.L., Li, Z.X., 2008. Global record of 1600–700 Ma Large Igneous Provinces (LIPs): implications for the reconstruction of the proposed Nuna (Columbia) and Rodinia supercontinents. *Precambrian Research* 160, 159–178.
- Evins, P., Smithies, H., Howard, H., Kirkland, C., Wingate, M.T.D., Bodorkos, S., 2010. Devil in the detail; the 1150–1000 Ma magmatic and structural evolution of the Ngaanyatjarra Rift, West Musgrave province, Central Australia. *Precambrian Research* 183, 572–588.
- Fan, H.R., Xie, Y.H., Wang, K.Y., Tao, K.J., Wilde, S.A., 2004a. REE daughter minerals trapped in fluid inclusions in the giant Bayan Obo REE–Nb–Fe deposit, Inner Mongolia, China. *International Geology Review* 46, 638–645.
- Fan, H.R., Xie, Y.H., Wang, K.Y., Wilde, S.A., 2004b. Methane-rich fluid inclusions in skarn near the giant REE–Nb–Fe deposit at Bayan Obo, northern China. *Ore Geology Reviews* 25, 301–309.
- Fan, H.R., Hu, F.F., Yang, K.F., Liu, X., Wang, K.Y., Pirajno, F., 2014. Integrated U–Pb and Sm–Nd geochronology for a REE-rich carbonatite dyke at the giant Bayan Obo REE deposit, northern China. *Ore Geology Reviews* 63, 510–519.
- Fischer, T.P., Burnard, P., Marty, B., Hilton, D.R., Furi, E., Palhol, F., Sharp, Z.D., Mangasini, F., 2009. Upper-mantle volatile chemistry at Oldoinyo Lengai volcano and the origin of carbonatites. *Nature* 459, 77–80.
- Fitton, J.G., Upton, B.G.J. (Eds.), 1987. *Alkaline Igneous Rocks*. Geological Society Special Publication 30. Blackwell Scientific Publications (568 pp.).
- Flint, D., Abeyasinghe, P.B., 2000. *Geology and Mineral Resources of the Gascoyne Region*. Geological Survey of Western Australia, Record 2000/7.
- Fourel, L., Milelli, L., Jaupart, C., Limare, A., 2013. Generation of continental rifts, basins, and swells by lithosphere instabilities. *Journal of Geophysical Research, Solid Earth* 118, 1–21.
- Genge, M.J., Price, G.D., Jones, A.P., 1995. Molecular dynamics simulation of CaCO₃ melts to mantle pressures and temperatures: implications for carbonatite magmas. *Earth and Planetary Science Letters* 131, 225–238.
- Gibson, S.A., Thompson, R.N., Day, J.A., 2006. Timescales and mechanisms of plume–lithosphere interactions: ⁴⁰Ar/³⁹Ar geochronology and geochemistry of alkaline igneous rocks from the Paraná–Etendeka large igneous province. *Earth and Planetary Science Letters* 251, 1–17.
- Graham, S., Lambert, D., Shee, S., 2004. The petrogenesis of carbonatite, melonite and kimberlite from the Eastern Goldfield Province, Yilgarn Craton. *Lithos* 76, 519–533.
- Grand, S.P., van der Hilst, R.D., Widiyantoro, S., 1997. Global seismic tomography: a snapshot of convection in the Earth. *GSA Today* 7 (4), 1–7.
- Griffin, W.L., Natapov, L.M., O'Reilly, S.Y., van Acherbergh, E., Cherenkova, A.F., Cherenkov, V.G., 2005. The Kharantai kimberlite field, Siberia: modification of the lithospheric mantle by the Siberian Trap event. *Lithos* 81, 167–187.
- Griffiths, R.W., Campbell, I.H., 1990. Stirring and structure in mantle plumes. *Earth and Planetary Science Letters* 99, 66–78.
- Groves, D.I., Bierlein, F.P., Meinert, L.D., Hitzman, L.W., 2010. Iron Oxide Copper–Gold (IOCG) deposits through Earth history: implications for origin, lithospheric setting, and distinction from other epigenetic iron oxide deposits. *Economic Geology* 105, 641–654.
- Gudfinsson, G.H., Presnall, D.C., 2005. Continuous gradations among primary carbonatitic, kimberlitic, melilitic, basaltic, picritic and komatiitic melts in equilibrium with garnet ilherzolite at 3–8 GPa. *Journal of Petrology* 46, 1645–1659.
- Guo, Z.F., Hertogen, J., Liu, J.Q., Pastels, P., Boven, A., Punzalan, L., He, H.Y., Luo, X.J., Zhang, W.H., 2005. Potassic magmatism in western Sichuan and Yunnan provinces, SE Tibet, China: petrological and geochemical constraints. *Journal of Petrology* 46, 33–78.
- Hanson, R.E., Martin, M.W., Bowring, S.A., Munyanywa, H., 1998. U–Pb zircon age for the Umkondo dolerites, eastern Zimbabwe: 1.1 Ga large igneous province in southern Africa–East Antarctica and possible Rodinia correlations. *Geology* 26, 1143–1146.
