



# Nature and origin of eclogite xenoliths from kimberlites

D.E. Jacob\*

*Institut für Geowissenschaften, Universität Mainz, Becherweg 21, D-55099 Mainz, Germany*

Received 21 June 2003; accepted 25 January 2004

Available online 25 May 2004

## Abstract

Eclogites from the Earth's mantle found in kimberlites provide important information on craton formation and ancient geodynamic processes because such eclogites are mostly Archean in age. They have equilibrated over a range of temperatures and pressures throughout the subcratonic mantle and some are diamond-bearing. Most mantle eclogites are biminerale (omphacite and garnet) rarely with accessory rutiles. Contrary to their overall mineralogical simplicity, their broadly basaltic-picritic bulk compositions cover a large range and overlap with (but are not identical to) much younger lower grade eclogites from orogenic massifs. The majority of mantle eclogites have trace element geochemical features that require an origin from plagioclase-bearing protoliths and oxygen isotopic characteristics consistent with seawater alteration of oceanic crust. Therefore, most suites of eclogite xenoliths from kimberlites can be satisfactorily explained as samples of subducted oceanic crust. In contrast, eclogite xenoliths from Kuruman, South Africa and Koidu, Sierra Leone stem from protoliths that were picritic cumulates from intermediate pressures (1–2 Ga) and were subsequently transposed to higher pressures within the subcratonic mantle, consistent with craton growth via island arc collisions. None of the eclogite suites can be satisfactorily explained by an origin as high pressure cumulates from primary melts from garnet peridotite.

© 2004 Elsevier B.V. All rights reserved.

*Keywords:* Eclogites; Xenoliths; Major elements; Trace elements; Isotopes; Mantle

## 1. Introduction

Eclogitic xenoliths are found in most kimberlites and from all cratonic areas are also abundant at off-craton localities (e.g. Appleyard, 2003; Pearson et al., 1995c; Robey, 1981). Most xenolith populations are dominated by peridotite (Gurney et al., 1991; Sobolev et al., 1977), but eclogite can make up the majority of mantle nodules at some localities, such as Roberts Victor, Bellsbank, Newlands (South Africa), and Zaga-dochnaya (Siberia). This overabundance may reflect

local enrichment of eclogite within the predominantly peridotitic subcontinental mantle, but is in some cases interpreted to be an effect of differences in xenolith preservation during sampling by the kimberlite. Schulze (1989) showed that despite the overabundance of eclogite nodules at Roberts Victor (estimated to be 80–98%, Hatton, 1978; MacGregor and Carter, 1970), garnet peridotite dominates the kimberlite heavy mineral concentrate. This author estimated the overall abundance of eclogite in the upper 200 km of subcontinental mantle to be < 1 vol.%.

Eclogite is the original host rock from which diamond was first recovered (Bonney, 1899) and diamond with eclogitic affinity dominates the dia-

\* Tel.: +49-6131-3923170; fax: +49-6131-3923070.

E-mail address: [jacobd@uni-mainz.de](mailto:jacobd@uni-mainz.de) (D.E. Jacob).

mond population at some localities, for example at Orapa (Botswana; Gurney et al., 1991). Best studied, therefore, are eclogite suites from major diamond mines on the Kaapvaal craton, Siberian platform, Congo and West African cratons. More recently, investigations have begun on those from the Slave and the Karelian cratons.

Off-craton eclogites are distinct from those from on-craton localities in that they are often associated with mafic garnet granulites that are less common (although not absent) in on-craton kimberlites (Griffin et al., 1979). Furthermore, they are finer grained than on-craton eclogites and show lower equilibration pressures (1–2.3 GPa; Robey, 1981). Equilibration temperatures overlap with those for the on-craton eclogites (Pearson et al., 1995c).

Although eclogite is apparently only a minor component of the Earth's mantle, it plays an important role in geodynamic processes, i.e. subduction, for which the gabbro-eclogite transition was believed to be the major driving force (Ringwood and Green, 1966). Since the onset of plate tectonics, this process transferred large amounts of basaltic material into the Earth's mantle, today impressively imaged by seismic tomography (e.g., van der Hilst, 1995) giving rise to tonalitic magmas upon subduction (e.g. Foley et al., 2002) and adding to mantle heterogeneity (e.g. Albarède and van der Hilst, 2002). Subducted oceanic crust in the form of eclogite is discussed by some authors as a component in continental flood basalts (e.g. Cordery et al., 1997; Lassiter and DePaolo, 1997) or as representing a "hidden reservoir" for Nb, Ta, Ti counterbalancing the mass imbalance for these elements in the silicate Earth (Rudnick et al., 2000). Furthermore, as most dated eclogite suites are Archean in age, the study of mantle eclogites allows insight into Archean geodynamic processes (e.g. Foley et al., 2003).

## 2. Hypotheses for the origin of eclogite xenoliths from kimberlites

Models for the origin of eclogites cover a wide range between two extremes. Those put forward mainly in the 1960s focussed on bimineralic eclogitic xenoliths and interpreted them to be of purely mantle origin, originating as high-pressure cumulates from mantle melts (Caporuscio and Smyth, 1990; Hatton

and Gurney, 1987; O'Hara, 1969; O'Hara et al., 1975; O'Hara and Yoder, 1967). Ringwood and Green (1966) proposed the basalt–eclogite transition as the driving force for subduction and later models consequently interpreted the eclogitic xenoliths as remnants of subducted oceanic crust (e.g. Helmstaedt and Doig, 1975; Jagoutz et al., 1984; MacGregor and Manton, 1986). This latter hypothesis experienced growing attention, refinement and alteration over time leading to a number of variants, e.g. models in which eclogites are residual subducted oceanic crust after tonalitic melt removal at high pressure (e.g. Barth et al., 2001; Hatton and Gurney, 1987; Ireland et al., 1994; Jacob and Foley, 1999). More recently, some eclogite suites were interpreted as subducted cumulates formed at low-pressures in the oceanic crust, such as gabbros and spinel-gabbros, (Barth et al., 2002b; Jacob et al., 2003b) attesting to the petrological diversity of oceanic crust. Finally, Schmickler et al. (in press) proposed sanidine-bearing eclogites from the margin of the Kaapvaal craton (Kuruman province) to be derived from phlogopite–pyroxenite veins in the upper lithospheric mantle. From this plethora of models involving a variety of subduction-related processes it is clear that eclogite in the Earth's mantle is not of uniform origin and it seems unjustified that the so-called "mantle hypothesis" has not received more attention over time. This may, however, be caused by the fact that evidence in favour of a subduction origin is much more evident and unequivocal, being based mainly on a deviation of the  $\delta^{18}\text{O}$  values from those of the Earth's mantle. Such oxygen isotopic anomalies can only be caused at low pressures and temperatures (Clayton et al., 1975) and are commonly interpreted as a seawater alteration signature of oceanic crust (e.g. Barth et al., 2001; Jacob et al., 1994; Jagoutz, 1984). Similarly striking evidence is not available for the identification of high-pressure cumulates or melts. However, as will be shown here, constraints on the characteristics of such rocks can be derived from forward-modelling and trace element systematics (e.g. Jacob et al., 2003b).

## 3. Petrography and classification

Eclogites from the Earth's mantle are high-grade metamorphic rocks with large grain sizes and few primary mineral phases. Whereas eclogites from

orogenic massifs may contain abundant inclusions (e.g. allanite), garnet and clinopyroxene in eclogite xenoliths from kimberlites are generally of high purity and very rarely contain mineral inclusions that give evidence for their prograde metamorphic history. Exceptions are coesite reported from some eclogites (e.g. Hatton, 1978; Sobolev, 1977; Schulze et al., 2000) and sanidine–coesite inclusions in garnets from Kuruman eclogites, interpreted as phlogopite (or phengite) breakdown products (Schmickler et al., *in press*). The large majority of mantle eclogites are biminerally garnet- and clinopyroxene-bearing rocks, commonly with rutile as an accessory. Other primary minor or accessory phases comprise coesite (typically transformed to quartz), ilmenite, sanidine, orthopyroxene, diamond, graphite, kyanite, corundum, apatite, zircon and sulphides. Recently, olivine has been described from eclogites of the Slave craton (Fung, 1998; Kopylova et al., 1999) and spinel-bearing eclogites are known from the Kaalvaallei and Orapa kimberlites (Shee and Gurney, 1979; Viljoen, 1994). Coesite or quartz was only rarely reported from mantle eclogites and this apparent scarcity used to be interpreted as one main difference between mantle- and massif eclogites, where quartz is abundant (e.g. Rudnick, 1995). More recent detailed studies, however, show that its abundance in mantle eclogites may have been underestimated, at least at Roberts Victor, RSA (Schulze et al., 2000). Amphibole, plagioclase and phlogopite in mantle eclogites are of secondary origin.

Exsolution features are common and comprise, for example, garnet exsolving from clinopyroxene (e.g. Udachnaya, Siberian platform, Jerde et al., 1993a), orthopyroxene exsolution from clinopyroxene (e.g. Kuruman, Kaapvaal craton; Schmickler et al., *in press*), rutile from garnet (common in Roberts Victor group II eclogites), sanidine from clinopyroxene (e.g. Kuruman, Kaapvaal craton; Schmickler et al., *in press*), and apatite from garnet (e.g. Ekati, Slave craton and Koidu, West African craton; Haggerty et al., 1994; Jacob et al., 2003a).

Classifications of eclogites are based on southern African and Siberian samples and can be applied reasonably well to mantle eclogites worldwide. MacGregor and Carter (1970) introduced a combined textural and geochemical classification scheme for Roberts Victor eclogites dividing them into groups I

and II. Group I eclogites contain subhedral or rounded garnets in a clinopyroxene matrix, sometimes displaying layering, whereas group II rocks have interlocking texture of anhedral garnet and clinopyroxene and are more gneissic in appearance, often very fresh. Group I garnets are higher in Na<sub>2</sub>O and MgO than those of group II, group I clinopyroxenes have more K<sub>2</sub>O, FeO, Cr<sub>2</sub>O<sub>3</sub>, CaO and MnO than group II clinopyroxenes. McCandless and Gurney (1989) refined and extended the classification of MacGregor and Carter (1970) and differentiated the often diamondiferous group I eclogites by their higher Na<sub>2</sub>O<sub>gt</sub> ( $\geq 0.09$  wt.%) and K<sub>2</sub>O<sub>cpx</sub> ( $\geq 0.08$  wt.%) from the non-diamondiferous group II eclogites at Roberts Victor. The elevated Na<sub>2</sub>O contents of garnets and K<sub>2</sub>O contents in clinopyroxenes most likely reflect higher pressures of equilibration for group I eclogites, as results from experimental petrology and diamond inclusion studies attest to the pressure dependent incorporation of these elements (Erlank and Kushiro, 1970; Harlow, 1997; Sobolev and Lavrentev, 1971). An alternative classification scheme was developed by Taylor and Neal (1989) on eclogite xenoliths from Bellsbank, Kaapvaal craton and is inspired by the eclogitic garnet classification by Coleman et al. (1965). This classification differentiates three groups based on MgO and Na<sub>2</sub>O contents (diopside and jadeite molecules) in clinopyroxenes: Group A rocks contain diopsidic clinopyroxenes with high MgO and low Na<sub>2</sub>O contents, group B eclogites are intermediate and group C rocks contain pronounced jadeite-rich clinopyroxenes. Rocks in group A are ultramafic, predominantly pyroxenites or show transitions towards this rock group; kyanite and corundum eclogites are group C according to this classification. Groups I and II eclogites of MacGregor and Carter (1970) and McCandless and Gurney (1989) are found in group B and C of Taylor and Neal (1989) and are not distinguished by this classification.

Pyroxenites are more common in off craton or craton-rim localities. Very few are described from the well-sampled kimberlite pipes on the cratons (e.g. Egger et al., 1987) and only at Orapa/Lethlakane is a considerable percentage of pyroxenite documented. Because some eclogite suites show compositional transitions towards pyroxenites (e.g. at Orapa or the Kuruman suite, Kaapvaal craton) it is sometimes

difficult to draw the line between eclogite and garnet pyroxenite. A useful distinction is the ratio of Tschermak's to jadeite molecules in clinopyroxene expressed as Al [6]/Al [4] ratio (Aoki and Shiba, 1973). Eclogites have Al [6]/Al [4] ratios  $\geq 2$ , reflecting higher equilibration pressures, but differentiation is difficult at low total Al-contents or when high-pressure pyroxenites are encountered (e.g. Green, 1966). The classification scheme of McCandless and Gurney (1989) helps in the identification of diamond-friendly conditions and is probably the most widely applied.

#### 4. Major element chemistry and equilibration conditions

##### 4.1. Mineral compositions

Garnets in mantle eclogites are Cr-poor pyrope–grossular–almandine mixtures with wide variations in grossular-contents. Eclogites in which the grossular component dominates are kyanite- and/or corundum-bearing, high Al bulk compositions and are termed grospsydites (*grossular–pyroxene–disthene* (=kyanite)) rocks (Sobolev et al., 1977). Most eclogite suites have a small grospsydite component. Data published on eclogite xenoliths from the Siberian Zagadochnaya pipe (Fig. 1G) represent an effort to sample explicitly the grospsydite component at this locality (Sobolev, 1977: the plotted data are therefore not representative for the Zagadochnaya eclogite population). Available data on the Jagersfontein eclogite suite has a very restricted chemical composition (Fig. 1C). Similarly, no grospsydites have been reported from eclogite suites of kimberlites from the Slave craton and Colorado/Wyoming (Fig. 1H). All eclogites here classify as group II, based on their garnet composition. The Orapa and Lethlakane mines (Botswana, Fig. 1F) yield a large amount of pyrope-rich, mostly pyroxenitic garnets that are group II. For most pipes, group I and group II eclogites overlap compositionally. The only pipe without compositional overlap between the two eclogite groups is Kaalvallei (South Africa, Fig. 1D), implying that two compositionally distinct suites of eclogite (of different age?) are stored at different depths in the litho-

sphere at this locality. At Koidu, Sierra Leone (Fig. 1E) two eclogite suites are identified by their bulk MgO-contents (e.g. Hills and Haggerty, 1989) and distinct origins are suggested for the low MgO and high MgO suites (Barth et al., 2001, 2002b). Some garnets of the “high MgO” suite, however, plot to high Fe-contents in Fig. 1E (triangles), illustrating the effects of kimberlite infiltration on the eclogite bulk compositions (see below).

