

A Non-cognate Origin for the Gibeon Kimberlite Megacryst Suite, Namibia: Implications for the Origin of Namibian Kimberlites

G. R. DAVIES^{1*}, A. J. SPRIGGS² AND P. H. NIXON²

¹FACULTEIT DER AARDWETENSCHAPPEN, VRIJE UNIVERSITEIT, DE BOELELAAN 1085, 1081 HV AMSTERDAM, THE NETHERLANDS

²SCHOOL OF EARTH SCIENCES, UNIVERSITY OF LEEDS, LEEDS LS2 9JT, UK

RECEIVED DECEMBER 17, 1999; REVISED TYPESCRIPT ACCEPTED JUNE 28, 2000

Trace element and Sr–Nd–Pb isotope analyses are presented on unaltered kimberlites and clinopyroxenes and garnets of the low-Cr megacryst suite from the Gibeon Province, Namibia. Significant Sr and Pb isotope disequilibrium between the kimberlites and megacrysts establishes that the megacrysts are not cognate material. Calculated equilibrium melts for the megacrysts have rare earth element contents comparable with those of alkali basalts. Rb–Sr mica ages, ~ 72 Ma, demonstrate that kimberlite volcanism occurred between 5 and 10 my after the inferred passage of the Discovery plume beneath the Gibeon region. Sr–Nd–Pb isotope relationships of the kimberlite and megacrysts are distinct from that of the inferred plume and hence it is argued that the plume contributed little mass to the volcanism. The megacryst suite has a strong DUPAL Pb isotope signature. Two hypotheses can explain the genesis of the kimberlite and megacryst suites. The first is that the DUPAL Pb isotope signature is derived from the lower mantle. The megacryst suite therefore represents the high-pressure crystallization product of deep plume-related magmatism. This magmatism interacts with the sub-continental lithospheric mantle (SCLM) to produce the kimberlite magmatism. Alternatively, fluid-rich melts derived from the Discovery plume migrated under the lithosphere and become concentrated in areas that were recently thermally perturbed as-thenosphere, causing small degrees of melting and kimberlite magmatism. In this scenario the megacrysts represent polybaric fractionation products from ‘basaltic’ asthenospheric-derived melts that ponded at the base of, but underwent interaction with, the sub-continental lithosphere. Storage of the megacrysts for an extended period (>10 and <100 my) is required to explain the homogeneous

major and trace element compositions of individual megacrysts. Currently the latter explanation is favoured, on the assumption that the DUPAL geochemical signal is derived from the SCLM.

KEY WORDS: megacryst; Namibia; kimberlite; Sr–Nd–Pb isotopes

INTRODUCTION

Kimberlites have been subdivided into Group I (basaltic) and Group II (micaceous) on the basis of their petrography and geochemistry (Smith, 1983). Low Cr-megacrysts are a characteristic of the majority of Group I kimberlites. These megacrysts, also termed discrete nodules, comprise large (>1 cm), rounded, fractured single crystals. Mutual inclusions and lamella intergrowths define a suite that consists predominantly of pyrope garnet, magnesian ilmenite (picro-ilmenite), clinopyroxene (sub-calcic to varying degrees), orthopyroxene, zircon and possibly olivine. These minerals are compositionally distinct from those found in associated peridotite xenoliths, which coupled with different deformation rules out derivation from peridotites or eclogites (e.g. Mitchell, 1987). The association with kimberlite, plus the homogeneity of single crystals, which can be up to 40 cm across, led workers to

*Corresponding author. Telephone: +31 20 4447 329. Fax: +31 20 6462 457. E-mail: davg@geo.vu.nl

propose a cognate origin for the megacrysts (e.g. Nixon *et al.*, 1963; Nixon & Boyd, 1973; Pasteris *et al.*, 1979). These workers invoked megacryst crystallization from the host kimberlite over a long period of time. P - T estimates have been made on rare composite nodules and by assuming a cogenetic origin for all megacrystic phases. Such calculations imply equilibration over a large temperature range (1400–1000°C) at a relatively constant pressure that equates to the base of the mechanical boundary layer of the lithosphere (50 kbar) (e.g. Nixon & Boyd, 1973; Harte & Gurney, 1981; Nixon *et al.*, 1981; Schulze, 1987; Hops *et al.*, 1992).

Despite the appeal of a simple model with megacrysts representing high-pressure cogenetic material there remains considerable controversy regarding the origin of megacryst suites. Comparable compositional variations are found in megacryst suites from kimberlites, alnoites and alkali basalts that contain a garnet-bearing suite. This observation suggests that all megacrysts were formed from a common source, which cannot be the host volcanism. Although individual megacrysts are homogeneous, megacrysts from individual intrusions record marked chemical variations. Many workers have reported co-variation in MgO (gnt, cpx, opx), CaO (cpx, opx), Cr₂O₃ (ilm, cpx, opx) and TiO₂ (gnt) as evidence that the megacrysts crystallized from an evolving kimberlitic magma (e.g. Nixon & Boyd, 1973; Gurney *et al.*, 1979; Schulze, 1987). Kimberlite magmas, however, do not record chemical evidence of significant fractional crystallization (Mitchell, 1986). In addition, in trace element studies it has generally been concluded that the megacryst suite is not in chemical equilibrium with kimberlite (Harte, 1983; R. Jones, 1987). These observations have led to models that propose early fractionation of the megacryst suite during diapiric upwelling, possibly before kimberlite melt separation (Gurney *et al.*, 1979; Harte & Gurney, 1981; R. Jones, 1987).

Previous isotope studies have shown general similarities between megacrysts and kimberlites (e.g. Kramers, 1977, 1979; Basu & Tatsumoto, 1980; Smith, 1983; Griffin *et al.*, 2000), although some studies have shown subtle differences (R. Jones, 1987; Hops *et al.*, 1992). In most studies, however, the megacrysts and kimberlites were not from the same intrusion and the kimberlite samples were always affected by low-temperature alteration. From existing data it is therefore difficult to establish unequivocally if megacrysts are cogenetic with their host kimberlites. We report chemical and isotopic composition of garnet and clinopyroxene megacrysts and unaltered host rocks from the Gibeon Kimberlite Province, to constrain the petrogenesis of the megacryst suite and by implication the processes responsible for formation and modification of the host kimberlites.

REGIONAL GEOLOGY AND SAMPLE PREPARATION

The Gibeon kimberlite province, Namibia, occupies an area 100 km by 80 km centred at 25°30'S and 18°E (Fig. 1). In excess of 70 kimberlite pipes are known and they tend to occur in NNE–SSW trending clusters suggesting basement control to the intrusion location (Fig. 2). The kimberlite province intrudes part of the Late Precambrian to Cambrian Nama Group and the overlying Carboniferous to Cretaceous Karoo Sequence. The rarely exposed basement, generally referred to as the Namibian Province, was stabilized at ~2.1 Ga (Tankard *et al.*, 1982). This province lies between the Archaean Kaapvaal Craton and the Mid-Proterozoic Namaqua–Natal Belt. There are no direct geophysical data on the crustal thickness in the region. Garnet and two-pyroxene granulite xenoliths yield pressures of ~13 kbar (Van Heerden, 1984). This suggests a crustal thickness comparable with that of Proterozoic belts (42 km) and greater than that of the Kaapvaal Craton (35 km; Durrheim & Green, 1992). Peridotite xenoliths from the Gibeon kimberlites are predominantly coarse-grained garnet lherzolites, although harzburgites, garnet harzburgites and sheared lherzolites have been found (Mitchell, 1984). The peridotites have a significantly different composition from xenoliths from Kaapvaal kimberlites (Hoal *et al.*, 1995). Gibeon peridotites are characterized by higher modal proportions of olivine, possibly reflecting formation as residues following lower degrees of melt extraction than the Archaean lithosphere (Pearson *et al.*, 1994). Re depletion ages from these xenoliths reach 2.2 Ga (Pearson *et al.*, 1994; Hoal *et al.*, 1995) comparable with the age of the crust. P - T estimates for the coarse-grained lherzolite xenoliths fall close to the 44 mW/m² geotherm between 35 and 50 kbar, indicating derivation from the graphite stability field (Franz *et al.*, 1996). Sheared lherzolites have equilibrated over a similar pressure range but on an elevated geotherm ~50 mW/m² (Mitchell, 1984; Franz *et al.*, 1996). Combined gravity and teleseismic studies indicate that there is a decrease in the thickness of the subcontinental lithospheric mantle (SCLM) westward across Namibia away from the Kaapvaal Craton (e.g. Fairhead & Reeves, 1976). Heat flow measurements also imply thinner lithosphere in central and western Namibia (M. Jones, 1981, 1987).