- Harrowfield, M., Holdgate, G.R., Wilson, C.J.L., McLoughlin, S., 2005. Tectonic significance of the Lambert graben, East Antarctica: reconstructing the Gondwana rift. *Geology* 33, 197–200.
- Hastings Rare Metals Ltd, 2012. Australian Stock Exchange (ASX) Announcement, 29th March 2012 (unpublished); <http://www.hastingsraremetals.com/projects/yangibana-project/previous-exploration/>.

- He, B., Xu, Y.-G., Chung, S.-L., Xiao, L., Wang, Y., 2003. Sedimentary evidence for a rapid, kilometer-scale crustal doming prior to the eruption of the Emeishan flood basalts. *Earth and Planetary Science Letters* 213, 391–405.
- Hoatson, D.M., Jaireth, S., Miezi, S., 2011. The major rare-earth element deposits of Australia: geological setting, exploration and resources. *Geoscience Australia Report*, (204 pp.).
- Hodgson, F.D.I., 1973. Petrography of the Brandberg intrusion, South West Africa. *Special Publication Geological Society of South Africa* 3, 339–342.
- Hou, Z.Q., Ma, H.W., Zaw, K., Zhang, Y.Q., Wang, M.J., Wang, Z., Guitang, P., Renli, T., 2003. The Himalayan Yulong porphyry copper belt: product of large scale strike-slip faulting in eastern Tibet. *Economic Geology* 98, 125–146.
- Hou, G., Santosh, M., Qian, X., Lister, G., Li, J., 2008. Configuration of the Late Paleoproterozoic supercontinent Columbia: insights from radiating mafic dyke swarms. *Gondwana Research* 14, 395–409.
- Hou, Z.Q., Tian, S.H., Xie, Y.L., Yang, Z.S., Yuan, Z.X., Yin, S.P., Yi, L.S., Fei, H.C., Bai, G., Li, X.Y., 2009. The Himalayan Mianing–Dechang REE belt associated with carbonatite–alkaline complexes, eastern Indo-Asian collision zone, SW China. *Ore Geology Reviews* 36, 65–89.
- Howard, H.M., Smithies, R.H., Evins, P.M., Kirkland, C.L., Werner, M., Wingate, M.T.D., Pirajno, F., 2011. Explanatory notes for the West Musgrave Province. *Geological Survey of Western Australia, Record* 2011/4, (349 pp.).
- Jaireth, S., Hoatson, D.M., 2014. Geological setting and resources of the major rare-earth element deposits in Australia. *Ore Geology Reviews* <http://dx.doi.org/10.1016/j.oregeorev.2014.02.008>.
- Johnson, S.P., Sheppard, S., Wingate, M.T.D., Kirkland, C.L., Belousova, E.A., 2011a. Emporal and hafnium isotopic evolution of the Glenburgh terrane basement: and exotic crustal fragment in the Capricorn Orogen. *Geological Survey of Western Australia, Report* 110, (27 pp.).
- Johnson, S.P., Sheppard, S., Rasmussen, B., Wingate, M.T.D., Kirkland, C.L., Muhling, J.R., Fletcher, I.R., Belousova, E.A., 2011b. Two collisions, two sutures: punctuated pre-1950 Ma assembly of the West Australian Craton during the Ophthalmanian and Glenburgh Orogenies. *Precambrian Research* 189 (3–4), 239–262.
- Johnson, S.P., Thorne, A.M., Cutten, H.N., Blay, O., 2012. Geological interpretation of the western Capricorn Orogen. *Geological Survey of Western Australia Record* 2011/25 (114 pp.).
- Johnson, S.P., Thorne, A.M., Tyler, I.M., Korsch, R.J., Kennett, B.L.N., Cutten, H.N., Goodwin, J., Blay, O., Blewett, R.S., Joly, A., Dentith, M.C., Aitken, A.R.A., Holzschuh, J., Salmon, M., Reading, A., Heinison, G., Boren, G., Ross, J.R.D., Costello, R.D., Fomin, T., 2013. Crustal architecture of the Capricorn Orogen, Western Australia and associated metallogeny. *Australian Journal of Earth Sciences* 60, 681–705.
- Kamanetsky, V.S., Maas, R., Sushchevskaya, N.M., Norman, M.D., Cartwright, I., Peyve, A.A., 2001. Remnants of Gondwanan continental lithosphere in oceanic upper mantle: evidence from the South Atlantic Ridge. *Geology* 29, 243–246.
- Kampunzu, A.B., Lubala, R.T. (Eds.), 1991. *Magmatism in Extensional Structural Settings – The Phanerozoic African Plate*. Springer-Verlag (637 pp.).
- Kinnaird, J.A., Bowden, P., 1991. Magmatism and mineralization associated with Phanerozoic anorogenic plutonic complexes of the African plate. In: Kampunzu, A.B., Lubala, R.T. (Eds.), *Magmatism in Extensional Structural Settings – The Phanerozoic African Plate*. Springer-Verlag, pp. 410–485.