Clinopyroxenes in eclogites are omphacites with jadeite contents typically between 20 and 70 mol%, corresponding to Na<sub>2</sub>O-contents between approximately 2 and 8 wt.%. They are distinguished from pyroxenitic clinopyroxenes by their higher jadeite component, as outlined above.

Orthopyroxene-bearing eclogites are not very common, but are more abundant in off-craton and craton-rim kimberlites than at on-craton localities (e.g. Rietfontein, Kuruman: Appleyard, 2003; Schmickler et al., *in press*). Some opx-eclogites are known from on-craton kimberlites such as Roberts Victor, Kaalvallei and Jagersfontein. Orthopyroxenes from Kuruman and Rietfontein have Mg-numbers (Mg-number =  $100 \times \text{Mg}/(\text{Mg} + \text{Fe})$ ) between 74 and 89 and contain between 0.58 and 0.95 wt.% Al<sub>2</sub>O<sub>3</sub>.

Unfortunately, no suitable barometers exist for biminerally eclogites. Equilibrium pressures can therefore not be determined for the majority of the on-craton suites, although the occurrence of coesite and diamond clearly show that they represent high-pressure assemblages. For orthopyroxene-bearing eclogites pressures can be determined using several geobarometers developed for peridotites. The most widely used and most accurate are those based on the Al exchange between garnet and orthopyroxene (e.g. Brey and Köhler, 1990).

Equilibration temperatures are usually calculated using the Mg–Fe<sup>2+</sup> exchange between garnet and clinopyroxene at an assumed pressure (often 5 GPa). The most widely used geothermometer is that by Ellis and Green (1979), which has the advantage of being calibrated for the range of Mg-numbers found in eclogites. The Fe<sup>3+</sup> correction required is commonly bypassed by assuming all Fe as Fe<sup>2+</sup>, which results in minimum temperatures. The newer refined calibration by Ai (1994) for a *P–T* range of 1 to 6 GPa and 600 to 1500 °C is believed to be the most accurate for Mg-rich eclogites.

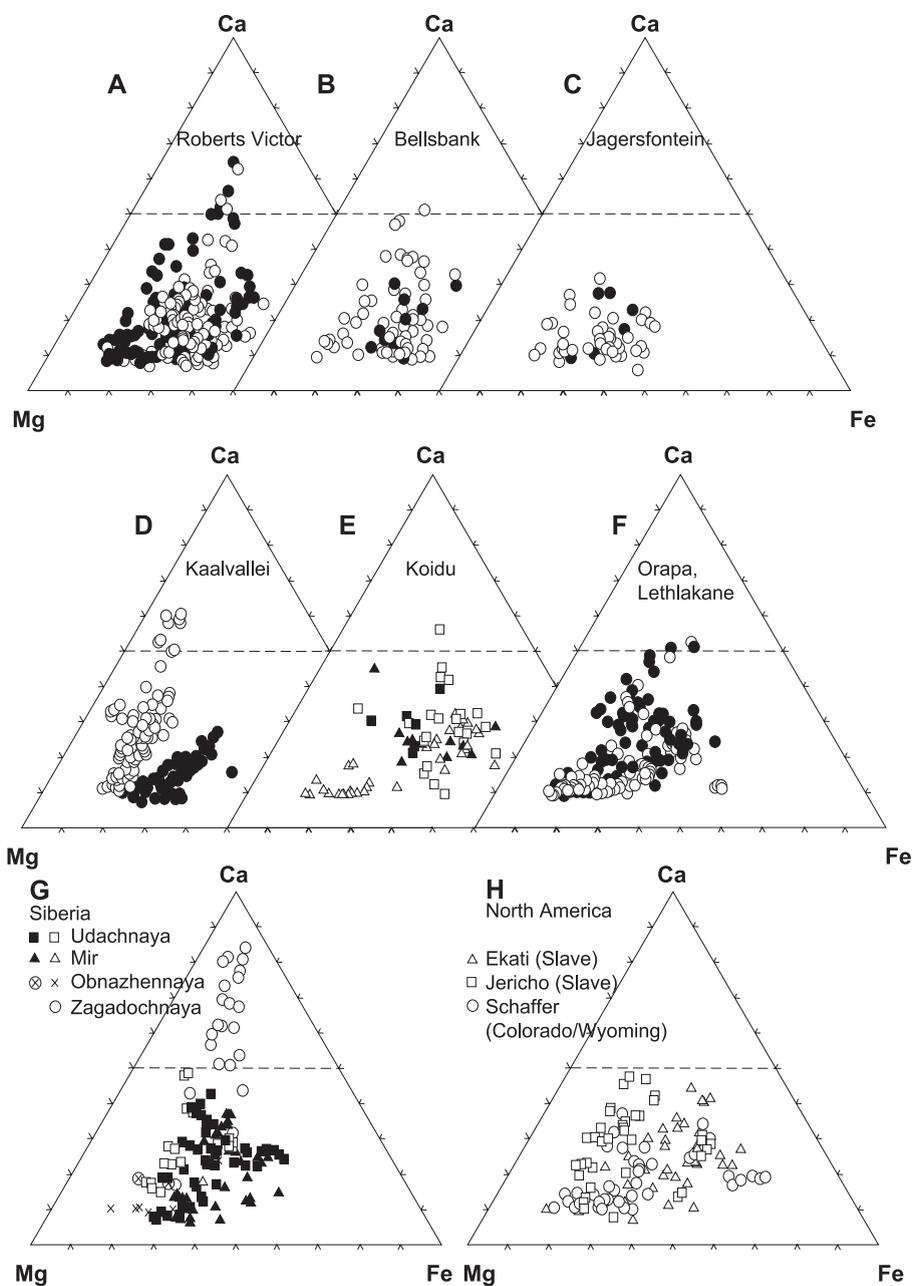


Fig. 1. A–H: Trilateral garnet compositions for bimineralic eclogites from individual kimberlite pipes of different cratons. Open symbols are garnets from group II eclogites; solid symbols are from group I eclogites after McCandless and Gurney (1989). The high-MgO suite of the Koidu pipe (label E) are plotted as triangles, low MgO eclogites are squares. (Data sources: database of the Kimberlite Research Group, University of Cape Town (queries are carried out on request) and Hatton, 1978; Hills and Haggerty, 1989; Jerde et al., 1993a,b; Kopylova et al., 1999; MacGregor and Manton, 1986; McCulloch, 1989; Pearson et al., 1999; Schulze et al., 2000; Shee and Gurney, 1979; Smith et al., 1989; Snyder et al., 1997; Sobolev et al., 1994; Stiefenhofer et al., 1997; Taylor and Neal, 1989; Viljoen, 1994; Viljoen et al., 1996.)

To derive a pressure estimate for biminerally eclogites, equilibration temperatures are often projected to a paleogeotherm, either deduced from peridotite xenoliths from the same pipe or to a shield geotherm (Pollack and Chapman, 1977). For the Kuruman eclogite suite, however, it has been shown that peridotite xenoliths and orthopyroxene-bearing eclogites describe distinct paleogeotherms. Extrapolation of the temperatures obtained for biminerally eclogites to the peridotite paleogeotherm would result in erroneously high pressures for the biminerally eclogites (Fig. 2; Schmickler et al., in press).

Equilibration temperatures (at an assumed pressure of 5 GPa) calculated with major element data plotted in Fig. 1 generally vary between approximately 800 and 1380 °C (Fig. 3). Eclogites from the Siberian pipes yield lower average temperatures than from pipes on other cratons, consistent with a slightly lower geothermal gradient proposed for this region (e.g. Griffin et al., 1999) and those from the Obnazhennaya pipes record the lowest (700 to 960 °C at 5 GPa). Group I eclogites yield overall higher

average temperatures than group II eclogites, supporting their higher pressure origin. Canadian eclogites show a very similar range of temperatures (800–1350 °C at 5 GPa).

MacGregor and Manton (1986) suggested that most eclogites are derived from the base of the lithosphere, from depths in excess of 150 km, and Gurney (1990) proposed that a pre-existing peridotitic cratonic keel may have been periodically underplated with diamondiferous eclogites. The presence of diamond in group I eclogites clearly indicates a high pressure origin of >5GPa (Kennedy and Kennedy, 1976) and latest results from seismic tomography showing a cratonic keel of at least 200–300 km below the Kaapvaal craton (Fouch et al., in press) provide the prerequisites for a deep lithospheric origin. In contrast, ranges of equilibration pressures and temperatures in group II eclogites, where diamond as a high pressure indicator mineral is lacking, are much larger and suggest that eclogites reside throughout the depth profile of the continental lithosphere.

#### 4.2. Bulk compositions

Generally of picritic to basaltic character, the original bulk major element composition can be affected by mantle metasomatism and infiltration by kimberlitic melts as well as by reactions between primary phases and the kimberlite. As a cautionary note, it should be pointed out that *all* eclogite xenoliths are infiltrated by kimberlitic material, and bulk analyses of trace elements and radiogenic isotopes never represent the eclogite composition, but are rather mixtures of eclogite and kimberlite (or its lower pressure alteration products). Pre-entrainment metasomatic changes in composition, less common than kimberlite infiltration, have been recognized in individual samples from most pipes, as well as in whole eclogite suites, as for example that from Kimberley, South Africa. Metasomatic effects may be both cryptic and patent (Dawson, 1984), and although trace elements are affected much more than major elements (see below) metasomatic overprints on the major element concentration have been recognized. Ireland et al. (1994) for example, compared Siberian eclogitic diamond inclusions occurring within

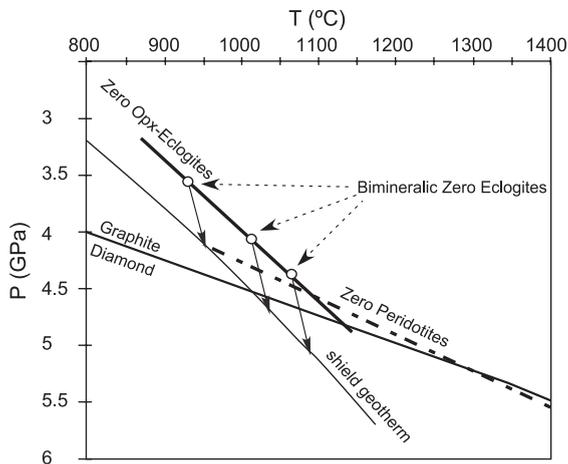


Fig. 2.  $P$ – $T$  plot for eclogites from the Kuruman kimberlites (Zero pipe, Schmickler et al., in press). Biminerally eclogites (open circles) are plotted onto the geotherm derived for opx-bearing eclogites from the same locality. Vectors denote loci of the biminerally samples if their temperatures are extrapolated onto a typical shield geotherm (Pollack and Chapman, 1977). This example shows that extrapolation of temperatures onto a shield geotherm may result in erroneous pressures for biminerally eclogites. Graphite–diamond boundary and geotherm for the Kuruman peridotites (Shee et al., 1989) are shown for comparison.

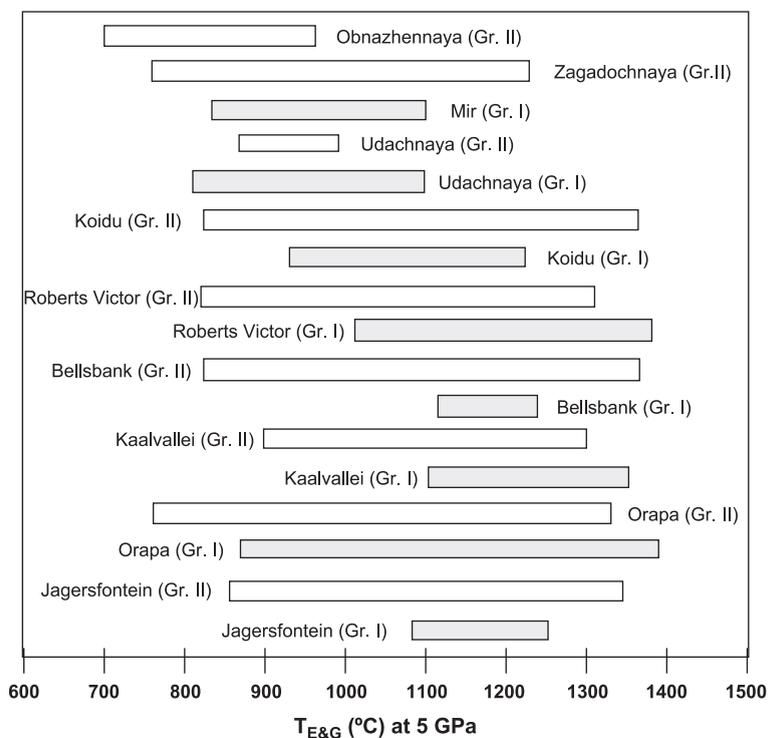


Fig. 3. Temperature ranges (Ellis and Green, 1979) for group I and group II eclogites (McCandless and Gurney, 1989) for individual kimberlite pipes calculated at 5 GPa. In general, group II eclogites cover a wider range of temperatures supporting the view that they are derived from a larger section throughout the lithosphere. Group I eclogites, derived from higher pressures than group II eclogites show generally more restricted and higher temperature ranges. Siberian eclogite suites plot to slightly lower temperatures, consistent with the lower Siberian geotherm (Griffin et al., 1999).

xenoliths with their eclogite host minerals. These authors found lower Mg-numbers in the diamond inclusions than in the eclogite host minerals and concluded that chemical exchange with kimberlitic melts with higher Mg-numbers and rich in incompatible elements changed the composition of those minerals not encapsulated in diamond. A similar observation was made by Barth et al. (2001) for some of the eclogite xenoliths from Koidu, West Africa. However, the major element bulk compositions of most eclogite xenoliths do not show signs of metasomatism, which prevents their reconstruction using mineral and modal analyses that would introduce further uncertainties.