Kimberlite is extremely vulnerable to alteration (e.g. Fesq *et al.*, 1975; Clement, 1982; Smith, 1983; Mitchell, 1986) such that fresh material can only be obtained from underground mines or drill cores. In addition, kimberlites tend to contain xenolith material derived from the crust and mantle. Namibian kimberlites typically contain 5–15% crustal material. Several hundred kimberlite drill cores were studied for obvious signs of alteration, and samples with calcite veins and altered olivines were

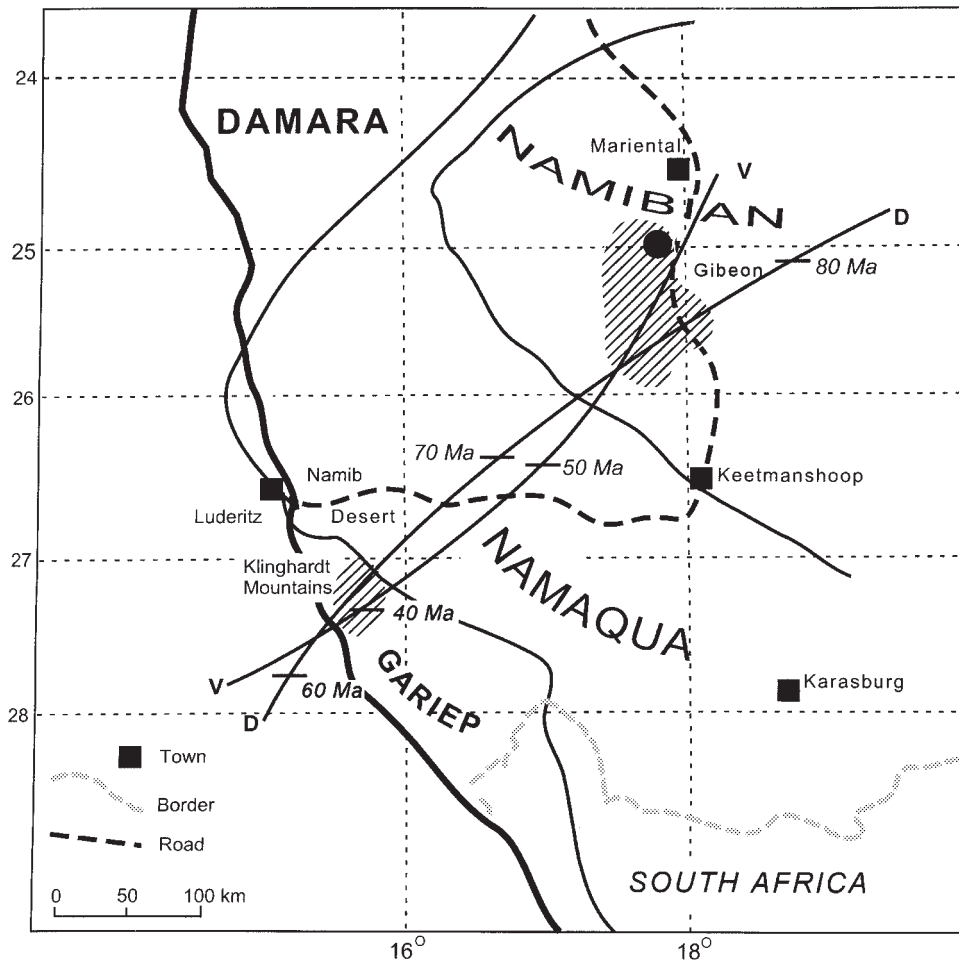


Fig. 1. Generalized location map of Namibia showing the distribution of volcanic rocks. The names of the major tectono-magmatic crustal provinces are shown in upper-case letters. Regions of volcanism marked by lined pattern. Continuous curves representing the tracks of the Vema (V) and Discovery (D) hotspots are from Hartnady & LeRoex (1983).

rejected. Subsequent petrographic examinations established samples with their primary kimberlite assemblage partially replaced by calcite, chlorite, serpentine, silica or clay minerals. Primary calcite and serpentine were identified on textural grounds. Petrographically unaltered samples were obtained only from drill cores that penetrated four kimberlites of hypabyssal facies. The petrology of each kimberlite locality is variable, implying either a very heterogeneous magma or that each locality contains several different intrusions. Multiple intrusions of kimberlite into the same pipe are a well-known phenomenon. Consequently, each sample is treated individually ($n = 16$). Core samples were coarsely crushed and crust and mantle xenoliths removed. Macrocrystic phlogopites were separated from the kimberlite and leached in 2M HCl for 10 min to remove any carbonate. Although fine-grained groundmass phlogopites are present in some samples, it proved impossible

to obtain multiple unaltered, inclusion-free separates of this material.

Garnet and clinopyroxene megacrysts were collected from the surfaces of seven diatreme facies kimberlites (Fig. 2; see Table 3, below). Unaltered megacrysts were selected to ensure no disruption of parent–daughter isotope ratios through alteration or interaction with the host kimberlite, which has concentrations of some trace element several orders of magnitude higher than the megacrysts. All samples were crushed, sieved and washed before hand picking under liquids. All mineral separates were treated ultrasonically in 6N HCl and subsequently washed several times in quartz-distilled H₂O. Sr isotope analyses were not performed on garnet megacrysts because the quality of the analyses is difficult to assess. This is due to garnets having extremely low Rb and Sr contents (<0.1 ppm), which makes them particularly susceptible to alteration. In addition, minor leaching during acid

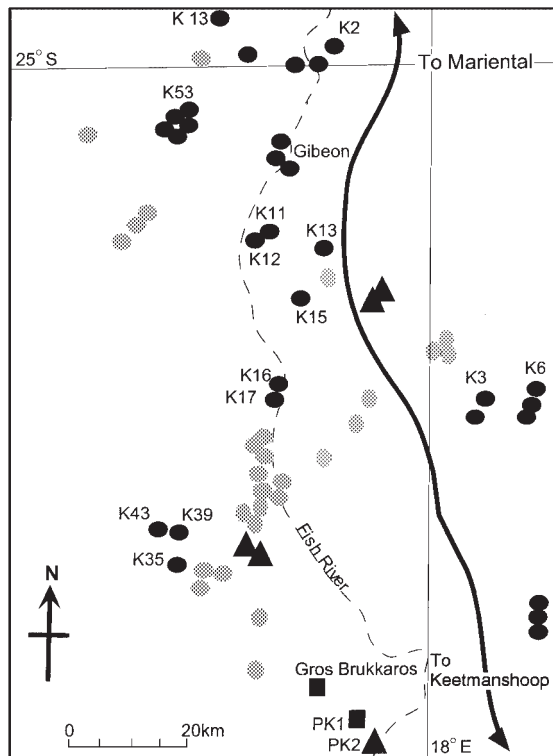


Fig. 2. Generalized location map of kimberlites in the Gibeon Province. Grey ellipses, kimberlites; filled ellipses, sampled kimberlites. Sample numbers indicate localities from which fresh samples or megacrysts were obtained. ■, carbonatite; △, associated melilitites. PK1 and PK2 are respectively carbonatite and monticellite peridotite dykes from the Blue Hills area for which fresh samples have been obtained.

treatment may significantly disrupt parent–daughter ratios, resulting in incorrect initial ratios.

A detailed discussion of the major element variations exhibited by the megacryst suite is beyond the scope of this study. Previous workers (Mitchell, 1987; Spriggs, 1988) have established that Namibian diopside and garnet megacrysts have large compositional ranges and concluded that, as a whole, the megacryst suite is generated by cumulate processes during the fractional crystallization of several batches of magma. Each kimberlite pipe in the Gibeon Province has its own characteristic megacryst suite that may have a large compositional range. For example, the Mukorob suite contains clinopyroxene with $\text{Ca}/(\text{Ca} + \text{Mg})$ ratios that vary from 0.35 to 0.45 and Cr_2O_3 contents of 0.1–0.9%. Such large compositional ranges imply formation from two or more magmatic sources.

ANALYTICAL TECHNIQUES

Major and trace element data were determined by a combination of X-ray fluorescence (XRF) and inductively

coupled plasma mass spectrometry (ICP-MS) at the University of Leeds and instrumental neutron activation analysis (INAA) at the Open University. Rare earth elements (REE) were determined on mineral separates by isotope dilution with standard ion exchange procedures (Thirlwall, 1982). Nd, Sr and Pb were measured at the University of Leeds. Data were normalized to $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.71024 for NBS 987 and 0.511845 for La Jolla. The routine total blanks at Leeds were Sr <2 ng, Nd <0.2 ng and Pb <0.25 ng. Between 100 and 150 mg of clinopyroxene was analysed for Pb isotopes. Sample sizes were sufficiently large that no significant blank correction is required. Further analytical details are given in Table 3 (below).

EMPLACEMENT AGES OF NAMIBIAN KIMBERLITES

To compare the Sr–Nd–Pb isotope systematics of the kimberlite and megacrysts the time of kimberlite eruption must be known to sufficient precision to allow comparison of initial isotopic ratios. The relatively low parent/daughter ratios of the kimberlites and clinopyroxene megacrysts means that initial Sr–Nd–Pb ratios can be compared provided ages are known to better than ± 20 my. Better precision is required to evaluate the relationship of the kimberlite volcanism to known hotspot tracks and to determine initial Nd isotope ratios of garnet megacrysts with high Sm/Nd ratios. Here, we report results of a Rb–Sr isotope study of phlogopitic micas from four Gibeon kimberlites. The low concentration of phlogopite in Gibeon kimberlites precludes the use of different size fractions to define a mineral isochron. Whole rock–mineral pairs are used to define two-point isochrons.

The Rb–Sr isotope results are presented in Table 1 (see Fig. 3). The macrocrystic phlogopites have $^{87}\text{Rb}/^{86}\text{Sr}$ ratios that range from 6.7 to 92.7. This large range is due to significant variations in Rb (210–500 ppm) and Sr (17.4–163 ppm) contents. There is no apparent major element control in phlogopite composition that helps predict high or low Rb/Sr ratios. Kimberlite K2 yields a mineral–whole-rock age of 71.3 ± 0.4 Ma. The phlogopite separate from sample K39 was a composite from several portions of drill core. Consequently, the Sr isotope heterogeneity recorded by the K39 pipe introduces an additional error to the age determination. Calculated ages, which vary from 70.9 ± 0.4 Ma to 71.3 ± 0.4 Ma, are within error of the age determined on kimberlite K2.