- Kjarsgaard, B.A., Peterson, T., 1991. Nephelinite–carbonatite liquid immiscibility at Shombole volcano, East Africa: petrographic and experimental evidence. *Mineralogy and Petrology* 43, 293–314.
- Kogarko, L.N., Lahaye, Y., Brey, G.P., 2010. Plume-related mantle source of super-large rare metal deposits from the Lovozero and Khibina massifs on the Kola peninsula, eastern part of the Baltic shield: Sr, Nd and Hf isotope systematics. *Mineralogy and Petrology* 98 (1–4), 197–208.
- Le Bas, M.J., 1989. Diversification of carbonatite. In: Bell, K. (Ed.), *Carbonatites. Genesis and Evolution*. Unwin Hyman Ltd, pp. 428–445.
- Le Bas, M.J., Yang, X.M., Taylor, R.N., Spiro, B., Milton, J.A., Peishan, Z., 2007. New evidence from a calcite–dolomite carbonatite dyke for the magmatic origin of the massive Bayan Obo ore-bearing dolomite marble, Inner Mongolia, China. *Mineralogy and Petrology* 91, 281–307.
- Le Maitre, R.W. (Ed.), 2002. *Igneous Rocks – A Classification and Glossary of Terms*. Cambridge University Press, Cambridge (236 pp.).
- Lentz, D.R., 1999. Carbonatite genesis: a re-examination of the role of intrusion-related pneumatolytic skarn processes in limestone melting. *Geology* 27 (4), 335–338.
- Li, Z.X., Li, Z.H., Kinny, P.D., Wang, J., 1999. The breakup of Rodinia: did it start with a mantle plume beneath South China? *Earth and Planetary Science Letters* 173, 171–181.
- Liu, Y.L., Yang, G., Chen, J.F., Du, A.D., Xie, Z., 2004. Re–Os dating of pyrite from the giant Bayan Obo REE–Nb–Fe deposit. *Chinese Science Bulletin* 49, 2627–2631.
- Lottermoser, B.G., 1990. Rare-earth element mineralisation within the Mt Weld carbonatite laterite, Western Australia. In: Hughes, F.E. (Ed.), *Geology and Mineral Deposits of Australia and Papua New Guinea*. Australasian Institute of Mining and Metallurgy, Monograph 14, pp. 151–167.
- Lowenstern, J.B., 2001. Carbon dioxide in magmas and implications for hydrothermal systems. *Mineralium Deposita* 36, 490–502.
- Mariano, A.N., 1989. Economic geology of rare earth elements. In: Lipin, B.R., McKay, G.A. (Eds.), *Geochemistry and Mineralogy of Rare Earth Elements*. Reviews in Mineralogy 21, pp. 309–337.
- Marsh, J.S., 1973. Relationships between transform directions and alkaline igneous rock lineaments in Africa and South America. *Earth and Planetary Science Letters* 18, 317–323.
- Martin, R.F., 2006. A-type granites of crustal origin ultimately result from open-system fenitization-type reactions in an extensional environment. *Lithos* 91, 125–136.
- Martin, D.M.B., Sheppard, S., Thorne, A.M., Farrell, T.R., Groenewald, P.B., 2006. Proterozoic geology of the western Capricorn Orogen – a field guide. *Geological Survey of Western Australia, Record* 2006/18, p. 41.
- Meert, J.G., 2014. Strange attractors, spiritual interlopers and lonely wanderers: the search for pre-Pangean supercontinents. *Geoscience Frontiers* 5, 155–166.
- Meert, J.G., Lieberman, B.S., 2008. The Neoproterozoic assembly of Gondwana and its relationship to the Ediacaran–Cambrian radiation. *Gondwana Research* 14, 5–21.
- Molnar, P., England, P., Martinod, J., 1993. Mantle dynamics, uplift of the Tibetan Plateau and the Indian Monsoon. *Reviews of Geophysics* 31, 357–396.
- Montelli, R., Nolet, G., Dahlen, F.A., Masters, G., 2006. A catalogue of deep mantle plumes: new results from finite-frequency tomography. *Geochemistry, Geophysics, Geosystems* 7 (11), Q11007. <http://dx.doi.org/10.1029/2006GC001248>.
- Morris, P.A., Pirajno, F., 2005. *Geology, geochemistry, and mineralization potential of Mesoproterozoic sill complexes of the Bangemall Supergroup, Western Australia*. Geological Survey of Western Australia, Report 99, (75 pp.).
- Mutschler, F.E., Griffin, M.E., Scott Stevens, D., Shannon, S.S., 1985. Precious metal deposits related to alkaline rocks in the North American Cordillera: an interpretive review. *Transactions of the Geological Society of South Africa* 88, 355–377.
- Natali, C., Beccaluva, L., Bianchini, G., Siena, F., 2011. Rhyolites associated to Ethiopian CFB: Clues for initial rifting at the Afar plume axis. *Earth and Planetary Science Letters* 312, 59–68.