Eclogitic xenoliths are distinct from those exposed in high-pressure orogenic terrains (“massif eclogites”) in that the former are on average picritic rather than basaltic (average MgO = 13.8 wt.% as opposed to 9.7 wt.% in the latter) and have lower average bulk

SiO<sub>2</sub>, TiO<sub>2</sub>, Na<sub>2</sub>O (Fig. 4) and Cr<sub>2</sub>O<sub>3</sub> contents at comparable CaO contents. The lower SiO<sub>2</sub> content also manifests itself by the general scarcity of quartz in eclogite xenoliths, although quartz is a common mineral in massif eclogites. Both rock groups are explained as subducted oceanic crust (e.g. Stosch and Lugmair, 1990), which in the case of massif eclogites is supported by the fact that many of their bulk compositions are similar to that of MORB. Rudnick (1995) for example, explained the discrepancies between eclogitic xenoliths and MORB/massif eclogites by loss of a tonalitic melt component at high pressures in the case of the xenoliths.

Although some eclogite suites have examples with high MgO values up to 20 wt.% (e.g. Koidu, West Africa: Barth et al., 2002b), none extend to the very high MgO values of komatiites (Fig. 4). Bulk TiO<sub>2</sub> contents range between 0.06 to about 2 wt.%, similar to the range found in komatiites, island arc

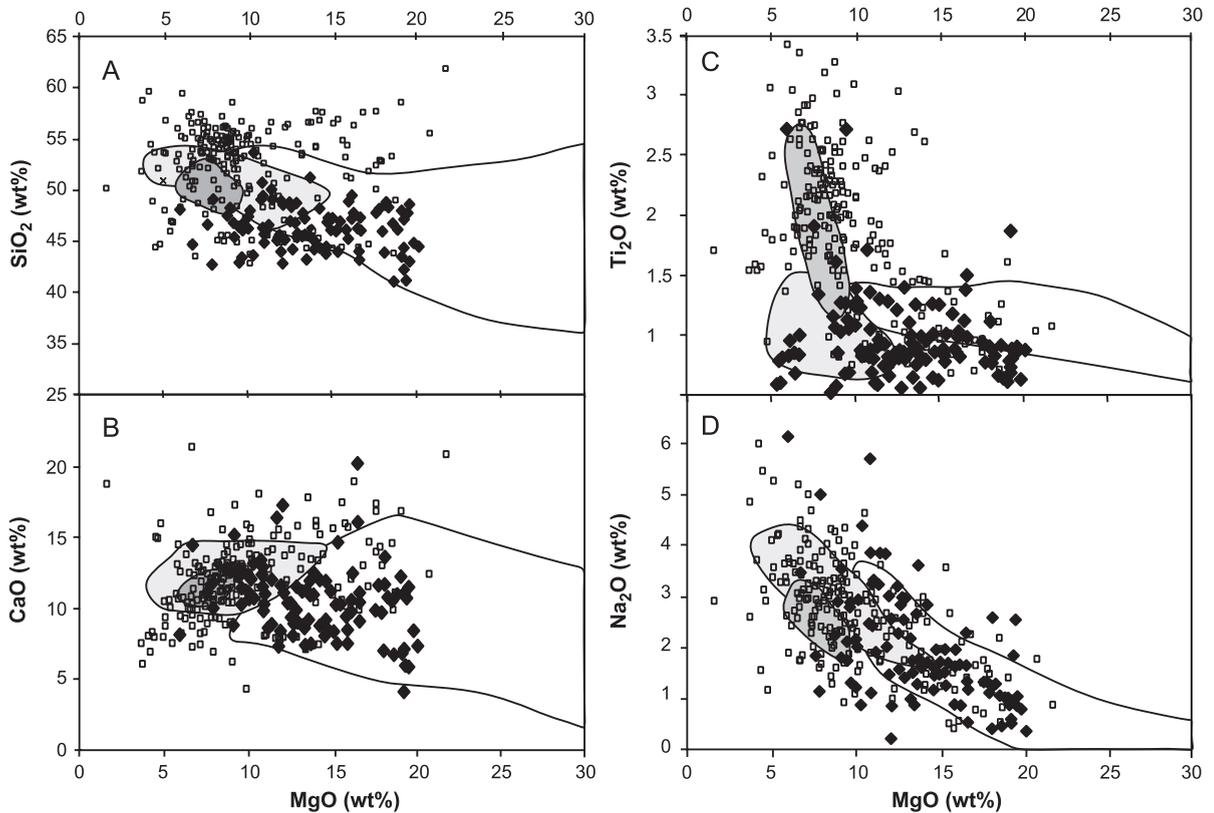


Fig. 4. A–D: Bulk major element plots for mantle eclogites (black diamonds, compiled from the literature) compared with eclogites from orogenic massifs (open squares) and fields for komatiites (white), oceanic gabbros (light grey, [Bach et al., 2001](#)) and MORB (dark grey). On average, mantle eclogites are distinct from massif eclogites and have lower  $\text{SiO}_2$ ,  $\text{TiO}_2$  and higher  $\text{MgO}$ . Massif eclogites are more similar to MORB, especially in  $\text{TiO}_2$ .

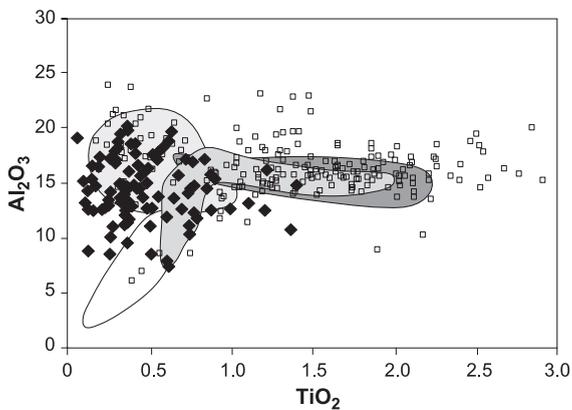


Fig. 5. Bulk  $\text{Al}_2\text{O}_3$  vs.  $\text{TiO}_2$  in weight percent for mantle eclogites and massif eclogites compared with fields for komatiites (white), Archean basalts (grey), oceanic gabbros (light grey) and MORB (dark grey). Symbols and data sources as in [Fig. 4](#).

related volcanics and oceanic gabbros, but are lower than the majority of mid-ocean ridge basaltic glasses ([Fig. 5](#)).

## 5. Trace element compositions

In contrast to major elements, the trace element bulk composition of *every* eclogite xenolith is significantly influenced by kimberlite infiltration. [Barth et al. \(2001\)](#) demonstrated by mass balance that the measured bulk trace element composition of Koidu eclogites was affected by 1% to 5% infiltration of kimberlitic melts ([Fig. 6](#)), significantly altering the incompatible trace element budget. Accordingly, trace element bulk compositions must be reconstructed using measured mineral compositions and estimated

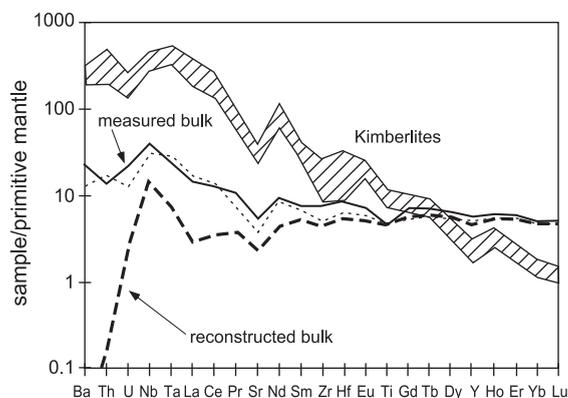


Fig. 6. Example for the effect of kimberlite infiltration upon the bulk trace element budget of mantle eclogites after Barth et al. (2001). The reconstructed trace element bulk composition of the eclogite sample (based on modal analyses and mass balance) is significantly depleted in LREE and LILE compared to the measured eclogite composition. However, addition of 5% of kimberlite to the reconstructed composition can mimic the measured whole rock eclogite.

modal mineralogy, as described below. During in situ mineral trace element analyses by Laser-Ablation ICP-MS, secondary alteration products may be encountered along even the finest cracks in otherwise pristine primary minerals. This, if unrecognized, can result in artificially high concentrations of elements that are enriched in kimberlite compared to eclogite, as for example Ba, Cs, Nb, La, Ce, U, Th. Most pristine, unaltered eclogitic garnets, for example, contain less than approximately 0.2 ppm Ba, so that this element can be used as an alteration monitor during analysis.

### 5.1. Rare earth elements

There are two typical types of patterns for rare earth elements in garnets from eclogitic xenoliths: Most rare earth element patterns of eclogitic garnets are light rare earth element (LREE) depleted with  $Ce_N$  as low as 0.01 (Jacob et al., 2002; Fig. 7). Heavy rare earth elements (HREE) are variably enriched; extreme enrichments of  $Lu_N = 70$  are found in Roberts Victor eclogites (Harte and Kirkley, 1997). The other type of REE pattern has nearly flat HREE with  $Lu_N < 10$  and positive Eu anomalies (Figs. 7 and 8) and is displayed by roughly half of all published garnet data (49% of the worldwide database). This second type is espe-

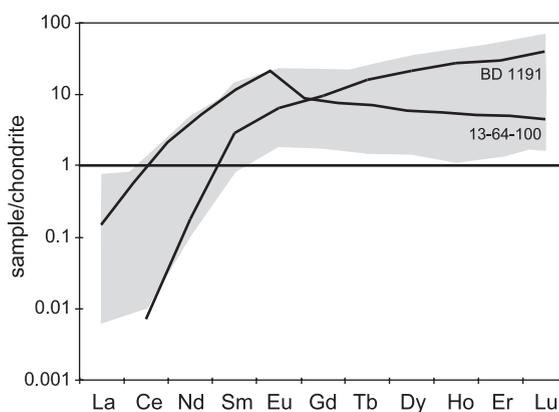


Fig. 7. Rare Earth element patterns of garnets from mantle eclogites worldwide (shaded field compiled from the literature and own unpublished data). The two black lines are examples for extreme LREE depletion found in some eclogitic garnets and patterns with positive Eu anomalies and flat HREE (sample 13-64-100 taken from Jacob et al., 2003b).

cially common amongst coesite-bearing eclogite xenoliths (e.g. Jacob et al., 2003b) as well as amongst those containing kyanite and/or corundum (e.g. Snyder et al., 1997; Jerde et al., 1993a,b; Harte and Kirkley, 1997). These type II patterns are very unlike equilibrium REE patterns for garnet from high-pressure rocks that are expected to have high abundances of HREE due to its mineral-melt partitioning charac-

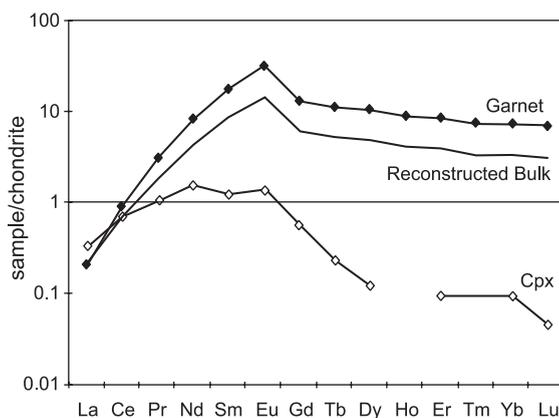


Fig. 8. A typical example for a mantle eclogite displaying positive Eu-anomalies in clinopyroxene, garnet and consequently in the reconstructed whole-rock. Note also the flat HREE pattern of garnet, not consistent with the mineral-melt equilibrium partitioning behaviour of this mineral (data for sample 13-64-100 from Jacob et al., 2003b).

teristics. These anomalies, when traced by the reconstructed bulk composition and associated with bulk positive Sr anomalies (see below), are interpreted as evidence of a prograde metamorphic reaction from plagioclase to garnet and are used in support of the subduction theory that explains the mantle eclogites as subducted oceanic crustal rocks (e.g. Jacob et al., 2003b; Jagoutz et al., 1984).

The majority of *clinopyroxenes* in mantle eclogites display convex-upward REE patterns with moderately enriched LREE concentrations of around  $Ce_N=30$  relative to HREE ( $Yb_N=0.02$ , Fig. 9) and may show positive Eu-anomalies when coexisting with garnets showing such anomalies (Fig. 8). Very LREE depleted ( $Ce_N=0.005$ ) and enriched clinopyroxenes with  $Ce_N=70$  to 100 are only reported from Roberts Victor and from Bellsbank (Jacob et al., 2002, 2003b; Taylor and Neal, 1989). The enrichment is most likely a result of reaction of primary cpx with metasomatic agents prior to entrainment in the kimberlite (Jacob et al., 2003b).

### 5.2. High field strength elements

Abundances of High Field Strength Elements (HFSE: Nb, Ta, Zr, Hf) in garnets and clinopyroxenes vary and are buffered by rutile, where present. Silicates of rutile-free assemblages show higher Ti and HFSE concentrations than rutile-bearing eclogites.