Fresh macrocrystalline separates were obtained from two altered kimberlites (K14 and K43). As a result of their altered state, whole-rock data from these kimberlites cannot be used to define an isochron. Below we will demonstrate that there is Sr and Nd isotope heterogeneity

Table 1: Rb–Sr Isotope data of phlogopites and host rocks

Sample	Rb (ppm)	Sr (ppm)	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	Age
<i>Kimberlites</i>						
K2/3/X WR	137	1915	0.072	0.2069	0.705530 ± 10	71.2 ± 0.4
K2 phlogopite*	210	84.56	2.483	7.189	0.712595 ± 16	
K39/W/M-WR	80	1264	0.063	0.1831	0.704048 ± 10	71.5 ± 0.4
K39 WR	81	1310	0.062	0.1788	0.704039 ± 10	71.0 ± 0.4
K39 phlogopite*	376.3	163.6	2.300	6.657	0.710634 ± 10	
K43 phlogopite*	552.6	17.39	31.78	92.66	0.787873 ± 18	
K35 WR	71	1099	0.065	0.1868	0.703801 ± 10	71.6 ± 0.4
K14 phlogopite*	434.4	95.94	4.528	13.11	0.716985 ± 10	
K53 WR	71	1089	0.065	0.1886	0.703839 ± 10	64.0 ± 0.4
<i>Gros Brukkaros carbonatite</i>						
Carbonate*	6.57	2749	0.0024	0.0069	0.703234 ± 8	73.6 ± 0.4
Phlogopite*	240.96	27.50	8.762	25.40	0.729796 ± 10	

*Rb and Sr determined by isotope dilution. Repetition of standard NBS 607 established that Rb is determined to better than 1% and Sr to <0.1%. Rb/Sr ratios determined by XRF on the host kimberlites are also better than 1%. Consequently a conservative error estimate of 1% is used in isochron calculations.

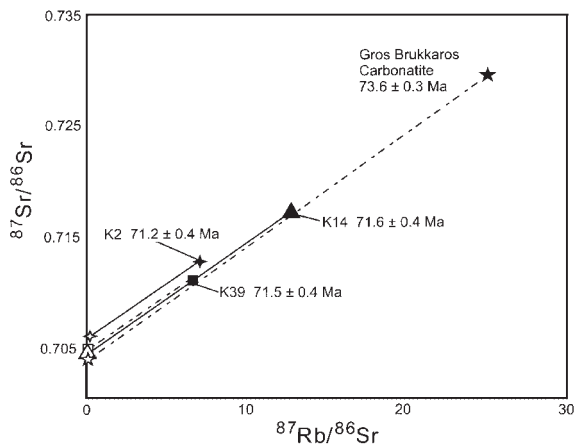


Fig. 3. Rb–Sr isochron diagram for phlogopitic micas from Gibeon kimberlites and the Gross Brukkaros carbonatite. Phlogopites represented by filled symbols. Open symbols are whole rocks.

in the kimberlites, possibly with a regional control. Therefore, to help constrain the eruption age of these two altered kimberlites we have used whole-rock Rb/Sr data from the nearest unaltered kimberlite. The phlogopite from K14 has an Rb/Sr ratio comparable with those of K2 and K39 and combined with a whole-rock analysis from the nearest fresh kimberlite (K53) yields an age of 71.6 ± 0.4 Ma, indistinguishable from those of pipes K2 and K39. Phlogopite from K43 has the most radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratio obtained in this study, 0.7879. Two analyses of K35, an adjacent kimberlite, are essentially

identical (see Table 3). The whole-rock phlogopite age of K43 is 64 ± 0.3 Ma, 7 my younger than the other samples. This younger age has three possible interpretations: (1) kimberlite magmatism in the Namibian Province was diachronous; (2) the extensive alteration of the kimberlite also disrupted the Rb/Sr systematics of the phlogopite; (3) we have made an invalid assumption regarding the unaltered whole-rock $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of K43. To obtain an older eruption age for K43 requires a lower whole-rock $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. K35 has the least radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the Namibian kimberlites and in Namibia only younger carbonatites have lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7032, e.g. Table 1; Spriggs, 1988). To produce a mica age close to 70 Ma an unrealistic whole-rock ratio of <0.700 is required. From the above data it is impossible to distinguish between the possibility of diachronous volcanism and disruption of the Rb/Sr systematics of the phlogopite. However, a 73.6 ± 0.4 Ma Rb–Sr biotite age for the adjacent Gross Brukkaros carbonatite complex (Table 1) appears to argue against significant diachronism in the regional volcanism.

Three micas define Rb/Sr ages that range from 71.0 to 71.6 Ma. Consequently, we feel confident in proposing an eruption age for the Gibeon kimberlites of ~ 71.5 Ma. It is important to note that the conclusions reached in the combined Sr–Nd–Pb isotope study of kimberlites and megacrysts presented below are unaffected by uncertainties of ± 5 Ma in the emplacement age of the kimberlites.

MEGACRYST REE DATA

There is a marked variation in the REE concentrations of garnet megacrysts (e.g. Yb contents range from 0.5 to 10.6 ppm; Table 2, Fig. 4a). Despite the large absolute REE variations there is little variation in REE fractionation; Sm/Nd ratios vary from 0.88 to 0.96. The clinopyroxene megacrysts also have significant variation in absolute REE abundance (Yb 0.12–0.17 ppm) with little variation in REE fractionation; Sm/Nd ratios vary from 0.253 to 0.256 (Fig. 4b). Clinopyroxenes are characterized by positive Eu anomalies (Eu^*/Eu up to 1.1) but garnets have no significant anomalies. Recent experimental studies have demonstrated that clinopyroxene will preferentially incorporate Eu^{2+} compared with other REE under oxidizing conditions and result in positive Eu anomalies (Wood *et al.*, 1999). The Namibian data therefore suggest that the clinopyroxenes crystallized at relatively high oxygen fugacity. By assuming that the megacryst suite is cogenetic, the REE data can be used to estimate clinopyroxene–garnet (cpx–gnt) REE partition coefficients. The calculated partition coefficients vary by over an order of magnitude for all REE (e.g. Nd 1–39; Yb 0.01–0.3). In contrast, published REE data for eclogites and garnet pyroxenites show limited K_d variation (e.g. Nd 2–9; Pearson *et al.*, 1993; Harte & Kirkley, 1997). The extreme variability of the REE cpx–gnt partition coefficients from Namibia strongly implies that they do not represent a cogenetic suite. Moreover, the clinopyroxenes record consistent heavy REE (HREE) fractionation. This observation clearly rules out clinopyroxene crystallization from magmas that had fractionated variable amounts of garnet and zircon. Although zircon is probably one of the latest phases to occur on the liquidus of the parental magma, even small amounts (<5%) of garnet fractionation would significantly fractionate light REE (LREE) from HREE in the residual liquid.

Sr–Nd–Pb ISOTOPE SYSTEMATICS OF NAMIBIAN KIMBERLITES AND MEGACRYSTS

The four kimberlite pipes have considerable variation in initial Sr and Nd isotope ratios; ϵ_{Nd} +1.6 to +4.0; ϵ_{Sr} –14 to +10 (Table 3, Fig. 5). These data are comparable with those for Group I kimberlites, to which the Namibian kimberlites have strong whole-rock chemical affinities (Smith, 1983; Fraser *et al.*, 1985; Smith *et al.*, 1985). Although pipe K2 has similar Nd isotope ratios to the other pipes it has $^{87}Sr/^{86}Sr$ ratios that trend to values more radiogenic than Group I kimberlites (Fig. 5). Compared with other Namibian kimberlites and typical Group I kimberlites, K2 has higher Ba, Rb, Sr, K and P

concentrations (Spriggs, 1988). The style of trace element enrichment (e.g. higher K/Nb, K/Ti and Ba/Nb) is similar to, although less extreme than, that of Group II kimberlites, which are also generally characterized by relatively radiogenic $^{87}Sr/^{86}Sr$ (Fraser *et al.*, 1985).

Present-day and initial Pb isotope ratios of Namibian kimberlites have relatively high $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios (Fig. 6) comparable with those of Group I kimberlites. Pipe K2 is again transitional towards Group II kimberlite compositions in having less radiogenic $^{206}Pb/^{204}Pb$. All Namibian kimberlites have high μ and κ values (50–80, 220–310). Pipe K2 has the lowest μ and κ values among the Namibian kimberlite suite. On a $^{207}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ diagram (Fig. 6) initial Pb isotope ratios of all the kimberlites have high $^{207}Pb/^{204}Pb$ relative to the Northern Hemisphere Reference Line (NHRL, Hart, 1984) and define an array with a slope sub-parallel to some South Atlantic ocean islands (e.g. Bouvet). The initial Pb data of the kimberlites are comparable with those for present-day South Atlantic islands (Fig. 6). The initial Pb isotope data of K2 overlap in part with the present-day field defined by Gough on $^{207}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ diagram but not on a $^{208}Pb/^{204}Pb$ vs $^{206}Pb/^{204}Pb$ diagram.