- Nelson, D.R., Chivas, A.R., Chappell, B.W., McCulloch, M.T., 1988. Geochemical and isotope systematics in carbonatites and implications for the evolution of ocean–island sources. *Geochimica et Cosmochimica Acta* 52, 1–17.
- Ngwenya, B.T., 1994. Hydrothermal rare earth mineralisation in carbonatites of the Tundulu complex, Malawi: processes at the fluid/rock interface. *Geochimica et Cosmochimica Acta* 58, 2061–2072.
- Nielsen, T.F.D., Veksler, I.V., 2002. Is natrocarbonatite a cognate fluid condensate? *Contributions to Mineralogy and Petrology* 142, 425–435.
- Nikishin, A.M., Ziegler, P.A., Abbott, D., Brunet, M.-F., Cloetingh, S., 2002. Permo-Triassic intraplate magmatism and rifting in Eurasia: implications for mantle plumes and mantle dynamics. *Tectonophysics* 351, 3–39.
- Olsen, K.H. (Ed.), 1995. *Continental Rifts: Evolution, Structure, Tectonics*. Development in Geotectonics 25. Elsevier Science, Amsterdam (449 pp.).
- Owen-Smith, T.M., Ashwal, L.D., Torsvik, T.H., Ganeröd, M., Nebel, O., Webb, S.J., Werner, S.C., 2013. Seychelles alkaline suite records the culmination of Deccan Traps continental flood volcanism. *Lithos* 182–183, 33–47.
- Pearson, J.M., 1996. *Alkaline Rocks of the Gifford Creek Complex, Gascoyne Province, Western Australia: Their Petrogenetic and Tectonic Significance*. PhD thesis (unpublished), University of Western Australia, Department of Geology and Geophysics, 286 pp.
- Pearson, J.M., Taylor, W.R., 1996. Mineralogy and geochemistry of fenitized alkaline ultrabasic sills of the Gifford Creek Complex, Gascoyne Province, Western Australia. *The Canadian Mineralogist* 34, 201–219.
- Pearson, J.M., Taylor, W.R., Barley, M.E., 1996. *Geology of the alkaline Gifford Creek Complex, Gascoyne Complex, Western Australia*. Australian Journal of Earth Sciences 43, 299–309.
- Peng, P., Zhai, M., Zhang, H.F., Guo, J.H., 2005. Geochronological constraints on the Paleoproterozoic evolution of the North China Craton: SHRIMP zircon ages of different types of mafic dikes. *International Geology Review* 47, 492–508.
- Peng, P., Zhai, M.G., Guo, J.H., Zhao, T.P., Liu, F., Hu, B., 2008. A 1.78 Ga large igneous province consisting of the North China dyke swarm and Xiong'er Volcanic Province. *Lithos* 101, 260–280.
- Pflumio, C., Boulègue, J., Tiercelin, J.-J., 1994. Hydrothermal activity in the northern Tanganyika Rift, East Africa. *Chemical Geology* 116, 85–109.
- Pirajno, F., 1987. A fossil hot spring systems in the Brandberg Complex, Damara Province, South West Africa/Namibia. *South African Journal of Geology* 90 (4), 509–513.
- Pirajno, F., 1990. *Geology, geochemistry and mineralisation of the Erongo Volcanic Complex, Namibia*. South African Journal of Geology 93 (3), 485–504.
- Pirajno, F., 1994. Mineral resources of anorogenic alkaline complexes in Namibia: a review. *Australian Journal of Earth Sciences* 41, 157–168.
- Pirajno, F., Butt, C.R.M., Winter, R., 1995. Gold enrichment in weathered carbonatite pyroclastics of the Kruidfontein Volcanic Complex, South Africa. *South African Journal of Geology* 98 (3), 319–325.
- Pirajno, F., 2000. *Ore Deposits and Mantle Plumes*. Kluwer Academic Publishers (distributed by Springer), (556 pp.).
- Pirajno, F., 2009. *Hydrothermal Processes and Mineral Systems*. Springer, Berlin (1250 pp.).
- Pirajno, F., 2013a. Effects of metasomatism on mineral systems and their host rocks: alkali metasomatism, skarns, greisens, tourmalinites, rodingites, black-wall alteration and listvenites. In: Harlov, D.E., Austrheim, H. (Eds.), *Metasomatism and Metamorphism: The Role of Fluids in Crustal and Upper Mantle Processes*. Lecture Series in Earth Science. Springer, pp. 203–252.
- Pirajno, F., 2013b. *The Geology and Tectonic Settings of China's Mineral Deposits*. Springer, Dordrecht (679 pp.).
- Pirajno, F., Chen, Y.J., 2005. The Xiong'er Group: a 1.76 Ga large igneous province in east-central China? www.largeigneousprovinces.org/ (June LIP of the Month).
- Pirajno, F., González-Álvarez, I., 2013. The ironstone veins of the Gifford Creek ferrocarbonatite complex, Gascoyne Province. *Geological Survey of Western Australia Record* 2013/12, (16 pp.).
- Pirajno, F., Hoatson, D.M., 2012. A review of Australia's Large Igneous Provinces and associated mineral systems: implications for mantle dynamics through geological time. *Ore Geology Reviews* 48, 2–54.