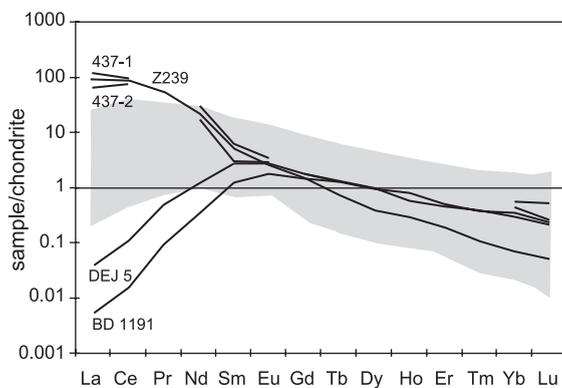


Fig. 9. Rare Earth element patterns of clinopyroxenes from mantle eclogites worldwide (shaded field). The black lines are examples for extreme LREE depletion and enrichment found in some eclogitic clinopyroxenes (Neal et al., 1990; Schmickler et al., in press and worldwide eclogite database).

Rutile, in turn, has very high partition coefficients for Ti and the HFSE and controls the bulk eclogite budget for these elements. Nb/Ta ratios in eclogitic rutiles are highly variable (Rudnick et al., 2000): very high Nb/Ta ratios ( $\leq 343$ ) and heterogeneous Nb and Ta concentrations (16,000 ppm Nb, 900–1700 ppm Ta) in addition to sometimes skeletal textures of rutiles in Koidu eclogites were used by Barth et al. (2001) to argue for a metasomatic origin of these rutiles. In contrast, very low and even undetectable Nb and Ta concentrations, low Nb/Ta and Zr/Hf ratios (Nb/Ta=3, Zr/Hf=16, Jacob et al., 2003b) in some Roberts Victor eclogites show that a partial melting event affected a rutile-free bulk composition, either under amphibolite facies conditions, below the rutile stability field, or at higher pressures, deep in the upper mantle, where all  $TiO_2$  is dissolved in cpx and garnet (Green and Sobolev, 1975; Klemme et al., 2002; Konzett, 1997). The low HFSE rutiles now present in these eclogites exsolved later from garnet upon cooling and decompression (Jacob et al., 2002).

Other minor and trace elements commonly analyzed are Cr, Mn, Co, Ni, Zn, Cu, Sc, V, Ga and recently also Li. Nickel in eclogitic garnets ranges between 4 and 300 ppm. The average for non-diamondiferous eclogite parageneses is 43 ppm, whereas garnets from diamondiferous eclogites have an average Ni content of 109 ppm. Principal carrier of Li in eclogites is clinopyroxene, which has about an order of magnitude higher Li contents than coexisting garnet (0.72–11.3 ppm as opposed to 0.052–0.82 ppm) (Woodland et al., 2002). Based on the small amount of data currently available, eclogite xenoliths have lower reconstructed bulk Li contents than eclogites from orogenic massifs.

### 5.3. Bulk rock trace element characteristics

Bulk rock reconstructions for the HFSE obviously are very sensitive to the modal abundances of rutile. Large xenoliths provide the opportunity to estimate rutile modal abundances via Ti mass balance (e.g. Barth et al., 2001, 2002b), but more often samples are too small for this approach. Modal abundances of garnet and clinopyroxene, in turn, may be estimated with an uncertainty of about 10–20%, depending on the size and homogeneity of the sample. Usually, garnet and clinopyroxene modal abundances range

between 70:30 and 40:60. Compositions dominated by garnet are more common than those where clinopyroxene makes up the majority. Jerde et al. (1993b) calculated bulk REE patterns for a suite of modal compositions with garnet/clinopyroxene ratios between 60:40% and 30:70%. They observed that many patterns are relatively insensitive to variance in mode and qualitative information is preserved even when the mode is changed by 30%.

Perhaps the most notable recalculated bulk trace element patterns are those that display positive Eu (Fig. 8) and Sr (not shown) anomalies and flat HREE. Eclogites with this type of pattern are known from many localities worldwide. Otherwise, trace element patterns are very variable with HREE enrichments in the order of  $\text{Lu}_N=30$  (e.g. Snyder et al., 1997) and variably depleted or enriched LREE. A sample with extreme HREE enrichment of  $\text{Lu}_N=100$  is known from the Koidu kimberlite pipe (Barth et al., 2002b), whereas eclogites from the Kuruman region have flat, chondritic REE patterns (Schmickler et al., in press). Metasomatism, related to the kimberlite but caused by passing melts before entrainment in the kimberlite affects the highly incompatible elements, such as Ba, Sr, Nb, Ta, Zr and the LREE (for a good assessment, see Barth et al., 2001, 2002b), resulting in “wavy” LREE patterns and overall enrichment of the affected trace elements. Ti and the HREE, however, are less influenced by metasomatism and have been found to give reliable information on the original eclogite composition. Compatible elements are also useful to constrain the eclogite precursor rocks. Reconstructed bulk Ni contents are between 60 and 900 ppm (100 to 300 ppm in diamondiferous eclogites), Co ranges between 30 and 100 ppm and Zn from 16 to 100 ppm.

A large number of mantle eclogites show Zr depletions relative to Sm together with depleted LREE in their recalculated bulk compositions. Barth et al. (2001) showed that in the case of Koidu eclogites this is not due to overlooked accessory phases such as zircons, because recalculated and measured bulk compositions both show Zr depletions. These depletions go with the observed general depletion of eclogite xenoliths in  $\text{SiO}_2$  (see above) and are a strong indication that many have lost a partial melt. Jacob and Foley (1999) reached a similar conclusion for a suite of diamondiferous eclogites from Udachnaya, which they interpreted as subducted oceanic

crust. While Mg-numbers of the samples are relatively high (up to 82) and could be an indication of olivine-accumulation in a picritic protolith, moderate reconstructed bulk Ni contents (102–237 ppm) do not support olivine-accumulation, that would have led to very high Ni contents. Loss of a silica-rich melt could explain this, as it would cause an increase in Mg-numbers but would give rise to only a slight increase in Ni content of the residue. It appears therefore that many eclogite xenoliths are residues after partial melting. Coesite/quartz-bearing eclogites may be an exception, because large amounts of partial melting would eliminate any free  $\text{SiO}_2$ . Yaxley and Green (1998) demonstrated experimentally that quartz is stable in an eclogite assemblage only up to 13% partial melting.

Melts produced by partial fusion of eclogites are of tonalitic composition (Rapp and Watson, 1995) and have been discussed as the major contributor to continental crustal growth. The HFSE systematics of continental crust, however, require amphibolite melting rather than eclogite melting (Foley et al., 2002).

## 6. Radiogenic and stable isotopes

Eclogite xenoliths are isotopically one of the most diverse rock groups on Earth. Their  $\epsilon_{\text{Nd}}$ -values, for example, cover 646 epsilon units between  $-38\epsilon$  (cpx from Mbuji Mayi, El Fadili and Dmaiffe, 1999) and  $+684\epsilon$  (gt from Roberts Victor, Jagoutz et al., 1984). This diversity is testimony to their antiquity as well as a reflection of a number of processes that affected the eclogites over time.

Bulk isotopic compositions for eclogite xenoliths are measured using the Re–Os isotopic system, whereas whole-rock compositions must be reconstructed from mineral analyses for the other radiogenic isotopic systems, introducing an uncertainty connected with the estimation of the mode, as discussed above. Neodymium, Sm, Lu and Hf have relatively similar partitioning behaviour for garnet and clinopyroxene, whereas Sr contents and most often Pb contents differ by up to two orders of magnitude between these two minerals. Strontium contents in garnet rarely exceed 1 ppm and are often less than 0.5 ppm (20 to 800 ppm in cpx). Lead concentrations measured by isotope dilution range

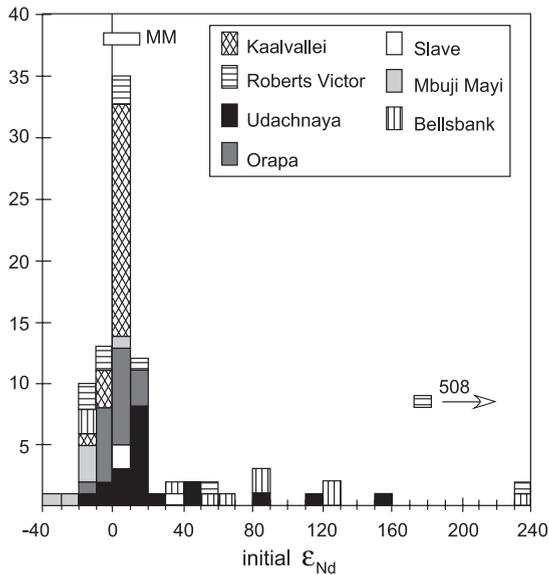


Fig. 10. Histogram of calculated bulk eclogite  $\epsilon_{Nd}$  at the time of kimberlite eruption. Mantle eclogites cover an immense range of Nd isotopic compositions between  $-26$  and  $+508$   $\epsilon_{Nd}$ . A vertical line denotes the composition of bulk Earth and the box labelled “MM” shows the isotopic composition of the mantle array onto which all major mantle melts fall. (Data sources: El Fadili and Demaiffe, 1999; Jacob and Jagoutz, 1995; Jacob et al., 1994; Jagoutz, 1988; Jagoutz et al., 1984; McCulloch, 1989; Neal et al., 1990; Pearson et al., 1995a; Roden et al., 1999; Smith et al., 1989; Snyder et al., 1997; Viljoen, 1994; Viljoen et al., 1996 and own unpubl. data.)

between 10 and 200 ppb and 170 to 523 ppb in coexisting clinopyroxenes (Jacob and Jagoutz, 1995; Jacob and Foley, 1999). Analysis of Pb isotopic data of garnets from kimberlitic xenoliths (peridotite and eclogite) shows that these very often form a linear array in  $^{206}\text{Pb}/^{204}\text{Pb}$ – $^{207}\text{Pb}/^{204}\text{Pb}$  space whose apparent age is identical to the kimberlite emplacement age, indicating U/Pb contamination by the kimberlite. It is therefore advisable to use only the Sr and Pb isotopic compositions of clinopyroxenes for comparisons, particularly as the garnet contributes little to the bulk rock composition. All data used here for interpretation and those shown in Fig. 10 are therefore recalculated bulk compositions using published modes or 60% gt: 40% cpx for Nd and Hf, whereas Pb and Sr data are from clinopyroxenes. The range of Nd and Sr isotopic compositions cover a much greater compositional space than all major magmas produced by the Earth’s mantle today, occupying all four quadrants of the Nd–Sr diagram. Some extremely radiogenic  $^{143}\text{Nd}/^{144}\text{Nd}$

ratios (Fig. 10) are consistent with strongly LREE depleted patterns and old Nd model ages found in the same rocks. These samples also possess some of the most unradiogenic initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios (Fig. 11), at time of kimberlite emplacement (0.70091, Jacob et al., 2003b; Jagoutz et al., 1984) and these characteristics are interpreted as effects of an ancient melting event. Very unradiogenic  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios found in eclogites range down to initial  $\epsilon_{Nd}$  values of  $-26$  (El Fadili and Demaiffe, 1999) and indicate long-term light REE enrichment, but are less extreme than those found in peridotitic rocks (Pearson et al., 1995a).

Few hafnium isotopic data exist for eclogite xenoliths. Initial measurements on Roberts Victor eclogites (Jacob et al., 2002) indicate extreme Hf isotopic variability reflecting the image shown by Sm–Nd systematics. Initial  $\epsilon_{Hf}$  range between  $-9.2$  to  $+166$  at initial  $\epsilon_{Nd}$ -values of  $-22$  to  $484$ . Most of the analysed samples plot below the mantle array, which defines the loci of all major magmas from the Earth’s mantle.

Only few Pb isotopic studies exist and these concentrate on Roberts Victor and Udachnaya eclogites (Jacob and Jagoutz, 1995; Jacob and Foley, 1999; Kramers, 1977, 1979). Clinopyroxene data plot

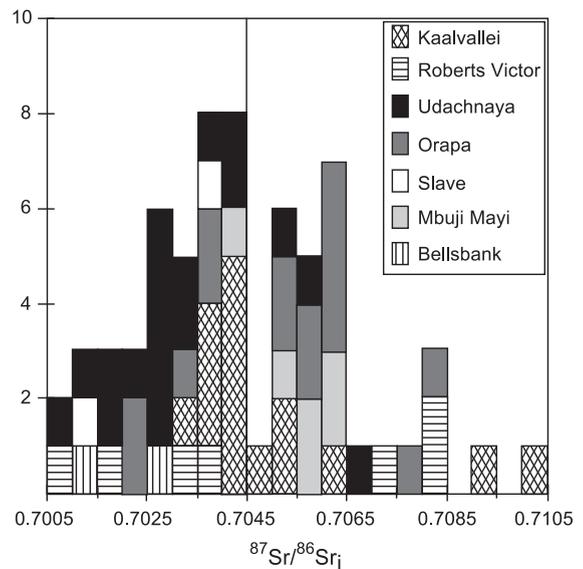


Fig. 11. Histogram of  $^{87}\text{Sr}/^{86}\text{Sr}$ -ratios of clinopyroxenes from mantle eclogites recalculated to the kimberlite eruption ages. Vertical bar denotes present bulk Earth composition of 0.7045. Data sources as in Fig. 10.

on both sides of the Geochron, and samples from both localities span a similar range of Pb isotopic ratios (Fig. 12).

Os isotopic compositions are commonly very radiogenic with initial  $\gamma_{Os}$  values ranging from close to chondritic to  $>6500$ . They are similar to those of Archean basalts and komatiites (e.g. Barth et al., 2002a; Menzies et al., 1998; Pearson et al., 1995a, Fig. 13) and are indicative of long-term evolution with basaltic–picritic Re/Os ratios.