Clinopyroxene megacrysts have a restricted range in initial Sr and Nd isotope ratios (ϵ_{Nd} 2.8–4.7; ϵ_{Sr} –10 to –18). ϵ_{Nd} values are generally higher and ϵ_{Sr} lower than for the host kimberlites (Fig. 5). On an Sr–Nd isotope co-variation diagram (Fig. 5) the megacrysts plot close to, or slightly below, the ‘mantle array’. Figure 5 demonstrates that only the clinopyroxene megacryst from Deutsche Erde plots within the field of Namibian kimberlites. The clinopyroxene megacrysts define three distinct groups; (1) Mukorob; (2) Hanaus and Koherab; (3) Deutsche Erde. Compared with the total variation of the suite, individual groups have relatively constant Rb and Sr concentrations and Rb/Sr ratios. There are, however, significant inter-group differences, strengthening the argument for geographic variation (e.g. Mukorob Sr 80.3–88.5 ppm, Rb/Sr 0.0009–0.00236; Koherab Sr 209–215 ppm, Rb/Sr 0.00703–0.00716). Previously published Sr–Nd isotope data from Namibian megacrysts (R. Jones, 1987) are not plotted in Fig. 4 because Rb/Sr ratios were not determined and hence no age corrections can be performed. Measured Sr and Nd isotope ratios are comparable with those of the present study.

Sm–Nd isotope data for garnet megacrysts are presented in Table 3. The large fractionation in Sm/Nd between garnet and clinopyroxene is favourable for obtaining age information. Assessing the degree of garnet–clinopyroxene isotopic equilibrium at the time of kimberlite eruption provides an additional test of whether the megacrysts and kimberlites are cogenetic. The inferred high equilibration temperature of the minerals implies that they would be above the blocking temperature of the

Table 2: REE data for Namibian megacrysts

	cpx3	cpx4	cpx6	gnt1	gnt2	gnt3	gnt4	gnt5	gnt6
Kimberlite:	K6	K17	K6	K11	K6	K15	K16	K3	K13
La	1.561	2.489	1.980		0.384	0.013	0.039	0.054	0.06
Ce	5.170	9.207	8.023		1.047	0.049	0.221	0.353	0.366
Nd	4.375	9.076	5.508	3.774	1.978	0.161	1.113	1.476	1.413
Sm	1.106	2.321	1.401	3.541	1.737	0.155	1.003	1.389	1.285
Eu	0.357	0.743	0.477	1.990	0.975	0.084	0.536	0.769	0.638
Gd	0.823	1.978	1.145	8.666	4.391	0.391	2.337	3.338	3.726
Dy	0.616	1.124	0.801	17.298	8.572	0.704	4.473	7.061	5.67
Er	0.204	0.337	0.275	11.896	5.883	0.492	2.900	4.535	3.821
Yb	0.121	0.165	0.154	10.559	5.587	0.472	2.692	4.411	3.696

REE determined by isotope dilution following the technique of Thirlwall (1982). Reproducibility typically better than 1% for all elements and Sm/Nd ratios <0.2%. La and Ce of gnt1 are not reported, as underspiking caused excessive error magnification. Sample sizes were typically >100 mg so no blank correction need be applied. (See Table 3 and Fig. 2 for details of host kimberlite locations.)

isotope systems until the time of eruption. Clinopyroxenes from each pipe have small variations in Sm–Nd isotope systematics, which will introduce an error of the order of ± 3 Ma in any age determination and immediately suggests that the megacrysts do not simply represent crystallization products of a single magmatic event. Only one garnet sample was analysed from each locality so it is impossible to assess the extent of Nd isotope heterogeneity of garnets. Garnet–clinopyroxene pairs from the Mukorob II pipe yield Sm–Nd ages between 72 and 78 ± 8 Ma. Mineral pairs from Koherab and Hanaus yield ages between 55 and 91 Ma. Consequently, some ages are consistent with the ~ 71.5 Ma kimberlite eruption. The large range in Sm–Nd garnet–clinopyroxene ages, however, provides further evidence that the two minerals do not have a cogenetic origin with the kimberlite.

The clinopyroxene megacrysts have large variations in Pb isotope ratios (e.g. $^{206}\text{Pb}/^{204}\text{Pb}$, 18–19.5, Fig. 5, Table 3) and are characterized by relatively elevated $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ such that they plot above the NHRL of Hart (1984). These data define approximate linear arrays with a relatively restricted range in $^{207}\text{Pb}/^{204}\text{Pb}$ but large variations in $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$. These arrays overlap in part with basalts from South Atlantic ocean islands and continental flood basalts from Namibia and Brazil (Hawkesworth *et al.*, 1984, 1986, 1988). Measured U/Pb ratios of the clinopyroxene megacrysts are low compared with values for the host kimberlite ($\mu = 4\text{--}22$). Consequently, comparison with the kimberlite data can only be made after age correction. Coupled variations in initial $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ of the kimberlites and megacrysts are distinct (Fig. 6). We are unable to compare ($^{208}\text{Pb}/^{204}\text{Pb}$)_i ratios as Th

was not determined on the megacrysts. It should be noted, however, that measured $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of the megacrysts are generally lower than the initial ratios of the kimberlites, demonstrating that initial ratios will be distinct. The clinopyroxene sample from Deutsche Erde, which has an Sr–Nd isotope composition indistinguishable from that of the kimberlites, has a Pb isotope composition most distinct from that of the kimberlites. Consequently, the combined Sr–Nd–Pb isotope data appear to demonstrate that the kimberlites and megacrysts do not represent a single cogenetic suite.

MEGACRYST PETROGENESIS

REE evidence

If partition coefficients from the literature are assumed, the REE abundances of the garnet and clinopyroxene megacrysts can be used to estimate the composition of a parental equilibrium liquid. There are few high-pressure REE–garnet or REE–cpx partition coefficients for kimberlitic compositions [see discussion by Wood *et al.* (1999)] such that it is possible to estimate parental compositions only by assuming that partition coefficients are comparable with those of basaltic systems. Estimated compositions have Yb concentrations comparable with that of the host kimberlite. The degree of LREE enrichment of the calculated parental liquid is, however, significantly lower than for the host kimberlites (La/Yb_n ~ 20 compared with 90–110 in host kimberlites; Fig. 4c). These data argue against a genetic relationship between the megacrysts and the kimberlites. Harte (1983) reached a similar conclusion. Kramers *et al.* (1981), however, argued