- Pirajno, F., Santosh, M., 2014. Rifting, intraplate magmatism, mineral systems and mantle dynamics in central Eurasia: an overview. *Ore Geology Reviews* 63, 265–295.
- Pirajno, F., Santosh, M., 2014w. Mantle plumes, supercontinents, intracontinental rifting and mineral systems. *Precambrian Research* (in review).
- Pirajno, F., Schlögl, H.U., 1987. The alteration–mineralization of the Kratztzberg tungsten deposit, South West Africa/Namibia. *South African Journal of Geology* 90 (4), 499–508.

- Pirajno, F., Phillips, D., Armstrong, R.A., 2000. Volcanology and eruptive histories of the Erongo Volcanic Complex and the Paresis Igneous Complex, Namibia: implications for mineral deposit styles. *Communications of the Geological Survey of Namibia* 12, 301–312.
- Pirajno, F., González-Álvarez, I., Chen, W., Simonetti, A., Kyser, K., leGrass, M., 2014. The Gifford Creek ferrocarnatite complex, Gascoyne Province, Western Australia: associated fenitic alteration and a putative link with the ~1075 Ma Warakurna LIP. *Lithos* 202–203, 100–119.
- Rajesh, H.M., Chisonga, B.C., Shindo, K., Beukes, N.J., Armstrong, R.A., 2013. Petrographic, geochemical and SHRIMP U–Pb titanite age characterization of the Thabazimbi mafic sills: extended time frame and a unifying petrogenetic model for the Bushveld Large Igneous Province. *Precambrian Research* 230, 79–102.
- Regelous, M., Niu, Y., Abouchami, W., Castillo, P.R., 2009. Shallow origin for the South Atlantic Dupal Anomaly from lower continental crust: geochemical evidence from the Mid-Atlantic Ridge at 26° S. *Lithos* 112, 57–72.
- Reichow, M.K., Saunders, A.D., White, R.V., Pringle, M.S., Al'Mukhamedov, A.I., Medvedev, A.I., Kirda, N.P., 2002. ⁴⁰Ar/³⁹Ar dates from the West Siberian Basin: Siberian flood basalt province doubled. *Science* 296, 1846–1849.
- Renaut, R.W., Jones, B., Tiercelin, J.-J., Traits, C., 2002. Sublacustrine precipitation of hydrothermal silica in rift lakes: evidence from Lake Baringo, central Kenya Rift Valley. *Sedimentary Geology* 148, 235–257.
- Renne, P.R., Glen, J.M., Milner, S.C., Duncan, A.R., 1996. Age of the Etendeka flood volcanism and associated intrusions in southwestern Africa. *Geology* 24, 659–662.
- Rocha-Júnior, E.R.V., Marques, L.S., Babinski, M., Nardy, A.J.R., Figueiredo, A.M.G., Machado, F.B., 2013. Sr–Nd–Pb isotopic constraints on the nature of mantle sources involved in the genesis of the high-Ti tholeiites from northern Paraná continental flood basalts (Brazil). *Journal of South American Earth Sciences* 46, 9–25.
- Rocha-Júnior, E.R.V., Puchtel, I.S., Marques, L.S., Walker, R.J., Machado, F.B., Nardy, A.J.R., Babinski, M., Figueiredo, A.M.G., 2012. Re–Os isotope and highly siderophile element systematics of the Paraná continental flood basalts (Brazil). *Earth and Planetary Science Letters* 337–338, 164–173.
- Rogers, J.J.W., Santosh, M., 2002. Configuration of Columbia, a Mesoproterozoic supercontinent. *Gondwana Research* 5 (1), 5–22.
- Rogers, J.J.W., Unrug, R., Sultan, M., 1995. Tectonic assembly of Gondwana. *Journal of Geodynamics* 19 (1), 1–34.
- Rogie, J.D., Kerrck, D.M., Chiodini, G., Frondini, F., 2000. Flux measurements of nonvolcanic CO₂ emission from some vents in central Italy. *Journal of Geophysical Research* 105 (B4), 8435–8445.
- Santosh, M., Zhao, G., 2009. Supercontinent dynamics (editorial). *Gondwana Research* 15, 225–227.
- Santosh, M., Maruyama, S., Yamamoto, S., 2009. The making and breaking of supercontinents: some speculations based on superplumes, super downwelling and the role of the tectosphere. *Gondwana Research* 15, 324–341.
- Schlüter, T., 1997. *Geology of East Africa*. Gerbrüder Borntraeger, Berlin.
- Schmitt, A.K., Emmertman, R., Trumbull, R.B., Bühn, B., Henjes-Kunst, F., 2000. Petrogenesis and ⁴⁰Ar/³⁹Ar geochronology of the Brandberg Complex, Namibia: evidence for a major mantle contribution in metaluminous and peralkaline granites. *Journal of Petrology* 41 (8), 1207–1239.
- Segev, A., 2000. Synchronous magmatic cycles during fragmentation of Gondwana: radiometric ages from the Levant and other provinces. *Tectonophysics* 325, 257–277.