### 6.1. Age determinations

Not many eclogite suites are dated, but those that are give Archean ages that most likely reflect emplacement into the lithosphere. Jagoutz et al. (1984) obtained a  $2.7 \pm 0.1$  Ga Sm–Nd isochron age for recalculated bulk eclogites from Roberts Victor. Jacob and Foley (1999) reported a Pb–Pb isochron age for eclogites from the Udachnaya kimberlite, Siberia of  $2.57 \pm 0.2$  Ga. This is within error of the age obtained using Re–Os on bulk eclogites for this suite by Pearson et al. (1995b) of  $2.9 \pm 0.4$  Ga. Osmium isotopic data, although a popular and useful tool in dating eclogites, often show considerable scatter and ages are burdened with relatively large errors. Never-

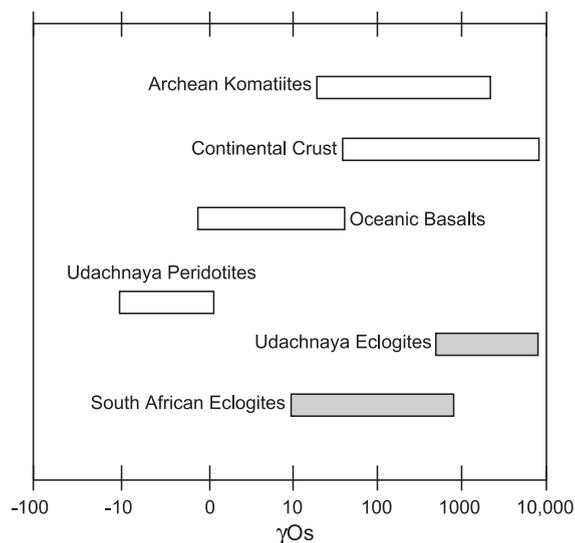


Fig. 13. Osmium isotopic compositions of mantle eclogites (modified after Pearson et al., 1995b).

theless, Re–Os isotopic systematics clearly indicate Archean ages for the Koidu and Newlands eclogites suite ( $3.44 \pm 0.76$  Ga: Barth et al., 2002a and  $3.6 \pm 0.6$  Ga: Menzies et al., 1998). Eclogites from the Slave craton (Jericho and Ekati/Lac de Gras pipes)

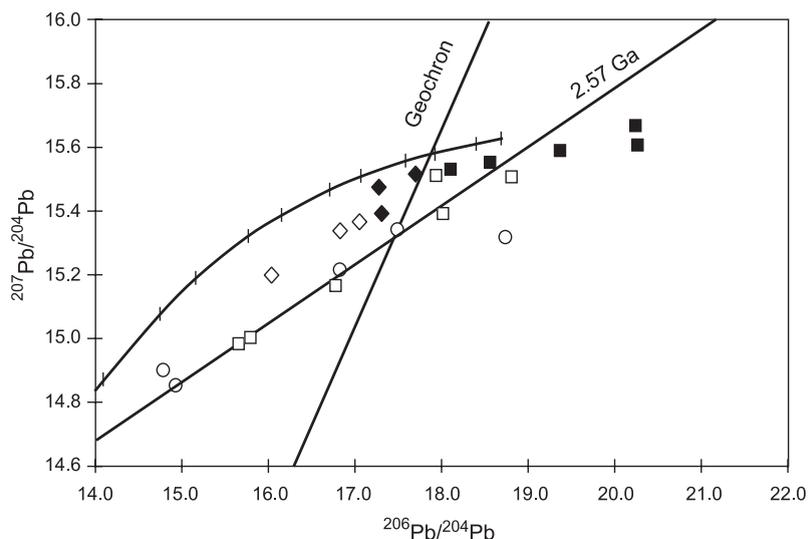


Fig. 12. Lead isotopic compositions of cpx (open symbols) and garnets (solid symbols) from mantle eclogites. Squares are samples from Udachnaya (Siberia, from Jacob and Foley, 1999) with a fitted 2.57 Ga isochron, interpreted as the age of eclogitization. Diamonds and circles are eclogites from Roberts Victor, South Africa. (Jacob and Jagoutz, 1995; Kramers, 1977, 1979). Also shown are the Geochron for 4.55 Ga, and the one-stage Pb model evolution line (Stacey and Kramers, 1975).

yield Proterozoic Nd-model ages (1.0–1.3 Ga, Heaman et al., 2003; Jacob et al., 2003a) that record carbonatite metasomatism within the eclogite facies. Eclogitization itself is currently estimated to be at least 1.79 Ga (Heaman et al., 2003), possibly connected to east-dipping subduction beneath the Slave craton during the 1.88–1.84 Ga Great Bear Magmatic arc event.

Mineral isochrons show a much more complicated picture. Eclogite xenoliths are stored in the subcratonic lithosphere at significantly higher temperatures than the closure temperature for Nd diffusion in pyrope (ca. 800 °C, van Orman et al., 2002) and garnet–clinopyroxene isochrons could therefore be expected to yield kimberlite emplacement ages for the eclogites. However, apparent Sm–Nd internal ages scatter over several orders of magnitude between 4 Ga and ages in the future, but only 7 out of 94 eclogite xenoliths worldwide show Sm–Nd isotopic equilibrium at the time of kimberlite emplacement. The most extreme apparent ages can occur among kyanite/corundum-bearing eclogites from the Kaalvallei and Bellsbank kimberlites (Neal et al., 1990; Viljoen, 1994), but eclogite suites from kimberlites worldwide show ages that deviate more than 100% from the age of the respective kimberlite. One-third of the database yield apparent ages in the future (so-called futurechrons, Jagoutz, 1995), i.e. the regression line between cpx and garnet shows a negative slope.

Some of the mineral ages can be attributed to preserved older equilibration events. Jagoutz (1988), for example showed in a very detailed study on a single eclogite xenolith from Tanzania that its Sm–Nd isotopic ratios evolved in a closed system to vastly different  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios between the mineral phases (>200‰ difference), and thus demonstrated that the garnet–cpx age of  $1.75 \pm 0.014$  Ga preserved information on an ancient equilibration event. Many other apparent inter-mineral ages might represent chemical equilibria that are “frozen in” near the closure temperature (Dodson, 1973) or, as has been shown recently, reflect slow cooling of the lithosphere at a temperature range above the closure temperature of the respective isotopic system (Albarede, 2003). Apparent future ages, however, cannot be explained by these hypotheses, but are results of metamorphic reactions involving finite

diffusion couples (Jenkin et al., 1995; Jacob and Jagoutz, in preparation). Similar to processes in high-grade metamorphic rocks (e.g. Mørk and Mearns, 1986), minerals newly formed during reactions such as exsolution or prograde formation of garnet from plagioclase inherit their precursors isotopic and trace element characteristics. As garnet and cpx in bimineraleclogites have very different and rather complementary equilibrium concentrations of REE, they represent finite diffusion couples for these elements. In this case, the rate of diffusion strongly depends on the concentration of the specific element in the exchange partner and diffusional equilibrium may not be reached over large geological timescales. In kyanite- or corundum-bearing eclogites these low-REE phases limit bulk diffusional exchange between cpx and garnet even further, and chemical heterogeneities are effectively preserved leading to extreme aberrant internal ages, as described above.

## 6.2. Oxygen isotopic compositions

Garlick et al. (1971) were the first to establish the unusually wide range of  $\delta^{18}\text{O}$ -values of mantle eclogites, now known to be characteristic of many eclogite suites worldwide (Fig. 14). Later studies established a minimal pressure-dependence of oxygen isotopic fractionation (Clayton et al., 1975) and a rather constant oxygen isotopic composition of the unchanged Earth’s mantle of  $5.5 \pm 0.4\text{‰}$  (Mattey et al., 1994) and provided a framework for the interpretation of mantle eclogites as subducted seawater-altered oceanic crust. The oxygen isotope argument still plays a key role in the subduction hypothesis and is based on similarities between the ranges of  $\delta^{18}\text{O}$  values observed in eclogite xenoliths and those of ophiolites (e.g. Gregory and Taylor, 1981) and modern oceanic crust (e.g. Alt et al., 1989). Low  $\delta^{18}\text{O}$  values can only be caused close to the Earth’s surface by hydrothermal alteration at elevated temperatures, such as in the dike sections of modern oceanic crust. Especially the low  $\delta^{18}\text{O}$  values found in eclogite suites from kimberlites are therefore truly indicative of their derivation from subducted altered oceanic crust.

Eclogites from the Roberts Victor mine show the widest range of  $\delta^{18}\text{O}$ -values from 2.5‰ to 8.0‰

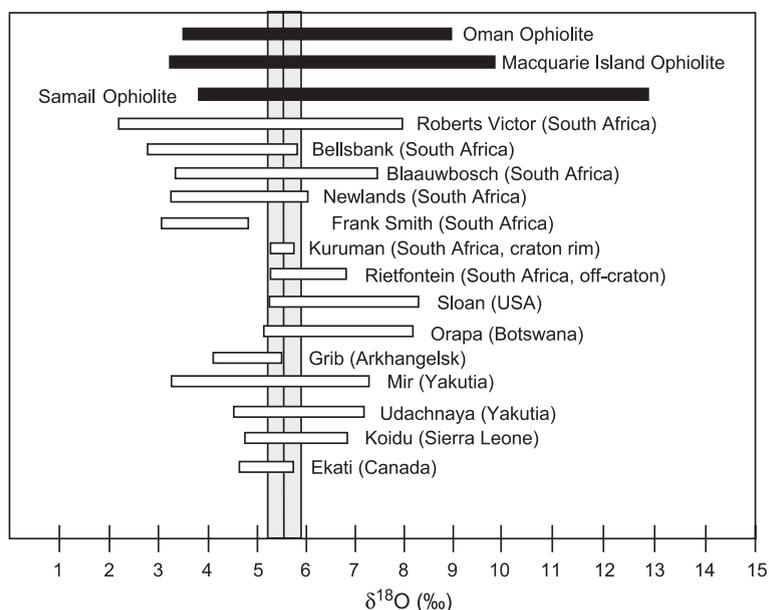


Fig. 14. Oxygen isotopic compositions of mantle eclogite suites worldwide compared to those of ophiolites (black bars). Vertical line denotes  $\delta^{18}\text{O}$  value of unchanged mantle with a grey error envelope ( $5.5\text{‰} \pm 0.4$ , Matthey et al., 1994). Note that only the Kuruman suite falls into the mantle field, whereas  $\delta^{18}\text{O}$ -values of all other suites deviate significantly. (Data sources: Oman ophiolite: Gregory and Taylor, 1981; Macquarie Island ophiolite: Cocker et al., 1982; Semail ophiolite: Gregory and Taylor, 1981; Roberts Victor eclogites: MacGregor and Manton, 1986; McDade, 1999, Lowry unpubl. data; Bellsbank: Neal et al., 1990; Blaauwbosch, Newlands, Frank Smith and Sloan: Schulze et al., 2003b; Kuruman: Schmickler et al., in press; Rietfontein: Appleyard, 2003; Orapa: Viljoen et al., 1996; Grib: Malkovets et al., 2003; Mir: Beard et al., 1996; Udachnaya: Jacob et al., 1994; Snyder et al., 1997; Koidu: Barth et al., 2001, 2002b; Ekati: Jacob et al., 2003a; Aulbach, 2003.)

(Fig. 14) and this variation originally measured by conventional fluorination techniques was confirmed by laser fluorination analysis (2.40‰ to 6.98‰, Lowry unpubl. data). Fractionation between clinopyroxene and garnet is approximately 0.3‰, the laser technique giving more systematic fractionation factors than the conventional method.

## 7. Discussion

The extreme range of possible solid solution in the clinopyroxene and garnet structures in the eclogite metamorphic facies can accommodate a wide variety of bulk compositions (Spear, 1993) and it is therefore not surprising that mantle eclogites represent a rather heterogeneous group of rocks. This fact alone shows that it is not justified to postulate a single origin for all eclogites found in the Earth's mantle. To model their petrogenesis can be rather challenging, as a number of modifying processes need to be taken into account.

Most prominent are the effects of kimberlite infiltration that generally can be bypassed by using ultra clean mineral separates for isotopic analyses and in situ methods for trace element determinations. However, the benefits are limited when pronounced metasomatism leads to growth of clinopyroxene and/or garnet. Further care has to be taken in recalculation of clean bulk compositions as accessory phases, e.g. rutile or quartz can easily be overlooked. Quartz is especially vulnerable as it is not in equilibrium with the quartz-undersaturated kimberlitic melt and, in addition to the fact that it maybe overlooked, reacts out easily. Furthermore, most eclogites from the Earth's mantle show trace element depletions and are thus residues after partial melting. Even coesite/quartz-bearing specimens show evidence for this (Jacob et al., 2003b), although the presence of free  $\text{SiO}_2$  limits the amount of partial melt lost to ca. 13% (Yaxley and Green, 1998). All these effects need to be recognized and unravelled in each individual sample, because their time-integrated effects have influence on

the radiogenic isotopic ratios and may lead to erroneous interpretation.

## 8. Hypotheses for the origin of eclogites evaluated

### 8.1. Recycling of seawater-altered oceanic crust

Many eclogites show evidence for protoliths of low-pressure (i.e. not mantle) origin. A crustal origin is well constrained for those eclogites that contain coesite/quartz or kyanite, because these phases are not in equilibrium with the peridotitic paragenesis in the Earth's mantle. Phase equilibrium considerations show that at pressures within the diamond stability field kyanite reacts with olivine to form pyroxenes and garnet, thus kyanite-bearing eclogites cannot be high-pressure cumulates from melts derived from garnet-peridotite (Jacob et al., 1998). The occurrence of coesite/quartz in eclogites also excludes a direct origin of the eclogites as melts from the Earth's mantle. Partial melts of peridotite are not quartz-normative at any pressure higher than 0.8 GPa, but are olivine + hypersthene-normative at higher pressures (Green and Falloon, 1998). Although free silica may occur in olivine tholeiites or olivine basalt compositions in eclogite facies conditions, these melts could not have originated by melting at pressures greater than 2 GPa (Green and Falloon, 1998).