Table 3: Sr–Nd–Pb isotope data for Namibian kimberlites and megacrysts

Sample	Rb (ppm)	Sr (ppm)	Rb/Sr	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr _i	ϵ_{87}	Sm	Nd	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd _i	U	Pb	Th	μ	K	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb _i		
Kimberlites																					
K2/W1/M	146	1560	0.0936	0.2707	0.70469 ± 1	0.70441	-2.87	14.00	109.8	0.07475	0.512775 ± 18	0.512740	5.81	8.54	27.7	50.56	242.82	19.137	15.631	39.094	
K2/W1/X	160	1629	0.0982	0.2841	0.70471 ± 1	0.70443	-2.68	13.80	109.4	0.07395	0.512764 ± 12	0.512729	5.81	8.65	28.0	45.80	280.07	19.125	15.654	39.153	
K2/W2/M	103	2819	0.0365	0.1057	0.70451 ± 2	0.70440	-3.07	14.23	113.2	0.07385	0.512712 ± 12	0.512677	5.09	8.29	30.9	45.80	280.07	19.226	15.645	39.267	
K2/W2/X	126	2050	0.0615	0.1778	0.70451 ± 1	0.70433	-4.08	14.23	110	0.07584	0.512749 ± 12	0.512714	5.09	8.02	30.9	45.80	280.07	19.184	15.636	39.143	
K2/W3/M	132	2504	0.0527	0.1525	0.70446 ± 1	0.70431	-4.38	13.87	107.3	0.07578	0.512699 ± 10	0.512664	5.81	6.91	28.7	67.81	280.04	19.372	15.66	39.432	
K2/W3/X	133	2480	0.0536	0.1551	0.70446 ± 1	0.70430	-4.46	13.91	106.8	0.07635	0.512740 ± 17	0.512704	7.00	7.91	28.7	67.81	280.04	19.336	15.646	39.357	
K2/3/M	140	1780	0.0787	0.2275	0.70525 ± 1	0.70502	5.70	13.82	105.8	0.07657	0.512713 ± 14	0.512677	7.00	7.72	28.7	67.81	280.04	19.323	15.651	39.354	
K2/3/X	137	1915	0.0715	0.2069	0.70553 ± 1	0.70532	10.02	13.16	102.7	0.07512	0.512753 ± 15	0.512718	7.00	7.72	28.7	67.81	280.04	19.323	15.651	39.354	
K35/W/X	70	1115	0.0628	0.1816	0.70380 ± 1	0.70362	-14.13	14.66	113.9	0.07545	0.512666 ± 16	0.512630	5.44	5.20	23.2	78.92	338.99	19.699	15.671	39.611	
K35/W/M	71	1099	0.0646	0.1868	0.70380 ± 1	0.70361	-14.23	14.33	116.7	0.07198	0.512703 ± 16	0.512669	5.44	5.13	23.2	78.92	338.99	19.725	15.675	39.624	
K39/W/M	32.4	1264	0.0256	0.0741	0.70399 ± 1	0.70382	-9.88	15.93	124	0.07531	0.512713 ± 16	0.512678	5.74	6.32	26.2	68.79	316.26	19.767	15.689	39.832	
K39/W/X	40.3	1400	0.0288	0.0833	0.70398 ± 1	0.70390	-10.18	15.53	123	0.07402	0.512666 ± 18	0.512630	5.74	6.32	26.2	68.79	316.26	19.767	15.686	39.704	
K39/1/M	61	1310	0.0466	0.1347	0.70403 ± 1	0.70390	-10.17	15.75	121.6	0.07583	0.512677 ± 18	0.512641	6.63	5.59	25.5	89.56	346.96	19.703	15.672	39.683	
K39/1/X	65	1385	0.0476	0.1377	0.70401 ± 1	0.70387	-10.59	16.02	127.8	0.07348	0.512705 ± 13	0.512671	6.19	5.89	25.5	89.56	346.96	19.817	15.69	39.831	
K53/W/M	71	1089	0.0652	0.1886	0.70384 ± 1	0.70365	-13.72	14.64	116.5	0.07367	0.512684 ± 15	0.512650	5.62	5.37	23.8	79.11	337.45	19.778	15.665	39.697	
K53/W/X	71	1110	0.0640	0.1850	0.70383 ± 2	0.70364	-13.84	13.99	110.7	0.07409	0.512728 ± 15	0.512693	5.62	5.24	23.8	79.11	337.45	19.815	15.691	39.772	
Megacrysts																					
<i>Koherab (K15)</i>																					
CPX1	0.147	209	0.0007	0.0020	0.703917 ± 12	0.70391	-9.92	1.614	6.322	0.14967	0.512838 ± 12	0.512768	0.097	0.453	15.41	15.41	17.964	15.600	37.909	17.964	
CPXE	0.154	215	0.0007	0.0021	0.703882 ± 12	0.70388	-10.42	1.612	6.315	0.14965	0.512853 ± 11	0.512783	0.005	0.195	15.41	1.76	18.411	15.637	38.301	18.411	
CPXF	0.152	213	0.0007	0.0021	0.703896 ± 12	0.70389	-10.22	1.596	6.257	0.14953	0.512827 ± 10	0.512757	0.004	0.187	15.41	1.52	18.534	15.642	38.421	18.534	
GNT3								0.155	0.161	0.56442	0.513076 ± 30	0.512811	5.17								
<i>Hansaus I (K11)</i>																					
CPX1	0.187	257	0.0007	0.0021	0.703889 ± 11	0.70389	-10.32	1.367	5.36	0.14951	0.512857 ± 10	0.512787	0.005	0.653	18.074	0.58	18.074	15.594	37.898	18.074	
GNT1								3.541	3.774	0.55007	0.513001 ± 17	0.512744	3.86								
<i>Diamontkop (K13)</i>																					
GNT6								1.285	1.413	0.53316	0.513025 ± 18	0.512776	4.48								
<i>Deutsche Erde I (K16)</i>																					
GNT4								1.003	1.113	0.52831	0.512829 ± 16	0.512682	2.65								
<i>Deutsche Erde II (K17)</i>																					
CPX4	0.307	171	0.0018	0.0052	0.703817 ± 12	0.70381	-11.39	2.321	9.076	0.14992	0.512759 ± 10	0.512689	0.097	0.301	23.06	23.06	17.886	15.597	37.835	17.886	
<i>Mukarob II (K6)</i>																					
CPX1	0.157	82.4	0.0019	0.0055	0.703856 ± 13	0.70385	-17.93	1.183	4.652	0.14908	0.512799 ± 11	0.512729	0.012	0.461	1.90	1.90	18.147	15.625	38.146	18.147	
CPX3	0.191	88.5	0.0022	0.0062	0.703335 ± 12	0.70333	-18.24	1.106	4.375	0.14820	0.512813 ± 10	0.512744	0.069	0.44	11.51	11.51	19.024	15.669	38.75	19.024	
CPX5	0.075	80.3	0.0009	0.0027	0.703373 ± 11	0.70337	-17.65	1.246	4.919	0.14850	0.512806 ± 12	0.512737	0.019	0.187	7.26	7.26	18.680	15.648	38.515	18.680	
CPX6	0.193	87.9	0.0022	0.0063	0.703369 ± 11	0.70336	-17.76	1.401	5.508	0.14911	0.512809 ± 10	0.512739	0.098	0.275	25.87	25.87	18.170	15.630	38.106	18.170	
CPX7	0.201	85.3	0.0024	0.0068	0.703401 ± 10	0.70339	-17.31	1.18	4.646	0.14889	0.512812 ± 12	0.512742	0.008	0.315	1.96	1.96	18.667	15.651	38.568	18.667	
GNT2								1.737	1.978	0.51483	0.512886 ± 15	0.512744	3.87								
<i>Vipersdorf (K3)</i>																					
GNT5								1.389	1.476	0.55171	0.513021 ± 17	0.512763	4.23								

U, Pb, Rb, Sr, Sm and Nd determined by isotope dilution for the megacrysts. Repetition of standard NBS 607 established that Rb is determined to better than 1% and Sr <0.1%. Rb/Sr ratios determined by XRF on the host kimberlites are also better than 1%. REE determined by ICP at Leeds University (Spriggs, 1988). Reproducibility of repeat samples was better than 1% for LREE and better than 5% for HREE where dissolution of minor phases may have been a contributing factor. REE abundances for selected samples were repeated by INAA at the Open University following procedures outlined by Rogers *et al.* (1985). Th was determined by INAA. Kimberlite locations shown in Fig. 2.

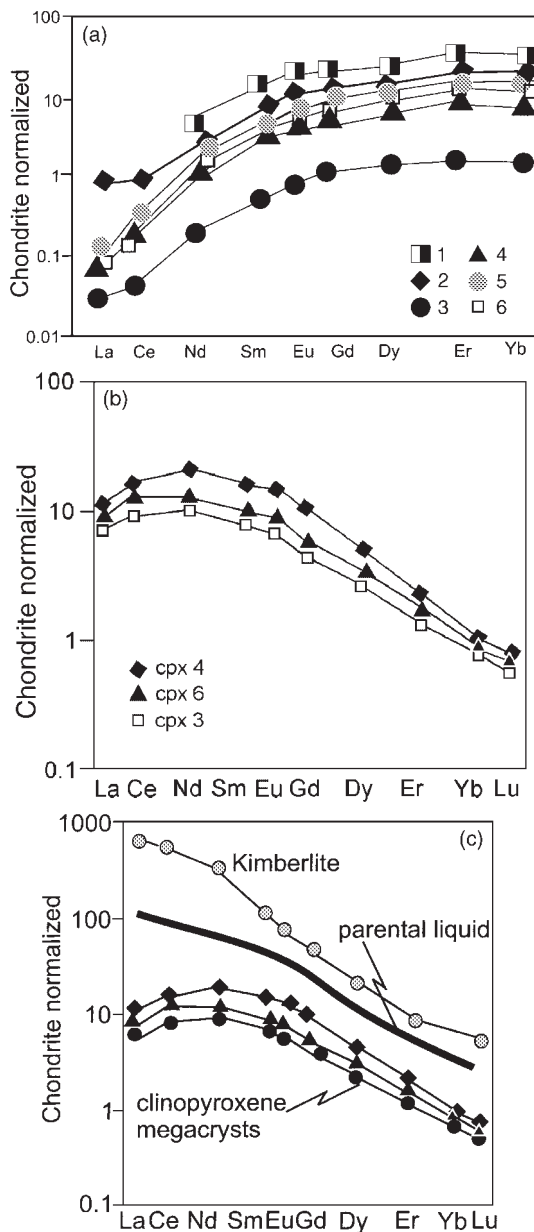


Fig. 4. (a) Chondrite-normalized REE diagram of garnet megacrysts. (See Table 2 for sample localities.) (b) Chondrite-normalized REE diagram of clinopyroxene megacrysts. (See Table 2 for sample localities.) (c) Chondrite-normalized REE diagram depicting difference between the calculated equilibrium liquid for the clinopyroxene megacrysts and the host kimberlite. The marked difference between calculated parental liquid and host kimberlite should be noted.

that because megacryst assemblages are cogenetic with their host kimberlites, clinopyroxene–kimberlite REE partition coefficients were up to an order of magnitude lower than in basaltic systems. To date, no experimental data have been presented to support this supposition.

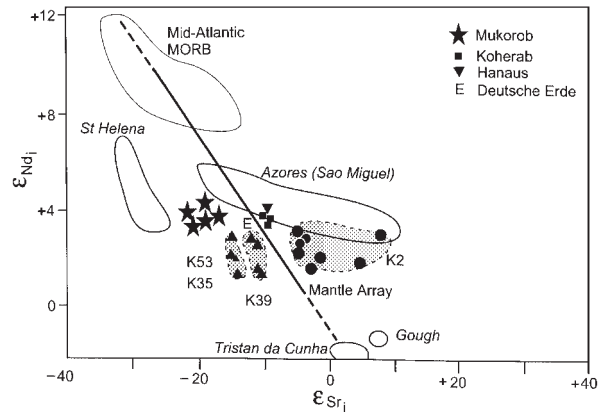


Fig. 5. Initial Sr–Nd isotope diagram for Namibian kimberlites and clinopyroxene megacrysts. Kimberlites indicated by small dots (K2) or triangles surrounded by shaded field. Megacrysts have individual symbols depending on locality. Fields of Mid-Atlantic MORB and representative Atlantic OIB taken from Zindler & Hart (1986), Chaffey *et al.* (1989) and Davies *et al.* (1989). Continuous line used for reference represents the 'mantle array' that joins MORB to Bulk Earth.