- Seltmann, R., Soloviev, S., Shatov, V., Pirajno, F., Naumov, E., Cherkasov, S., 2010. Metallogeny of Siberia: tectonic, geologic and metallogenic settings of selected significant deposits. *Australian Journal of Earth Sciences* 57, 655–706.
- Şengör, A.M.C., 2001. Elevation as indicator of mantle-plume activity. In: Ernst, R.E., Buchan, K.L. (Eds.), *Mantle Plumes: Their Identification Through Time*. Geological Society of America Special Paper 352, pp. 183–225.
- Şengör, A.M.C., Natal'in, B.A., 2001. Rifts of the world. In: Ernst, R.E., Buchan, K.L. (Eds.), *Mantle Plumes: Their Identification Through Time*. Geological Society of America Special Paper 352, pp. 389–482.
- Sheppard, S., Rasmussen, B., Muhling, J.R., Farrell, T.R., Fletcher, I.R., 2007. Grenvillian-aged orogenesis in the Paleoproterozoic Gascoyne Complex, Western Australia: 1030–950 Ma reworking of the Proterozoic Capricorn orogeny. *Journal of Metamorphic Geology* 25, 477–494.
- Sheppard, S., Johnson, S.P., Wingate, M.T.D., Kirkland, C.L., Pirajno, F., 2010. Explanatory notes for the Gascoyne Province. Geological Survey of Western Australia (336 pp.).
- Sinclair, W.D., Richardson, J.M., 1992. Quartz–tourmaline orbicules in the Seagull batholith, Yukon Territory. *Canadian Mineralogist* 30, 923–935.
- Slack, J.F., 1996. Tourmaline associations with hydrothermal ore deposits. *Reviews in Mineralogy* 33, 560–640.
- Smith, M.P., 2007. Metasomatic silicate chemistry at the Bayan Obo Fe–REE–Nb deposit, Inner Mongolia, China: contrasting chemistry and evolution of fenitising and mineralising fluids. *Lithos* 93, 126–148.
- Smith, M.P., Henderson, P., 2000. Preliminary fluid inclusion constraints on fluid evolution in the Bayan Obo Fe–REE–Nb deposit, Inner Mongolia, China. *Economic Geology* 95, 1371–1388.
- Smith, M.P., Henderson, P., Campbell, L.S., 2000. Fractionation of the REE during hydrothermal processes: constraints from the Bayan Obo Fe–REE–Nb deposit, Inner Mongolia, China. *Geochimica et Cosmochimica Acta* 64, 3141–3316.
- Smithies, R.H., Kirkland, C.L., Korhonen, F.J., Aitken, A.R.A., Howard, H.M., Maier, W.D., Wingate, M.T.D., Quentin de Gromard, R., Gessner, K., 2014. The Mesoproterozoic thermal evolution of the Musgrave Province in central Australia – plume vs. the geological record. *Gondwana Research* <http://dx.doi.org/10.1016/j.gr.2013.12.014>.
- Solovova, I.P., Giris, A.V., Kogarko, L.N., Kononkova, N.N., Stoppa, F., Rosatelli, G., 2005. Composition of magmas and carbonate–liquid immiscibility in the Vulture alkaline igneous complex, Italy. *Lithos* 85, 113–128.
- Stampfli, G.M., Hochard, C., Vêrard, C., vonRaumer, W.J., 2013. The formation of Pangea. *Tectonophysics* 593, 1–19.
- Stica, J.M., Zalán, P.V., Ferrari, A.L., 2014. The evolution of rifting on the volcanic margin of the Pelotas Basin and the contextualization of the Paraná–Etendeka LIP in the separation of Gondwana in the South Atlantic. *Marine Petroleum and Geology* 50, 1–21.
- Stoppa, F., Rosatelli, G., Wall, F., Jeffries, T., 2005. Geochemistry of carbonatite–silicate pairs in nature: a case history from central Italy. *Lithos* 85, 26–47.
- Stoppa, F., Jones, A.P., Sharygin, V.V., 2009. Nyerereite from carbonatite rocks at Vulture volcano: implications for mantle metasomatism and petrogenesis of alkali carbonate melts. *Central European Journal of Geosciences* 1 (2), 131–151.
- Storey, B.C., Alabaster, T., Pankhurst, R.J., 1992. Magmatism and the causes of continental break-up. Geological Society, London, Special Publications 68 (404 pp.).
- Tiercelin, J.-J., Thouin, C., Kala, T., Mondegue, A., 1989. Discovery of sublacustrine hydrothermal activity and associated massive sulphides and hydrocarbons in the North Tanganyika trough, East African Rift. *Geology* 17, 1053–1056.
- Toyoda, K., Horiuchi, H., Tokonami, M., 1994. Dupal anomaly of Brazilian carbonatites: geochemical correlations with hotspots in the South Atlantic and implications for mantle source. *Earth and Planetary Science Letters* 126, 315–331.