Geochemical evidence for low-pressure protoliths comes from distinct trace element patterns with positive Eu- and Sr-anomalies and flat HREE in reconstructed bulk compositions (Fig. 8) that can be found not only in coesite/quartz- and kyanite-bearing eclogites, but also in many bimineralic samples. These trace element patterns are very unusual for garnet-bearing high-pressure rocks, but are characteristic of metamorphosed plagioclase-bearing rocks (Mørk and Brunfelt, 1988; Mørk and Mearns, 1986). These characteristics clearly point to a low-pressure origin, but the key arguments for a derivation from oceanic crust are the variation in  $\delta^{18}\text{O}$  values observed in most suites of eclogites (Fig. 14). Clayton et al. (1975) showed that the effect of pressure on the fractionation of oxygen isotopes is negligible, thus excluding fractionation processes at mantle pressures and temperatures as a reason for the observed variations. However, seawater alteration of oceanic crust at

different temperatures, as recorded in modern oceanic basalts and by ophiolites, causes very similar (although more extreme) ranges of  $\delta^{18}\text{O}$  values (e.g. Gregory and Taylor, 1981). The match with the oxygen isotopic variation in mantle eclogites is close, although not identical, which has led to criticism regarding its proposed analogous origin (e.g. Haggerty, 1999). However, more extreme  $\delta^{18}\text{O}$  values of +16.1‰ are reported for coesites included in diamond (translating to +14.6‰ for hypothetical garnet in equilibrium with this coesite (Schulze et al., 2003a). This shows that minerals shielded by diamond from equilibration with surrounding mantle do indeed preserve their original more extreme  $\delta^{18}\text{O}$  values.

Within ophiolites and modern oceanic crust, the variations in  $\delta^{18}\text{O}$  values are caused by alteration by circulating seawater at variable temperatures, leading to elevated  $\delta^{18}\text{O}$  values at low temperatures, close to the seafloor and low  $\delta^{18}\text{O}$  values at higher temperatures (see for example Alt, 1995). In contrast to meteoric water that can have extreme negative  $\delta^{18}\text{O}$  values, seawater has a  $\delta^{18}\text{O}$  value close to 0, and thus shifts the isotopic composition of the altered rocks to significantly less extreme values (Fig. 15). This provides a tool to differentiate between those eclogites whose precursors were altered by meteoric waters (e.g. at Dabie Shan, China, Yui et al., 1997) and those that show a seawater alteration signature. Fig. 15 also illustrates that apparently unchanged  $\delta^{18}\text{O}$  values in individual samples are not necessarily evidence for an unaltered state of the rocks. The  $\delta^{18}\text{O}$  trend caused by seawater alteration shows a cross-over with the magmatic fractionation line which is simply a function of the endmember compositions and creates magmatic, apparently unchanged  $\delta^{18}\text{O}$  values in strongly seawater altered samples (Gregory and Taylor, 1981).

Further supporting evidence for a seawater alteration history of the protoliths of mantle eclogites can be found in trace element and isotopic compositions (e.g. Jacob et al., 1994), as the effects of this process on the geochemical signature of oceanic crust are well constrained. Perhaps most significant are Cs concentrations in eclogitic minerals from Udachnaya and Roberts Victor that overlap with those of MORB (Jacob et al., 1994). Due to the incompatible behaviour of the alkali elements, high-pressure cumulate rocks from melts from garnet peridotite are not expected to contain measurable Cs concentrations,

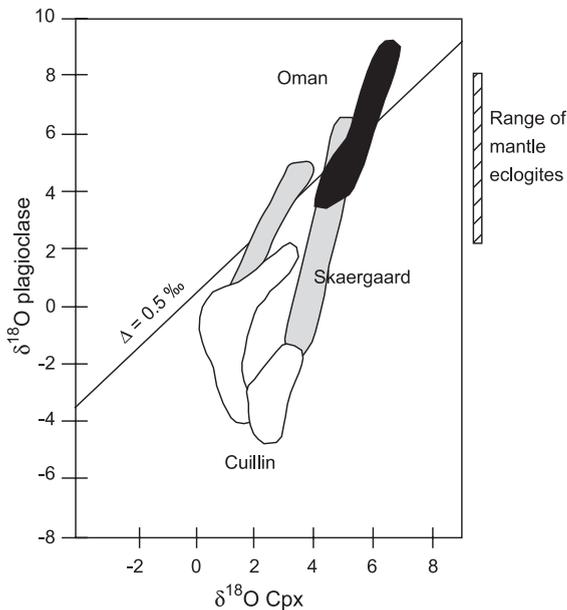


Fig. 15. Oxygen fractionation between plagioclase and clinopyroxene in layered intrusions (Skaergaard and Cuillin) altered by meteoric waters compared to fractionation effects in the seawater-altered Oman ophiolite (redrawn after Gregory and Taylor, 1981). Meteoric waters have more extreme (negative)  $\delta^{18}\text{O}$ -values than seawater and consequently generate more extreme ranges in altered mineral pairs. The black line denotes the magmatic equilibrium fractionation between clinopyroxene and plagioclase. Note that, although altered by seawater with a  $\delta^{18}\text{O}$  value of 0, some minerals, by coincidence, fall on the magmatic fractionation line.

whereas oceanic basalts contain significant amounts (Hofmann and White, 1982).

An origin of a *suite* of mantle eclogites from seawater-altered oceanic protoliths is well justified in cases where evidence for plagioclase in the protolith (“ghost-plagioclase signatures”, such as positive Eu-anomalies) coincides with variations in  $\delta^{18}\text{O}$  values. However, it is more difficult on the scale of *individual* samples for the following reasons: an apparently pristine  $\delta^{18}\text{O}$  value similar to that of the unchanged Earth’s mantle does not necessarily exclude a seawater alteration history due to the cross-over issue outlined above. Furthermore, as the oceanic crust is made up of basalts and ultramafic cumulates in addition to gabbros, a lack of “ghost-plagioclase” signature, indicative for a gabbro precursor rock, does also not necessarily exclude an oceanic origin for the eclogite protoliths.

## 8.2. Models explaining eclogites as cumulates

Models explaining mantle eclogites as cumulates from primary melts from garnet peridotite were the first to be put forward (e.g. O’Hara, 1969; O’Hara et al., 1975). One of the major problems with this hypothesis, however, is that the accumulation of garnet and clinopyroxene as liquidus minerals from a mantle-derived melt is unlikely; the first phase to accumulate from a primary mantle melt in geologically realistic conditions should be olivine, which is a rare mineral in mantle eclogites.

This is shown in Fig. 16 using the example of high-pressure experimental results on a synthetic olivine tholeiite (Green and Ringwood, 1967). Garnet and clinopyroxene occur together at the liquidus at a pressure of 2.5 GPa for this rock, and close to the liquidus above 1.7 GPa. They are replaced to lower pressures by orthopyroxene alone and then by olivine. Although it appears that it is reasonable that garnet and clinopyroxene should co-crystallize from this

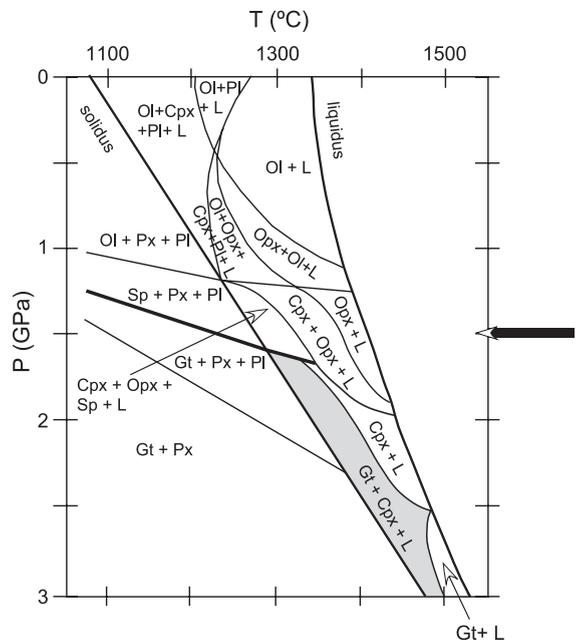


Fig. 16. Phase diagram of olivine tholeiite (modified after Green and Ringwood, 1967). Grey field shows  $P$ - $T$  area in which accumulation of cpx and garnet occurs. This is, however, at greater pressures than the origin of the magma (black arrow, Jaques and Green, 1980; Falloon and Green, 1988) and therefore geodynamically impossible. Fat black line is the garnet-in line.

melt, it must be taken into account that the pressure of origin for olivine tholeiites is known to be on the order of 1.5–1.7 GPa from partial melting experiments on peridotite (arrow in Fig. 16; Jaques and Green, 1980; Falloon and Green, 1988), meaning that the garnet and clinopyroxene liquidus fields are exclusively at higher pressures than the origin of the melt. This would require the geodynamically unlikely situation of displacement of a melt to higher pressures in order to generate eclogites by accumulation.

The lack of a multiphase saturation point at the liquidus in Fig. 16 demonstrates that the rock does not represent a primary melt, but that it must have lost olivine, which would be the liquidus phase at pressures below the pressure of origin (Foley, 1989). Although this is only an example it is a general rule that multiphase saturation points, although required for primary melts, are generally not found in experimental studies of basaltic rocks. Furthermore, garnet is mostly restricted to pressures above 1.5 GPa even at temperatures well below the liquidus, so that considerable fractionation must occur to bring it to the liquidus. This means that garnet should be an intercumulus, rather than a cumulus phase in any mantle cumulate derived from a partial melt of peridotite. Garnet may eventually crystallize as a liquidus mineral from a strongly fractionated rest-melt, but in this case the Mg-number of the garnet would not correspond to those seen in the eclogite xenoliths, but instead would be much lower. Phase topologies similar to that shown in Fig. 16 are commonly found in experiments on basaltic rocks, indicating that these conclusions are generally applicable.

The formation of cumulates from primitive melts may be modelled using the pMELTS program (Ghiorso and Hirschmann, 2002) and this has been applied to investigate the possible formation of eclogite suites by Jacob et al. (2003a) and Schmickler et al. (in press). For the Kuruman eclogite suite, it could be shown that the eclogites can be explained by accumulation from a picritic melt, but only by low pressure accumulation of olivine, clinopyroxene and plagioclase followed by subsequent metamorphic transformation to eclogites by an increase in pressure. The pressure of accumulation of around 1 GPa is too high for modern oceanic crust, but may apply for thick ocean crust in late Archean or earliest Proterozoic times (Schmickler et al., in press). Al-

though this may appear to imply a convergence of the originally opposed subduction and mantle cumulate hypotheses, it retains the main feature of the subduction hypothesis that eclogite does not result directly by crystallization of garnet and clinopyroxene from a melt.

In summary, the vast majority of eclogite suites are dominated by samples that correspond to parts of subducted oceanic crust; either volcanics or plagioclase-bearing cumulates. Recent results have revived the idea of an origin by accumulation of liquidus minerals from mantle-derived melts, but at pressures below the eclogite facies followed by eclogite metamorphism.

### Acknowledgements

Reviews by Dan Schulze and Chris Hatton are gratefully acknowledged. C. Appleyard, P. McDade, K.S. Viljoen, C. Hatton and D. Lowry generously provided unpublished Theses and data. Financial support from the 8IKC and the DFG made my attendance to the Conference and this publication possible.

### References

- Ai, Y., 1994. A revision of the garnet–clinopyroxene Fe<sup>2+</sup>–Mg exchange geothermometer. *Contrib. Mineral. Petrol.* 115 (4), 467–473.
- Albarede, F., 2003. The thermal history of leaky chronometers above their closure temperature. *Geophys. Res. Lett.* 30 (1), 1015 (doi 10.1029/2002GL016484).
- Albarede, F., van der Hilst, R.D., 2002. Zoned mantle convection. *Philos. Trans. R. Soc. Lond., A* 360 (1800), 2569–2592.
- Alt, J.C., 1995. Subseafloor processes in Mid-ocean ridge hydrothermal systems. In: Humphris, S.E., Zierenberg, R.A., Mullineaux, L.S., Thomson, R.E. (Eds.), *Seafloor Hydrothermal Systems: Physical, Chemical, Biological and Geological Interactions*. AGU, Washington, pp. 85–114.
- Alt, J.C., Anderson, T.F., Bonnell, L., Muehlenbachs, K., 1989. Mineralogy, chemistry, and stable isotopic composition of hydrothermally altered sheeted dikes: ODP hole 504B, Leg 111. *Proc. ODP, Sci. Results* 111, 27–40.
- Aoki, K., Shiba, I., 1973. Pyroxenes from lherzolite inclusions of Itinomegata, Japan. *Lithos* 6, 41–51.
- Appleyard, C., 2003. The geochemistry of a suite of eclogite xenoliths from the Rietfontein kimberlite, South Africa. *Ext. Abstr. 8th Internat. Kimberlite Conference*.
- Aulbach, S., 2003. The lithospheric mantle beneath the Slave