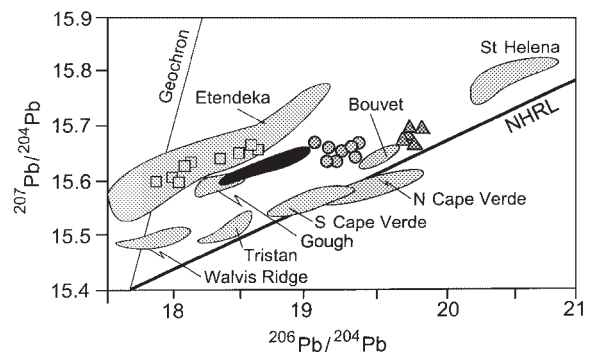


Fig. 6. Present-day and initial $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ for Namibian kimberlites and clinopyroxene megacrysts. Fields of representative Atlantic OIB taken from Zindler & Hart (1986) and Davies *et al.* (1989). Etendeka flood basalts from Hawkesworth *et al.* (1986, 1988). Grey squares, clinopyroxene megacrysts initial ratios; grey triangles, kimberlites with 'typical' Group I kimberlite characteristics (35, 39, 53); grey circles, kimberlite K2, which is characterized by chemistry transitional to Group II kimberlite. Initial ratios of the kimberlites plotted as a black field. NHRL, Northern Hemisphere Reference Line.

Combined Sr–Pb–Nd isotope evidence

The megacrysts and host kimberlites are in Sr–Nd–Pb isotope disequilibrium (Figs 5 and 6). If this relationship is inherited from the mantle then it palpably rules out a cogenetic relationship and hence negates the argument proposed by Kramers *et al.* (1981) for lower melt–clinopyroxene partition coefficients in kimberlitic melts. Before accepting a non-genetic relationship a full assessment of crustal contamination is required. The relatively constant REE slopes of the garnet and clinopyroxene megacrysts imply that any isotope variations are not a consequence of variable contamination by the host

kimberlites or fluids derived from them associated with low-temperature alteration. Both processes would tend to introduce an LREE enriched signature to the megacrysts. Namibian kimberlites have high trace element contents (e.g. Sr, Nd and Pb contents of 1100, 100 and 5–8 ppm, respectively). Mass balance calculations demonstrate that 10% crust must be assimilated by the kimberlites to change their $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from that of the megacrysts (0.7033) to an initial ratio of 0.7039 (by assuming an $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.73 for the Proterozoic basement). Given the low SiO_2 and high MgO, Cr and Ni contents of the kimberlites this clearly has not occurred. The isotope distinction between megacrysts and kimberlites therefore implies derivation from different sources.

Previous workers have noted apparent isotope disequilibrium between megacrysts and host kimberlite (e.g. Harte, 1983). This difference has been variously ascribed to minor alteration or the early precipitation of the megacryst suite from a 'proto-kimberlitic' melt followed by compositional change in the melt as a result of fractional crystallization and assimilation (e.g. Kramers *et al.*, 1981; R. Jones, 1987). These models are appealing in that they avoid the need for a complex petrogenetic history. Such models are, however, very difficult to reconcile with aspects of megacryst and kimberlite chemical and isotopic compositions. For example, the kimberlites record little evidence of the extraction of the megacryst suite. Fractionation of olivine and spinel would rapidly deplete the magma in MgO, Ni and Cr. Clinopyroxene fractionation would deplete Sc and garnet fractionation Sc, Ni and HREE. The major element variations recorded by megacryst suites (e.g. Mitchell, 1987) do, however, imply that they formed from a magma that underwent significant fractionation.

The topology of the Sr–Nd–Pb isotope diagrams helps to constrain possible source components in megacryst and kimberlite genesis. Sr/Nd ratios of most mantle-derived magmas are within the range 10–20 (excluding subduction-related volcanism) such that mixing lines between components on an Sr–Nd isotope diagram are close to straight lines. In terms of their Sr–Nd isotopes the kimberlites and megacrysts could be explained through some form of mixing process that involves a relatively depleted source [mid-ocean ridge basalt (MORB) or ocean-island basalt (OIB) component] and an enriched component (OIB). The Mukorob megacryst samples have relatively unradiogenic Sr isotope ratios such that they have characteristics transitional toward HIMU and EM1 ocean islands (e.g. St Helena and Cape Verde, Chaffey *et al.*, 1989; Davies *et al.*, 1989). On Pb/Pb diagrams the megacryst suite forms arrays sub-parallel to those of some Atlantic ocean islands and has relatively high $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios that are more extreme than those of Atlantic ocean islands with the DUPAL isotope signature (e.g. Gough, Walvis Ridge and

Cape Verde; Fig. 6). In contrast, HIMU islands generally have radiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ and relatively low $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$. Consequently, the Sr–Nd–Pb isotope systematics of the megacryst suite is most comparable with that of ocean islands with a DUPAL signature.

In the following discussion petrogenetic models are proposed that assess why the megacryst suite has a DUPAL Pb isotopic signature. It is acknowledged, however, that the combined Sr–Nd–Pb isotopic composition of the megacryst suite is not directly comparable with DUPAL in that Sr isotope ratios are generally less radiogenic. Any proposed model for megacryst kimberlite formation must explain why the megacryst suite has a markedly greater DUPAL Pb isotope signature than the host kimberlite. In this respect, the source of the DUPAL anomaly is crucial. It is well known that the DUPAL isotopic anomaly is not restricted to the oceanic environment, being present in Parana and Karoo–Etendeka basalts at both margins of the Atlantic (e.g. Hawkesworth *et al.*, 1984, 1988). The origin of the DUPAL chemical and isotopic anomaly has been the subject of considerable debate. Le Roex (1986) proposed that the regional DUPAL anomaly in the Southern Hemisphere was a consequence of asthenospheric derived hotspot magmatism, which also produced Group I and II kimberlites in southern Africa. Weaver *et al.* (1986) are some of many who have proposed that recycling of sediments into the source of plumes was responsible, whereas Richardson *et al.* (1984), Hawkesworth *et al.* (1986) and Davies *et al.* (1989) argued for an SCLM origin. It is, however, difficult to distinguish between the chemical characteristics of ancient recycled and modified sediments and the SCLM. The strongest evidence for a shallow lithospheric origin of the DUPAL anomaly comes from age information inferred from Sr–Nd–Pb isotopes, which suggests that the sources of continental flood basalts, SCLM xenoliths and the crust are of equal age (Hawkesworth *et al.*, 1984; Pearson, 1999). This observation implies that the entire lithosphere stabilized together. We therefore favour a lithospheric origin for the DUPAL signal. With this assumption a model is proposed to explain the kimberlite–megacryst association (see Fig. 8, below).

Knowledge of the thermal and tectonic processes that have modified the lithosphere in the region and are responsible for the initiation of kimberlite magmatism may be a key to understanding the genetic relationship between the megacryst suite and host rock. Plate reconstructions suggest that several currently active South Atlantic hotspots were initiated and passed beneath southern Africa in the last 200 my (Fig. 7; Duncan, 1981; Morgan, 1983). The latest reconstruction (O'Connor *et al.*, 1999) suggests slower recent motion than previously predicted and that the Discovery plume track was active under the Gibeon province between 75 and 80 Ma. Previous magmatism associated with opening of the South

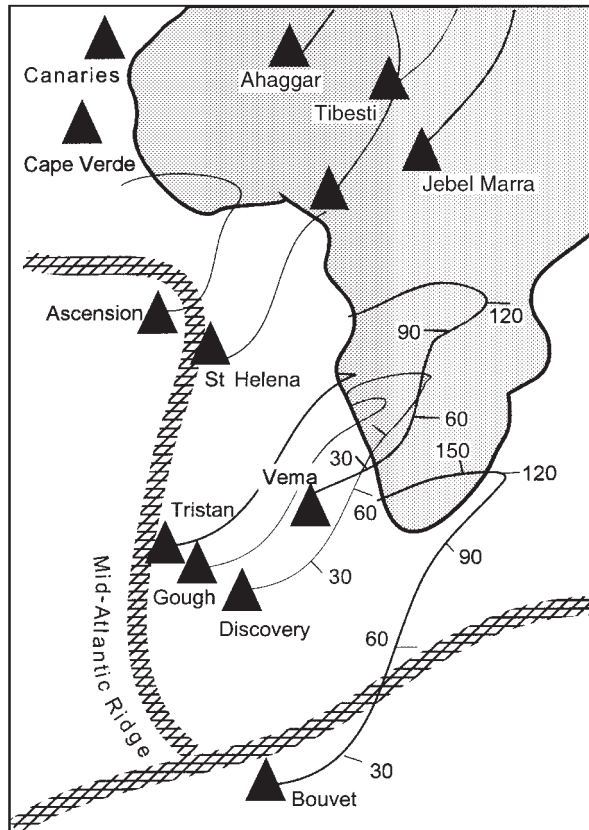


Fig. 7. Schematic figure showing the approximate track of hotspots that have passed beneath southern Africa in the last 200 my, from Hartnady & LeRoex (1983). Before the passage of the Discovery plume beneath Namibia the lithosphere had been modified by magmatic activity associated with the Bouvet plume at ~175 Ma and initial formation of the South Atlantic at ~130 Ma. The detailed passage of the Vema and Discovery hotspots can be seen in Fig. 1.

Atlantic and the initiation of Tristan and Gough (Etendeka and Parana flood volcanism, Griffiths & Campbell, 1990) will also have extensively modified the SCLM beneath Namibia. Kimberlite volcanism in Namibia occurs ~5–10 my after the passage of the Discovery plume beneath the Gibeon region. This time delay and the different Sr–Nd–Pb isotope signature of the current plume-related ocean-island basalts and the Namibian kimberlites suggests that the plume did not act as the direct source of the kimberlites.