- Trumbull, R.B., Bühn, B., Romer, R.L., Volker, F., 2003. The petrology of basanite–tephrite intrusions in the Erongo Complex and implications for a plume origin of Cretaceous alkaline complexes in Namibia. *Journal of Petrology* 44 (1), 93–111.
- Trumbull, R.B., Harris, C., Frindt, S., Wigand, M., 2004. Oxygen and neodymium isotope evidence for source diversity in Cretaceous anorogenic granites from Namibia and implications for A-type granite genesis. *Lithos* 73, 21–40.
- Van der Hilst, R.D., Widiyantoro, S., Engdahl, E.R., 1997. Evidence for deep mantle circulation from global tomography. *Nature* 386, 578–584.
- Veksler, I.V., Lentz, D., 2006. Parental magmas of plutonic carbonatites, carbonate–silicate immiscibility and decarbonation reactions: evidence from melt and fluid inclusions. *Mineralogical Association of Canada Short Course* 36, pp. 123–149.
- Verwoerd, W.J., 1966. Fenitization of basic igneous rocks. In: Tuttle, O.F., Gittins, J. (Eds.), *Carbonatites*. Interscience Publication (John Wiley & Sons), London, pp. 295–308.
- Verwoerd, W.J., 1993. Update on carbonatites of South Africa and South West Africa. *South African Journal of Geology* 96, 75–95.
- Verwoerd, W.J., 2006. The Pilanesberg Alkaline Province. In: Johnson, M.R., Anhaeusser, C.R., Thomas, R.J. (Eds.), *The Geology of South Africa*. Geological Society of South Africa and Council for Geoscience, South Africa, pp. 381–393.
- Verwoerd, W.J., 2008. The Goudini carbonatite complex, South Africa: a re-appraisal. *The Canadian Mineralogist* 46, 825–830.
- von Knorring, O., 1986. A note on tin tantalum pegmatites in the Damara orogen and alkali rocks associated with the Brandberg complex. *Communications Geological Survey of South West Africa/Namibia* 2, 63–64.
- von Knorring, O., Hornung, G., Concliff, E., Emmertman, R., 1985. Brandberg intrusion in Namibia. Abstracts of the 13th Colloquium of African Geology 2. University of St. Andrews, Scotland, p. 281.
- Wang, J., Tatsumoto, M., Li, X., Premo, W.R., Chao, E.C.T., 1994. A precise ²³²Th–²⁰⁸Pb chronology of fine grained monazite: age of the Bayan Obo REE–Fe–Nb ore deposit, China. *Geochimica et Cosmochimica Acta* 58, 3155–3169.
- Wang, Y.J., Fan, W.M., Zhang, Y.H., Guo, F., Zhang, H.F., Peng, T.P., 2004. Geochemical, ⁴⁰Ar/³⁹Ar geochronological and Sr–Nd isotopic constraints on the origin of Paleoproterozoic mafic dikes from the southern Taihang Mountains and implications for the ca. 1800 Ma event of the North China Craton. *Precambrian Research* 135, 55–77.
- Watkins, R.T., McDougall, I., Le Roex, A.P., 1994. K–Ar ages of the Brandberg and Okenyanya igneous complexes, north-western Namibia. *Geologische Rundschau* 83, 348–356.
- Watkinson, D.H., Wyllie, P.J., 1971. Experimental study of the join NaAlSi₃O₈–CaCO₃–H₂O and the genesis of alkali rock–carbonatite complexes. *Journal of Petrology* 12, 357–378.
- White, R., McKenzie, D., 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. *Journal of Geophysical Research* 94, 7685–7729.
- Wigand, M., Schmitt, A.K., Trumbull, R.B., Villa, I.M., Emmertman, R., 2004. Short-lived magmatic activity in an anorogenic subvolcanic complex: ⁴⁰Ar/³⁹Ar and ion microprobe U–Pb zircon dating of the Erongo, Damaraland, Namibia. *Journal of Volcanic Geothermal Research* 130, 285–305.
- Wilson, M., 1989. *Igneous Petrogenesis – A Global Tectonic Approach*. Unwin Hyman, (466 pp.).
- Wingate, M.T.D., Pirajno, F., Morris, P.A., 2004. The Warakurna large igneous province: a new Mesoproterozoic large igneous province in west-central Australia. *Geology* 32, 105–108.
- Wolfe, E., Ebinger, C., Yirgu, G., Deino, A., Ayalew, D., 2004. Evolution of the northern main Ethiopian rift: birth of a triple junction. *Earth and Planetary Science Letters* 224, 213–228.
- Woolley, A.R., 1987. Lithosphere metasomatism and the petrogenesis of the Chilwa Province of alkaline igneous rocks and carbonatites, Malawi. *Journal of African Earth Sciences* 6, 891–898.
- Woolley, A.R., 2001. Alkaline rocks and carbonatites of the world – part 3: Africa. *The Geological Society of London* (372 pp.).
- Woolley, A.R., 2003. Igneous silicate rocks associated with carbonatites: their diversity, relative abundances and implications for carbonatite genesis. *Periodico di Mineralogia* 72, 9–17.