- Craton and Alberta, Canada, PhD thesis, Macquarie University. 282 pp.
- Bach, W., Alt, J.C., Nau, Y., Humphris, S.E., Erzinger, J., Dick, H.J.B., 2001. The geochemical consequences of late-stage low-grade alteration of lower ocean crust at the SW Indian Ridge: results from ODP Hole 735B (Leg 176). *Geochim. Cosmochim. Acta* 65, 3267–3288.
- Barth, M., Rudnick, R.L., Horn, I., McDonough, W.F., Spicuzza, M., Valley, J.W., Haggerty, S.E., 2001. Geochemistry of xenolithic eclogites from West Africa: Part I. A link between low MgO eclogites and Archean crust formation. *Geochim. Cosmochim. Acta* 65, 1499–1527.
- Barth, M., Rudnick, R.L., Carlson, R.W., Horn, I., 2002a. Re–Os and U–Pb geochronological constraints on the eclogite–tonalite connection in the Archean Man Shield, West Africa. *Precambrian Res.* 118, 267–283.
- Barth, M., Rudnick, R.L., Horn, I., McDonough, W.F., Spicuzza, M., Valley, J.W., Haggerty, S.E., 2002b. Geochemistry of xenolithic eclogites from West Africa: Part II. Origins of the high MgO eclogites. *Geochim. Cosmochim. Acta* 66, 4325–4345.
- Bonney, T.G., 1899. The parent-rock of the diamond in South Africa. *Geol. Mag.* 6, 309–321.
- Brey, G.P., Köhler, T., 1990. Geothermobarometry in four-phase lherzolites: II. New thermobarometers, and practical assessment of existing thermobarometers. *J. Petrol.* 31, 1353–1378.
- Caporuscio, F.A., Smyth, J.R., 1990. Trace element crystal chemistry of mantle eclogites. *Contrib. Mineral. Petrol.* 105, 550–561.
- Clayton, R.N., Goldsmith, J.R., Karel, V.J., Mayeda, T.K., Newton, R.C., 1975. Limits on the effect of pressure on isotopic fractionation. *Geochim. Cosmochim. Acta* 39, 1197–1201.
- Cocker, J.D., Griffin, B.J., Muehlenbachs, K., 1982. Oxygen and carbon isotope evidence for seawater-hydrothermal alteration of the Macquarie Island ophiolite. *Earth Planet. Sci. Lett.* 61, 112–122.
- Coleman, R.G., Lee, D.E., Beatty, L.B., Brannock, W.W., 1965. Eclogites and eclogites: their differences and similarities. *Geol. Soc. Amer. Bull.* 76, 483–508.
- Cordery, M.J., Davies, G.F., Campbell, I.H., 1997. Genesis of flood basalts from eclogite-bearing mantle plumes. *J. Geophys. Res.* 102, 20179–20197.
- Dawson, J.B., 1984. Contrasting types of upper-mantle metasomatism. In: Kornprobst, J. (Ed.), *Kimberlites: II. The Mantle and Crust–mantle Relationships*. Elsevier, Amsterdam, pp. 289–294.
- Dodson, M.H., 1973. Closure temperature in cooling geochronological and petrological systems. *Contrib. Mineral. Petrol.* 40, 259–274.
- Eggler, D.H., McCallum, M.E., Kirkley, M.B., 1987. Kimberlite-transported nodules from Colorado–Wyoming: a record of enrichment of shallow portions of an infertile lithosphere. In: Morris, E.M., Pasteris, J.D. (Eds.), *Mantle Metasomatism and Alkaline Magmatism*. Special Paper - Geol. Soc. America, vol. 215, pp. 77–90.
- El Fadili, E.S., Demaiffe, D., 1999. Petrology of eclogite and granulite nodules from the Mbuji Mayi kimberlites (Kasai, Congo): significance of kyanite–omphacite intergrowths. In: Gurney, J.J., Gurney, J.L., Pascoe, M., Richardson, S.H. (Eds.), *Proc. 7th International Kimberlite Conference*. Red Roof Design cc, Cape Town, pp. 205–213.
- Ellis, D.J., Green, D.H., 1979. An experimental study of the effect of Ca upon garnet–clinopyroxene Fe–Mg exchange equilibria. *Contrib. Mineral. Petrol.* 71, 13–22.
- Erlank, A.J., Kushiro, I., 1970. Potassium contents of synthetic pyroxenes at high temperatures and pressures. *Carnegie Inst. Washington, Yearbook* 68, 433–439.
- Falloon, T.J., Green, D.H., 1988. Anhydrous partial melting of peridotite from 8 to 35 kb and the petrogenesis of MORB. *J. Petrol.*, 379–414 (Special Lithospheric Issue).
- Foley, S., 1989. The genesis of lamproitic magmas in a reduced fluorine-rich mantle. In: Ross, J., et al. (Eds.), *Kimberlites and Related Rocks*. Special Publication - Geol. Soc. Australia, vol. 14, pp. 616–631. Perth.
- Foley, S., Tiepolo, M., Vannucci, R., 2002. Growth of early continental crust controlled by melting of amphibolite in subduction zones. *Nature* 417 (6891), 837–840.
- Foley, S.F., Buhre, S., Jacob, D.E., 2003. Evolution of the Archean crust by delamination and shallow subduction. *Nature* 421 (6920), 249–252.
- Fouch, M.J., James, D.E., Vandecar, J.C., van der Lee, S. Kaapvaal Seismic Group, 2004. Mantle seismic structure beneath the Kaapvaal and Zimbabwe cratons. *S. Afr. J. Geol.* 107 (1–2), 35–46.
- Fung, A.T., 1998. Petrochemistry of upper mantle eclogites from the Grizzly, Leslie, Pigeon and Sable kimberlites in the Slave Province, Canada. *Ext. Abs. 7th. Int. Kimberlite Conf.*, pp. 230–232.
- Garlick, G.D., MacGregor, I.D., Vogel, D.E., 1971. Oxygen isotope ratios in eclogites from kimberlites. *Science* 172, 1025–1027.
- Ghiorso, M.S., Hirschmann, M.M., 2002. pMELTS: a revision of MELTS for improved calculation of phase relations and major element partitioning related to partial melting of the mantle to 3 GPa. *Geochem. Geophys. Geosyst.* 3, U1–U36 (May 31, 2002).
- Green, D.H., 1966. The origin of the ‘eclogites’ from Salt Lake Crater, Hawaii. *Earth Planet. Sci. Lett.* 1 (6), 103–190.
- Green, D.H., Falloon, T.J., 1998. Pyrolite: a Ringwood concept and its current expression. In: Jackson, I. (Ed.), *The Earth’s Mantle*. Cambridge Univ. Press, Melbourne, pp. 311–380.
- Green, D.H., Ringwood, A.E., 1967. The genesis of basaltic magmas. *Contrib. Mineral. Petrol.* 15, 217–229.
- Green, D.H., Sobolev, N.V., 1975. Coexisting garnets and ilmenites synthesized at high pressures from pyrolite and olivine basanite and their significance for kimberlitic assemblages. *Contrib. Mineral. Petrol.* 50, 217–229.
- Gregory, R.T., Taylor, H.P., 1981. An oxygen isotope profile in a section of cretaceous oceanic crust, Samail ophiolite, Oman: evidence for  $\delta^{18}\text{O}$  buffering of the oceans by deep (>5 km) seawater-hydrothermal circulation at mid-ocean ridges. *J. Geophys. Res.* 86 (B4), 2737–2755.
- Griffin, W.L., Carswell, D.A., Nixon, P.H., 1979. Lower crustal granulites from Lesotho, South Africa. In: Boyd, F.R., Meyer, H.O.A. (Eds.), *The Mantle Sample: Inclusions in Kimberlites and other Volcanics*. American Geophysical Union, pp. 59–86.
- Griffin, W.L., Ryan, C.G., Kaminsky, F.V., O’Reilly, S.Y., Natapov, L.M., Win, T.T., Kinny, P.D., Ilupin, I.P., 1999. The Siberian

- lithosphere traverse: mantle terranes and the assembly of the Siberian Craton. *Tectonophysics* 310 (1–4), 1–35.
- Gurney, J.J., 1990. The diamondiferous root of our wandering continents. *S. Afr. J. Geol.* 93, 424–437.
- Gurney, J.J., Moore, R.O., Otter, M.L., Kirkley, M.B., Hops, J.J., McCandless, T.E., 1991. Southern African kimberlites and their xenoliths. In: Kampunzu, A.B., Luballa, R.T. (Eds.), *Magma-tism in Extensional Structural Settings*. Springer, pp. 495–536.
- Haggerty, S.E., 1999. A diamond trilogy: superplumes, supercontinents, and supernovae. *Science* 285, 851–860.
- Haggerty, S.E., Fung, A.T., Burt, D.M., 1994. Apatite, phosphorus and titanium in eclogitic garnet from the upper mantle. *Geophys. Res. Lett.* 21, 1699–1702.
- Harlow, G.E., 1997. K in clinopyroxene at high pressure and temperature: an experimental study. *Am. Mineral.* 82, 259–269.
- Harte, B., Kirkley, M.B., 1997. Partitioning of trace elements between clinopyroxene and garnet: data from mantle eclogites. *Chem. Geol.* 136, 1–24.
- Hatton, C.J., 1978. The geochemistry and origin of xenoliths from the Roberts Victor mine. PhD thesis, University of Cape Town.
- Hatton, C.J., Gurney, J.J., 1987. Roberts Victor eclogites and their relation to the mantle. In: Nixon, P.H. (Ed.), *Mantle Xenoliths*. Wiley, London, pp. 453–463.
- Heaman, L.M., Creaser, R.A., Cookenboo, H.O., Chacko, T., 2003. Multi-stage modification of the mantle lithosphere beneath the Slave Craton: evidence from an unusual suite of zircon-bearing eclogite xenoliths entrained in the Jericho kimberlite, Canada. *Ext. Abs. 8th. Int. Kimberlite Conf.*
- Helmstaedt, H., Doig, R., 1975. Eclogite nodules from kimberlite pipes in the Colorado plateau—samples of subducted Franciscan type oceanic lithosphere. *Phys. Chem. Earth* 9, 95–111 (First international conference on kimberlites).
- Hills, D.V., Haggerty, S.E., 1989. Petrochemistry of eclogites from the Koidu Kimberlite Complex, Sierra Leone. *Contrib. Mineral. Petrol.* 103, 397–422.
- Hofmann, A.W., White, W.M., 1982. Ba, Rb and Cs in the Earth's mantle. *Z. Naturforsch.* 38a, 256–266.
- Ireland, T.R., Rudnick, R.L., Spetsius, Z., 1994. Trace elements in diamond inclusions reveal links to Archean granites. *Earth Planet. Sci. Lett.* 128, 199–213.
- Jacob, D.E., Foley, S.F., 1999. Evidence for Archean ocean crust with low high field strength element signature from diamondiferous eclogite xenoliths. *Lithos* 48, 317–336.
- Jacob, D., Jagoutz, E., 1995. A diamond–graphite bearing eclogitic xenolith from Roberts Victor (South Africa): indications for petrogenesis from Pb-, Nd- and Sr-isotopes. In: Meyer, H.O.A., Leonardos, O.H. (Eds.), *Kimberlites, Related Rocks and Mantle Xenoliths*. CPRM Spec. Publ., vol. 1/94, pp. 304–317.
- Jacob, D.E., Jagoutz, E., in preparation. Chemical Equilibrium between Garnet and Clinopyroxene in Eclogitic Rocks from Kimberlites.
- Jacob, D., Jagoutz, E., Lowry, D., Matthey, D., Kudrjavitseva, G., 1994. Diamondiferous eclogites from Siberia: remnants of Archean oceanic crust. *Geochim. Cosmochim. Acta* 58, 5191–5207.
- Jacob, D.E., Jagoutz, E., Lowry, D., Zinngrube, E., 1998. Comment on “The origin of Yakutian eclogite xenoliths”. *J. Petrol.* 39, 1527–1533.
- Jacob, D.E., Bizimis, M., Salters, V.J.M., 2002. Lu–Hf isotopic systematics of subducted ancient oceanic crust: Roberts Victor eclogites. *Geochim. Cosmochim. Acta* 66 (15A), A360.
- Jacob, D.E., Fung, A.T., Jagoutz, E., Pearson, D.G., 2003a. Petrology and geochemistry of eclogite xenoliths from the Ekati kimberlites area. *Ext. Abstr. 8th Internat. Kimberlite Conference*.
- Jacob, D.E., Schmickler, B., Schulze, D.J., 2003b. Trace element geochemistry of coesite-bearing eclogites from the Roberts Victor kimberlite, Kaapvaal craton. *Lithos* 71, 337–351.
- Jagoutz, E., 1988. Nd and Sr systematics in an eclogite xenolith from Tanzania: evidence for frozen mineral equilibria in the continental mantle. *Geochim. Cosmochim. Acta* 52, 1285–1293.
- Jagoutz, E., 1995. Isotopic constraints on garnet equilibration. *Terra Abstr.* 7, 339.
- Jagoutz, E., Dawson, J.B., Hoernes, S., Spettel, B., Wänke, H., 1984. Anorthositic oceanic crust in the Archean Earth. 15th Lunar Planet. Sci. Conf., pp. 395–396. Abs.
- Jaques, A.L., Green, D.H., 1980. Anhydrous melting of peridotite at 0–15 Kbar pressure and the genesis of tholeiitic basalts. *Contrib. Mineral. Petrol.* 73 (3), 287–310.
- Jenkin, G.R.T., Rogers, G., Fallick, A.E., Farrow, C.M., 1995. Rb–Sr closure temperatures in bi-mineralic rocks: a mode effect and test for different diffusion models. *Chem. Geol.* 122, 227–240.
- Jerde, E.A., Taylor, L.A., Crozaz, G., Sobolev, N.V., 1993a. Exsolution of garnet within clinopyroxene of mantle eclogites: major- and trace-element chemistry. *Contrib. Mineral. Petrol.* 114, 148–159.
- Jerde, E.A., Taylor, L.A., Crozaz, G., Sobolev, N.V., Sobolev, V.N., 1993b. Diamondiferous eclogites from Yakutia, Siberia: evidence for a diversity of protoliths. *Contrib. Mineral. Petrol.* 114, 189–202.
- Kennedy, C.S., Kennedy, G.C., 1976. The equilibrium boundary between graphite and diamond. *J. Geophys. Res.* 81, 2467–2470.
- Klemme, S., Blundy, J.D., Wood, B.J., 2002. Experimental constraints on major and trace element partitioning during partial melting of eclogite. *Geochim. Cosmochim. Acta* 66, 3109–3123.
- Konzett, J., 1997. Phase relations and chemistry of Ti-rich K-rich-terite-bearing mantle assemblages: an experimental study to 8.0 GPa in a Ti-KNCMASH system. *Contrib. Mineral. Petrol.* 128, 385–404.
- Kopylova, M.G., Russell, J.K., Cookenboo, H.O., 1999. Mapping the lithosphere beneath the North Central Slave Craton, The J B Dawson volume. *Proceedings of the VIIth International Kimberlite Conference, Cape Town, vol. 1. Red Roof Design cc, pp. 468–479.*
- Kramers, J.D., 1977. Lead and strontium isotopes in cretaceous kimberlites and mantle-derived xenoliths from Southern Africa. *Earth Planet. Sci. Lett.* 34, 419–431.
- Kramers, J.D., 1979. Lead, uranium, strontium, potassium and rubidium in inclusion-bearing diamonds and mantle-derived xenoliths from Southern Africa. *Earth Planet. Sci. Lett.* 42, 58–70.
- Lassiter, J.C., DePaolo, D.J., 1997. Plume/lithosphere interaction in the generation of continental and oceanic flood basalts:

- chemical and isotopic constraints. In: Mahoney, J.J. (Ed.), Large Igneous Provinces. American Geophysical Union, Monograph 100, pp. 335–355.
- MacGregor, I.D., Carter, J.L., 1970. The chemistry of clinopyroxenes and garnets of eclogite and peridotite xenoliths from the Roberts Victor mine, South Africa. *Phys. Earth Planet. Inter.* 3, 391–397.
- MacGregor, I.D., Manton, W.I., 1986. Roberts Victor eclogites: ancient oceanic crust. *J. Geophys. Res.* 91 (B14), 14063–14079.
- Malkovets, V., Taylor, L.A., Griffin, W., O'Reilly, S., Pokhilenko, N., Verichev, E., Golovin, N., Litasov, K., Valley, J., Spicuzza, M., 2003. Eclogites from the Grib Kimberlite Pipe Arkhangelsk, Russia. Ext. Abstr. 8th International Kimberlite Conference.
- Mattey, D., Lowry, D., MacPherson, C., 1994. Oxygen isotope composition of mantle peridotite. *Earth Planet. Sci. Lett.* 128, 231–241.
- McCandless, T.E., Gurney, J.J., 1989. Sodium in garnet and potassium in clinopyroxene; criteria for classifying mantle eclogites. In: Ross, J., Jaques, A.L., Ferguson, J., Green, D.H., O'Reilly, S.Y., Danchin, R.V., Janse, A.J.A. (Eds.), *Kimberlites and Related Rocks*. Special Publication—Geological Society of Australia, vol. 14, pp. 827–832. Perth.
- McCulloch, M.T., 1989. Sm–Nd systematics in eclogite and garnet peridotite nodules from kimberlites: implications for the early differentiation of the earth. In: Ross, J., Jaques, A.L., Ferguson, J., Green, D.H., O'Reilly, S.Y., Danchin, R.V., Janse, A.J.A. (Eds.), *Kimberlites and Related Rocks*. Special Publication—Geological Society of Australia, vol. 14, pp. 864–876. Perth.
- McDade, P., 1999. An Experimental Study of the Products of Extreme Metamorphic Processes: Ultrahigh-temperature Granulites and the Roberts Victor Eclogites University of Edinburgh. 390 pp.
- Menzies, A.H., Shirey, S.B., Carlson, R.W., Gurney, J.J., 1998. Re–Os isotope systematics of diamond bearing eclogites and peridotites from Newlands kimberlite. Ext. Abstr. 7th Internat. Kimberlite Conference, Cape Town, pp. 579–581.
- Mørk, M.B.E., Brunfelt, A.O., 1988. Geochemical comparisons of coronitic olivine gabbro and eclogites: metamorphic effects and the origin of eclogite protoliths (Flemsøy, Sunnmøre, Western Norway). *Nor. Geol. Tidsskr.* 68, 51–63.
- Mørk, M.B.E., Mearns, E.W., 1986. Sm–Nd isotopic systematics of a gabbro-eclogite transition. *Lithos* 19, 255–267.
- Neal, C.R., Taylor, L.A., Davidson, J.P., Holden, P., Halliday, A.N., Nixon, P.H., Paces, J.B., Clayton, R.N., Mayeda, T.K., 1990. Eclogites with oceanic crustal and mantle signatures from the Bellsbank kimberlite, South Africa, Part 2: Sr, Nd, and O isotope geochemistry. *Earth Planet. Sci. Lett.* 99, 362–379.
- O'Hara, M., 1969. The origin of eclogite and ariégite nodules in basalt. *Geol. Mag.* 106, 322–330.
- O'Hara, M.J., Yoder, H.S., 1967. Formation and fractionation of basic magmas at high pressure. *Scott. J. Geol.* 3, 67–117.
- O'Hara, M.J., Saunders, M.J., Mercy, E.L.P., 1975. Garnet–peridotite, primary ultrabasic magma and eclogite; interpretation of upper mantle processes in kimberlite. *Phys. Chem. Earth* 9, 571–604 (First international conference on kimberlites).
- Pearson, D.G., Shirey, S.B., Carlson, R.W., Boyd, F.R., Pokhilenko, N.P., Shimizu, N., 1995a. Re–Os, Sm–Nd, and Rb–Sr isotope evidence for thick Archean lithospheric mantle beneath the Siberian craton modified by multistage metasomatism. *Geochim. Cosmochim. Acta* 59, 959–978.
- Pearson, D.G., Snyder, G.A., Shirey, S.B., Taylor, L.A., Carlson, R.W., Sobolev, N.V., 1995b. Archean Re–Os age for Siberian eclogites and constraints on Archean tectonics. *Nature* 374, 711–713.
- Pearson, N.J., O'Reilly, S.Y., Griffin, W.L., 1995c. The crust–mantle beneath cratons and craton margins: a transect across the southwest margin of the Kaapvaal craton. *Lithos* 36, 257–287.
- Pearson, N.J., Griffin, W.L., Doyle, B.J., O'Reilly, S.Y., van Acherbergh, E., Kivi, K., 1999. Xenoliths from kimberlite pipes of the Lac de Gras area, Slave Craton, Canada. In: Gurney, J.J., Gurney, J.L., Pascoe, M., Richardson, S.H. (Eds.), *Proceedings of the VIIIth International Kimberlite Conference*, Cape Town, pp. 644–658.
- Pollack, H., Chapman, D., 1977. On the regional variation of heat flow, geotherms, and lithospheric thicknesses. *Tectonophysics* 38, 279–296.
- Rapp, R.P., Watson, E.B., 1995. Dehydration melting of metabasalt at 8–32 kbar. Implications for continental growth and crust–mantle recycling. *J. Petrol.* 36, 891–931.
- Ringwood, A., Green, D., 1966. An experimental investigation of the gabbro-eclogite transformation and some geophysical implications. *Tectonophysics* 3, 383–427.
- Robey, J.V., 1981. Kimberlites of the Central Cape Province, RSA. PhD thesis, University of Cape Town. 261 pp.
- Roden, M.F., Lazko, E.E., Jagoutz, E., 1999. The role of garnet pyroxenites in the Siberian lithosphere: evidence from the Mir kimberlite. In: Gurney, J.J., Gurney, J.L., Pascoe, M., Richardson, S.H. (Eds.), *Proc. 7th International Kimberlite Conference*. Red Roof Design cc, Cape Town, pp. 714–720.
- Rudnick, R.L., 1995. Eclogite xenoliths: samples of Archean ocean floor. Ext. Abstr. 6th Intern. Kimberlite, 473–475.
- Rudnick, R.L., Barth, M., Horn, I., McDonough, W.F., 2000. Rutile-bearing refractory eclogites: missing link between continents and depleted mantle. *Science* 287, 278–281.
- Schmickler, B., Jacob, D.E., Foley, S.F., 2004. Eclogite xenoliths from the Kuruman kimberlites, South Africa: geochemical fingerprinting of deep subduction and cumulate processes. *Lithos* (in press).
- Schulze, D.J., 1989. Constraints on the abundance of eclogite in the upper mantle. *J. Geophys. Res.* 94 (B4), 4205–4212.
- Schulze, D.J., Valley, J.W., Spicuzza, M., 2000. Coesite eclogites from the Roberts Victor kimberlite, South Africa. *Lithos* 54, 23–32.
- Schulze, D.J., Harte, B., Valley, J.W., Brenan, J.M., Channer, D.M.D.R., 2003a. Extreme crustal oxygen isotope signatures preserved in coesite in diamond. *Nature* 423, 68–70.
- Schulze, D.J., Valley, J.W., Viljoen, K.S., Spicuzza, M.J., 2003b. Oxygen isotope composition of mantle eclogites. Ext. Abstr. 8th Internat. Kimberlite Conference.
- Shee, S.R., Gurney, J.J., 1979. The mineralogy of xenoliths from Orapa, Botswana. In: Boyd, F.R., Meyer, H.O.A. (Eds.), *The Mantle Sample: Inclusions in Kimberlites and Related Rocks*. AGU, Washington, pp. 37–49.
- Shee, S.R., Bristow, J.W., Bell, D.R., Smith, C.B., Allsopp, H.L.,

- Lo, C.H., 1989. The petrology of kimberlites, related rocks and associated mantle xenoliths from the Kuruman Province, South Africa. In: Ross, J., Jaques, A.L., Ferguson, J., Green, D.H., O'Reilly, S.Y., Danchin, R.V., Janse, A.J.A. (Eds.), *Kimberlites and Related Rocks. Spec. Publ. - Geol. Soc. Australia*, vol. 14, pp. 60–82. Perth.
- Smith, C.B., et al., 1989. Sr and Nd isotopic systematics of diamond-bearing eclogitic xenoliths and eclogite inclusions in diamond from southern Africa. In: Ross, J., Jaques, A.L., Ferguson, J., Green, D.H., O'Reilly, S.Y., Danchin, R.V., Janse, A.J.A. (Eds.), *Kimberlites and Related Rocks. Special Publication - GSA*, vol. 14, pp. 853–863. Perth.
- Snyder, G.A., Taylor, L.A., Crozaz, G., Halliday, A.N., Beard, B.L., Sobolev, V.N., Sobolev, N.V., 1997. The origins of Yakutian eclogite xenoliths. *J. Petrol.* 38, 85–113.
- Sobolev, N.V., 1977. Deep-seated inclusions in kimberlites and the problem of the composition of the upper mantle. *American Geophysical Union, Washington, DC*. 279 pp.
- Sobolev, N.V., Lavrentev, Y.G., 1971. Isomorphic sodium admixture in garnets formed at high pressure. *Contrib. Mineral. Petrol.* 31, 1–12.
- Sobolev, N.V., Pokhilenko, N.P., Lavrentev, Y.G., Yefimova, E.S., 1977. Deep-seated xenoliths, xenocrysts in kimberlites and crystalline inclusions in diamonds from "Udachnaya" kimberlite pipe, Yakutia. *Ext. Abstr. 2nd Internat. Kimberlite Conference: Unpaginated*.
- Sobolev, V.N., Taylor, L.A., Snyder, G.A., 1994. Diamondiferous eclogites from the Udachnaya kimberlite pipe, Yakutia. *Int. Geol. Rev.* 36, 42–64.
- Spear, F.S., 1993. *Metamorphic Phase Equilibria and Pressure-Temperature-Time Paths*. Min. Soc. America Monograph, Washington, DC. 799 pp.
- Stacey, J.S., Kramers, J.D., 1975. Approximation of terrestrial lead isotope evolution by a two-stage model. *Earth Planet. Sci. Lett.* 26, 207–221.
- Stiefenhofer, J., Viljoen, K.S., Marsh, J.S., 1997. Petrology and geochemistry of peridotite xenoliths from the Letlhakane kimberlites, Botswana. *Contrib. Mineral. Petrol.* 127 (1–2), 147–158.
- Stosch, H.-G., Lugmair, G.W., 1990. Geochemistry and evolution of MORB-type eclogites from the Münchberg massif, southern Germany. *Earth Planet. Sci. Lett.* 99, 230–249.
- Taylor, L.A., Neal, C.R., 1989. Eclogites with oceanic crustal and mantle signatures from the Bellsbank kimberlite, South Africa. Part I: mineralogy, petrography and whole rock chemistry. *J. Geol.* 97 (5), 551–567.
- van der Hilst, R.D., 1995. Complex morphology of subducted lithosphere in the mantle beneath the Tonga trench. *Nature* 374, 154–157.
- van Orman, J.A., Grove, T.L., Shimizu, N., Layne, G.D., 2002. Rare earth element diffusion in a natural pyrope single crystal at 2.8 GPa. *Contrib. Mineral. Petrol.* 142 (4), 416–424.
- Viljoen, K.S., 1994. The petrology and geochemistry of a suite of mantle-derived eclogite xenoliths from the Kaalvallei kimberlite, South Africa. PhD thesis, Witwatersrand, Johannesburg.
- Viljoen, K.S., Smith, C.B., Sharp, Z.D., 1996. Stable and radiogenic isotope study of eclogite xenoliths from the Orapa kimberlite, Botswana. *Chem. Geol.* 131, 235–255.
- Woodland, A.B., Seitz, H.M., Altherr, R., Marschall, H., Olker, B., Ludwig, T., 2002. Li abundances in eclogite minerals; a clue to crustal or mantle origin? *Contrib. Mineral. Petrol.* 143 (5), 587–601.
- Yaxley, G.M., Green, D.H., 1998. Reactions between eclogite and peridotite: mantle refertilisation by subduction of oceanic crust. *Schweiz. Mineral. Petrogr. Mitt.* 78, 243–255.
- Yui, T.F., Rumble, D., Chen, C.H., Lo, C.H., 1997. Stable isotope characteristics of eclogites from the ultra-high-pressure metamorphic terrain, east-central China. *Chem. Geol.* 137 (1–2), 135–147.