The relatively short time delay between the inferred passage of the Discovery plume and kimberlite magmatism implies that the plume is probably ultimately responsible for magma production. The isotopic distinction between the plume and the volcanism, however, establishes that the plume did not provide significant mass to the magmatism. One explanation is that conduction of heat from the hotspot into the thermo-chemical boundary layer at the base of the SCLM resulted in the subsequent

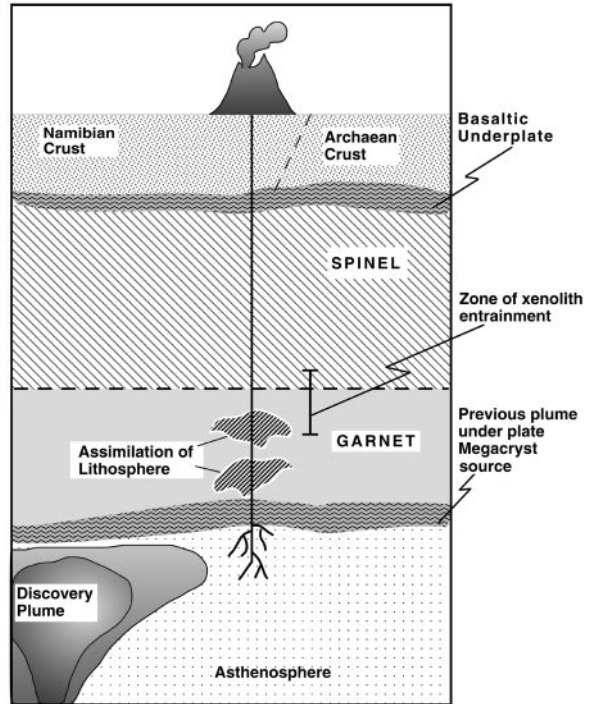


Fig. 8. Schematic model to explain the Gibeon kimberlite–megacryst association. Previous passage of the Discovery plume causes topography at the base of the lithosphere, possibly associated with the tectonic boundary between Namaqua and Namibian aged lithosphere. Fluids from the nearby plume migrate along the base of the SCLM and are concentrated into the recently thermally perturbed asthenosphere causing kimberlite genesis. Kimberlite magmas are variably modified as they pass through the base of the SCLM, assimilating melts derived from the most easily fused parts of the SCLM before rapid emplacement from depth. During emplacement kimberlites incorporate xenoliths from various depths in the lithosphere. The megacryst suite was formed in previous periods of plume-related underplating, most probably at the base of the lithosphere. Residence at relatively high temperatures for >10 my is required to explain the homogeneous nature of individual megacrysts.

melting of a shallower non-plume-related source. In this model volatile-rich regions of the SCLM with the lowest solidus temperature would be the source of the kimberlite (Wilson *et al.*, 1995). This argument is not favoured because the isotopic signature of the kimberlites is generally not compatible with an SCLM origin. The exception is kimberlite K2, which has a composition that is transitional to that of DUPAL, implying some interaction with old trace element enriched lithosphere. Given that xenoliths are transported from above the zone of melt generation, kimberlite formation is inferred to be deeper and hence most probably immediately below the base of the lithosphere (e.g. Le Roex, 1986).

Passage of the Discovery plume will have caused topography at the base of the lithosphere, possibly associated with the tectonic boundary between Namaqua and Namibian lithosphere. A model is proposed wherein fluid-rich melts emanating from the plume, which at the time

of volcanism was within ~ 100 km, migrate along the base of the SCLM and are concentrated into the recently thermally perturbed asthenosphere, causing small degrees of melting and kimberlite genesis (Fig. 8). Kimberlite magmas are variably modified as they pass through the base of the SCLM, assimilating melts derived from the most easily fused parts of the SCLM before rapid emplacement from depth (Fig. 8). During emplacement the kimberlites sample xenoliths from various depths in the lithosphere. A test of this hypothesis of kimberlite genesis would be to examine the rare gas isotope ratios of the kimberlite; high $^3\text{He}/^4\text{He}$ and $^{21}\text{Ne}/^{22}\text{Ne}$ associated with the plume may be transferred to the magmatism even though it is argued that little mass has been. In addition, kimberlites such as K2 that have characteristics transitional to Group II kimberlites, as a result of inferred assimilation of SCLM, would have a more uranogenic He and Ne isotope signature than Group I kimberlites. Unfortunately, widespread alteration of kimberlites and high U contents mean that only pristine melt-inclusion-free olivines can be used for such a study and such material is difficult to obtain.

The megacryst suite is incorporated into the kimberlite as it migrates through the SCLM. The extreme DUPAL Pb isotope signature of the megacryst suite implies a significantly greater lithospheric contribution than for the kimberlites. The relatively depleted Nd and Sr isotope signature, however, demonstrates that the megacrysts did not simply crystallize from melts of old SCLM. The low Pb contents of the megacrysts mean that their Pb isotopes are sensitive to interaction with the SCLM, and high Sr/Pb ratios mean that Sr isotope ratios will not record an SCLM signature unless $>10\%$ SCLM material is assimilated. Slopes defined by the megacrysts on a $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 6) are shallower than that of the Etendeka and Parana flood basalts. This observation implies that, in addition to the SCLM, a source with isotopic systematics comparable with South Atlantic Ocean islands such as Bouvet and Ascension was involved in megacryst petrogenesis. On a $^{208}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (not shown) the megacrysts define a slope that is steeper than that for the flood basalts but shallower than the NHRL and for most OIB. These relationships imply that the megacrysts formed from asthenospheric-derived melts that have undergone extensive crystallization within and contamination by the SCLM (Fig. 8). We therefore favour megacryst formation within and immediately above the thermo-chemical boundary layer at the base of the SCLM that, beneath Namibia, has been extensively modified by the passage of several plumes (Fig. 7). An implication of this model is that the megacryst suite would have He and Ne isotope ratios with a relatively uranogenic signature compared with ambient upper mantle as a result of interaction with the relatively U-enriched SCLM.

The Pb–Sr–Nd relationships outlined above give little direct age information but provide useful constraints on the timing of megacryst formation. The megacrysts have relatively high Sm/Nd and low μ and Rb/Sr so that with time the suite records ingrowth in radiogenic ^{143}Nd but little change in Sr and Pb isotope ratios. ^{143}Nd in garnets changes, relative to Bulk Earth, at a rate of $\sim 4.5 \epsilon_{\text{Nd}}$ units per 100 my. In addition, radiogenic Pb growth in the megacrysts has been insufficient to disrupt the shallow Pb/Pb array on a $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 6) despite variable μ values. The Nd isotope ratios of the megacrysts are similar to those of many melts recently derived from the asthenosphere. In combination these data are consistent with the megacryst suite having a short but significant residence in the base of the SCLM, >10 and <100 my. Major and trace element diffusion during this mantle residence provides the explanation of why individual megacrysts have homogeneous compositions.

The favoured model presented above makes the assumption that the megacrysts have undergone greater interaction with the SCLM than the host kimberlites. Griffin *et al.* (2000) also favour an SCLM origin for megacrystic zircons based on a Hf isotope study. Mass balance calculations based on Hf isotope studies of continental crust and depleted mantle suggest that a significant reservoir exists in the Earth with low Nd/Hf and non-radiogenic Hf ratios (Blichert-Toft & Albarède, 1997). Kimberlites and lamproites have such non-radiogenic Hf and low Nd/Hf ratios (Nowell *et al.*, 1998a, 1998b). These data may potentially be used to distinguish if kimberlites are derived from great depth or from the SCLM (Nowell & Pearson, 1998). If the SCLM is characterized by non-radiogenic Hf and low Nd/Hf ratios then this would represent the source of kimberlites. If, however, the SCLM reservoir lies on the Nd–Hf isotope correlation then a deep mantle source would be favoured in kimberlite genesis. Coupled Hf and He isotope studies are under way to substantiate which of the two models is correct.

CONCLUSIONS

Significant trace element variations within low-Cr megacryst suites from individual kimberlite localities suggest their derivation from a magma, or series of magmas that underwent extensive differentiation. Host kimberlite magmas record no evidence of extensive fractional crystallization of the megacryst suite. Calculated equilibrium melts that may have crystallized the low-Cr megacryst suite have LREE enrichment comparable with that of alkalic basalts not kimberlite. Sr–Nd–Pb isotope disequilibrium between the megacryst and host kimberlites palpably establishes that the megacryst suite is not cognate

in origin. The generally similar isotopic character of the two rock types, however, implies derivation from a similar source region that, from P - T estimates, must be close to the base of the lithosphere. We favour a model in which megacrysts formed near the base of the SCLM associated with basaltic volcanism. Kimberlite magmatism occurs as a result of the localization of fluid-rich melts derived from the Discovery plume and subsequently entrains the megacrysts. Currently, however, derivation of the DUPAL isotope signature from depths significantly below the SCLM cannot be ruled out. Homogeneous major and trace element compositions and isotope systematics suggest that the megacryst suite had extended residence times at the base of the SCLM, of >10 and <100 my.

ACKNOWLEDGEMENTS

Without the drill core samples and field support provided by DeBeers this research project would not have been possible. This paper has benefited from discussion with several colleagues, notably Graham Pearson, Craig Smith and Klaus Mezger. Constructive reviews by Dorrit Jacob and an anonymous reviewer are gratefully acknowledged. A.J.S. was funded by an NERC studentship.