- Woolley, A.R., Kempe, D.R.C., 1989. Carbonatites: nomenclature, average chemical compositions, and element distribution. In: Bell, K. (Ed.), *Carbonatites, Genesis and Evolution*. Unwin Hyman, Boston, pp. 1–14.
- Woolley, A.R., Kjarsgaard, B.A., 2008. Carbonatite occurrences of the world: map and database. Geological Survey of Canada Open File 5796.

- Xia, L., Xia, Z., Xu, X., Li, X., Ma, Z., 2012. Mid-Late Neoproterozoic rift-related volcanic rocks in China: geological records of rifting and break-up of Rodinia. *Geoscience Frontiers* 3 (4), 375–399.
- Xie, Y.L., Hou, Z.Q., Xu, W.Y., Yuan, Z.X., Bai, G., Li, X.Y., 2006. Discovery of Cu–Zn, Cu–Sn intermetallic minerals and its significance for genesis of the Mianning–Dechang REE metallogenic belt, Sichuan Province, China. *Science China (Series D)* 49, 597–603.
- Xie, Y.L., Hou, Z.Q., Yin, S.P., Dominy, S.C., Xu, J.H., Tian, S.H., Xu, W.Y., 2009. Continuous carbonatitic melt-fluid evolution of a REE mineralization system: evidence from inclusions in the Maoniuping REE deposit, western Sichuan, China. *Ore Geology Reviews* 36, 90–105.
- Xie, Y.L., Li, Y.X., Cooke, D.R., Danyushevsky, L., Dominy, S.C., 2014. Ore genesis and deposit model of carbonatite-hosted REE deposits: the Mianing–Dechang REE belt, Western Sichuan Province, China. *Acta Geological Sinica (English edition)* 88 (Suppl. 2), 475–477.
- Xu, C., Huang, Z.L., Liu, C.Q., Qi, L., Li, W.B., Guan, T., 2003. Geochemistry of carbonatites in Maoniuping REE deposit, Sichuan Province, China. *Science China (Series D)* 46, 246–256.
- Yang, X.M., Le Bas, M.J., 2004. Chemical composition of carbonate minerals from Bayan Obo, Inner Mongolia, China: implications for petrogenesis. *Lithos* 72, 97–116.
- Yang, Z.X., William-Jones, A.E., Pu, G.P., 2000. Geological features of Maoniuping REE deposit, Sichuanm China. *Journal of Mineralogy and Petrology* 20, 28–34 (in Chinese with English abstract).
- Yang, Z.X., William-Jones, A.E., Pu, G.P., 2001. A fluid inclusion study of Maoniuping REE deposit, Sichuan, China. *Journal of Mineralogy and Petrology* 21, 26–33 (in Chinese with English abstract).
- Yang, K.F., Fan, H.R., Santosh, M., Hu, F.F., Wang, K.Y., 2011a. Mesoproterozoic mafic and carbonatitic dykes from the northern margin of the North China Craton: implications for the final breakup of Columbia supercontinent. *Tectonophysics* 498, 1–10.
- Yang, K.F., Fan, H.R., Santosh, M., Hu, F.F., Wang, K.Y., 2011b. Mesoproterozoic carbonatitic magmatism in the Bayan Obo deposit, Inner Mongolia, North China: constraints for the mechanism of superaccumulation of rare earth elements. *Ore Geology Reviews* 40, 122–131.
- Yang, X.Y., Lai, X.D., Pirajno, F., Liu, Y.L., Ling, M.X., Sun, W.D., 2014. Formation of the super-large Bayan Obo Fe–REE–Nb deposit, Inner Mongolia. *Acta Geological Sinica (English edition)* 88 (Suppl. 2), 478–479.
- Yirgu, G., Ebinger, C.J., Maguire, P.K.H. (Eds.), 2006. The Afar Volcanic Province Within the East African Rift System. Geological Society (London) Special Publications 259 (327 pp.).
- Yoshida, M., Santosh, M., 2011. Supercontinents, mantle dynamics and plate tectonics: a perspective based on conceptual vs. numerical models. *Earth-Science Reviews* 105, 1–24.
- Yuan, Z.X., Shi, Z.M., Bai, G., Wu, C.Y., Chi, R.A., Li, X.Y., 1995. The Maoniuping Rare Earth Ore Deposit. Seismological Press, Beijing (150 pp., in Chinese).
- Zhao, D.P., 2004. Global tomographic images of mantle plumes and subducting slabs: insights into deep Earth dynamics. *Physics of the Earth and Planetary Interiors* 146, 3–34.
- Zhao, D.P., 2007. Seismic images under 60 hot spots: search for mantle plumes. *Gondwana Research* 12, 335–355.
- Zhao, D.P., 2009. Multiscale seismic tomography and mantle dynamics. *Gondwana Research* 15, 297–323.
- Zhu, X. (Ed.), 2007. *Mineral Facts of China*. Elsevier, Amsterdam (776 pp.).
- Ziegler, P.A., Cloetingh, S., 2004. Dynamic processes controlling the evolution of rifted basins. *Earth-Science Reviews* 64, 1–50.