REFERENCES

- Basu, A. R. & Tatsumoto, M. (1980). Nd isotopes in selected mantle derived rocks and minerals and their implications for mantle evolution. *Contributions to Mineralogy and Petrology* **75**, 43–54.
- Blichert-Toft, J. & Albarède, F. (1997). The Lu–Hf geochemistry of chondrites and the evolution of the mantle–crust system. *Earth and Planetary Science Letters* **148**, 243–258.
- Chaffey, D. J., Cliff, R. A. & Wilson, B. M. (1989). Characterization of the St Helena magma source. In: Saunders, A. D. & Norry, M. J. (eds) *Magmatism in the Ocean Basins*. Geological Society, London, *Special Publications* **42**, 257–276.
- Clement, C. R. (1982). A comparative geological study of some major kimberlite pipes in the Northern Cape and Orange Free State. Ph.D. thesis, University of Cape Town.
- Davies, G. R., Norry, M. J., Gerlach, D. C. & Cliff, R. A. (1989). A combined chemical and Pb–Sr–Nd isotope study of the Azores and Cape Verde hot-spots: the geodynamic implications. In: Saunders, A. D. & Norry, M. J. (eds) *Magmatism in the Ocean Basins*. Geological Society, London, *Special Publications* **42**, 231–255.
- Duncan, R. A. (1981). Hotspots in the Southern Oceans—an absolute frame for the motion of the Gondwana continents. *Tectonophysics* **74**, 29–42.
- Durrheim, R. J. & Green, R. W. E. (1992). A seismic refraction investigation of the Archaean Kaapvaal Craton, South Africa, using mine tremors as the energy source. *Geophysical Journal International* **108**, 812–832.
- Fairhead, J. D. & Reeves, C. V. (1976). Teleseismic delay times, Bouguer anomalies and inferred thickness of the African lithosphere. *Earth and Planetary Science Letters* **36**, 63–76.
- Fesq, H. W., Kable, E. J. D. & Gurney, J. J. (1975). Aspects of the geochemistry of kimberlites from Premier mine and other S. African occurrences with particular reference to the REE. *Physics and Chemistry of the Earth* **9**, 686–707.
- Franz, L., Brey, G. P. & Okrusch, M. (1996). Re-equilibration of ultramafic xenoliths from Namibia by metasomatic processes at the mantle boundary. *Contributions to Mineralogy and Petrology* **126**, 181–198.
- Fraser, K. J., Hawkesworth, C. J., Erlank, A. J., Mitchell, R. H. & Scott-Smith, B. H. (1985). Sr, Nd and Pb isotopes and trace elements of lamproites and kimberlites. *Earth and Planetary Science Letters* **76**, 57–70.
- Griffin, W. L., Pearson, N. J., Belousova, E., Jackson, S. E., van Achterbergh, E., O'Reilly, S. Y. & Shee, S. R. (2000). The Hf isotope composition of cratonic mantle: LAM–MC–ICPMS analysis of zircon megacrysts in kimberlites. *Geochimica et Cosmochimica Acta* **64**, 133–147.
- Griffiths, R. W. & Campbell, I. H. (1990). Stirring and structure in mantle plumes. *Earth and Planetary Science Letters* **99**, 66–78.
- Gurney, J. J., Jakob, W. R. O. & Dawson, J. B. (1979). Megacrysts from the Monastery kimberlite pipe, South Africa. In: Boyd, F. R. & Meyer, H. O. A. (eds) *The Mantle Sample*. Washington, DC: American Geophysical Union, pp. 227–243.
- Hart, S. R. (1984). A large scale isotopic anomaly in the southern hemisphere mantle. *Nature* **309**, 753–758.
- Harte, B. (1983). Mantle peridotites and processes—the kimberlite sample. In: Hawkesworth, C. J. & Norry, M. J. (eds) *Continental Basalts and Mantle Xenoliths*. Nantwich: Shiva, pp. 46–91.
- Harte, B. & Gurney, J. J. (1981). The mode of formation of Cr poor megacryst suites from kimberlites. *Journal of Geology* **89**, 749–753.
- Harte, B. & Kirkley, H. B. (1997). Partitioning of trace elements between clinopyroxene and garnet: data from mantle eclogites. *Chemical Geology* **136**, 1–24.
- Hartnady, C. J. M. & LeRoex, A. P. (1983). Southern Ocean hotspot tracks and the Cenozoic absolute motion of the African, Antarctic and South American plates. *Earth and Planetary Science Letters* **75**, 245–257.
- Hawkesworth, C. J., Marsh, J. S., Duncan, A. R., Erlank, A. J. & Norry, M. J. (1984). The role of continental lithosphere in the generation of Karoo volcanic rocks. *Geological Society of South Africa Special Publication* **13**, 341–354.
- Hawkesworth, C. J., Mantovani, M. M., Taylor, P. N. & Palacz, Z. (1986). Evidence from the Parana of Brazil for a continental contribution to Dupal basalts. *Nature* **322**, 356–359.
- Hawkesworth, C. J., Mantovani, M. M. & Peate, D. W. (1988). Lithosphere remobilisation during Parana magmatism. *Journal of Petrology (Lithosphere Special Issue)* 205–223.
- Hoal, B. G., Hoal, K. E. O., Boyd, F. R. & Pearson, D. G. (1995). Age constraints on the crustal and mantle lithosphere beneath the Gibeon kimberlite field, Namibia. *South African Journal of Geology* **98**, 112–118.
- Hops, J., Gurney, J. J. & Harte, B. (1992). The Jagersfontein Cr-poor megacryst suite—towards a model for megacryst petrogenesis. *Journal of Volcanology and Geothermal Research* **50**, 143–160.
- Jones, M. Q. W. (1981). Heatflow and heat production studies in the Namaqua mobile belt and Kaapvaal Craton. Ph.D. thesis, University of Johannesburg.
- Jones, M. Q. W. (1987). Heatflow and heat production in the Namaqua mobile belt, South Africa. *Journal of Geophysical Research* **92**, 6273–6289.
- Jones, R. A. (1987). Sr and Nd isotopes and REE evidence for the genesis of megacrysts in kimberlites of southern Africa. In: Nixon, P. H. (ed.) *Mantle Xenoliths*. Chichester: Wiley, pp. 711–724.
- Kramers, J. D. (1977). Pb and Sr isotopes in Cretaceous kimberlites and mantle derived xenoliths from Southern Africa. *Earth and Planetary Science Letters* **34**, 419–431.

- Kramers, J. D. (1979). Pb, U, Sr, K, and Rb in inclusion bearing diamonds and mantle derived xenoliths from R.S.A. *Earth and Planetary Science Letters* **42**, 58–70.
- Kramers, J. D., Smith, C. B., Lock, N., Harmon, R. S. & Boyd, F. R. (1981). Can kimberlites be generated from an ordinary mantle? A trace element point of view. *Nature* **241**, 53–56.
- Le Roex, A. P. (1986). Geochemical correlations between S. African kimberlites and S. African hotspots. *Nature* **324**, 243–245.
- Mitchell, R. H. (1984). Garnet lherzolites from Hanaus-1 and Louwrensia kimberlites of Namibia. *Contributions to Mineralogy and Petrology* **86**, 178–188.
- Mitchell, R. H. (1986). *Kimberlites; Mineralogy, Geochemistry and Petrology*. New York: Plenum.
- Mitchell, R. H. (1987). Megacrysts in kimberlites from the Gibeon field, Namibia. *Neues Jahrbuch für Mineralogie, Abhandlungen* **157**, 267–283.
- Morgan, W. J. (1983). Hotspot tracks and the early rifting of the Atlantic. *Tectonophysics* **94**, 123–139.
- Nixon, P. H. & Boyd, F. R. (1973). The discrete nodule association in kimberlites in northern Lesotho. In: Nixon, P. H. (ed.) *Lesotho Kimberlites*. Cape Town: Cape and Transvaal Printers, pp. 67–75.
- Nixon, P. H., von Knorring, O. & Rooke, J. M. (1963). Kimberlites and associated inclusions of Basutoland: a mineralogical and geochemical study. *American Mineralogist* **48**, 1090–1132.
- Nixon, P. H., Rogers, N. W., Gibson, I. L. & Grey, A. (1981). Depleted and fertile mantle xenoliths from southern African kimberlites. *Annual Review of Earth and Planetary Sciences* **9**, 285–309.
- Nowell, G. M. & Pearson, D. G. (1998). Hf isotope constraints on the genesis of kimberlitic megacrysts: evidence for a deep mantle component in kimberlites. 7th International Kimberlite Conference, Extended Abstracts, University of Cape Town, pp. 634–636.
- Nowell, G. M., Kempton, P. D. & Pearson, D. G. (1998a). Hf–Nd systematics of kimberlites: relevance to terrestrial Hf–Nd systematics. 7th International Kimberlite Conference, Extended Abstracts, University of Cape Town, pp. 628–630.
- Nowell, G. M., Pearson, D. G., Kempton, P. D., Irving, A. J. & Turner, S. (1998b). A Hf isotope study of lamproites: implications for their origin and relationship to kimberlites. 7th International Kimberlite Conference, Extended Abstracts, University of Cape Town, pp. 637–639.
- O'Connor, J. M., Stoffers, P., van den Bogaard, P. & McWilliams, M. (1999). First seamount evidence for a significantly slower African plate motion since 19 to 30 Ma. *Earth and Planetary Science Letters* **171**, 575–589.
- Pasteris, J. D., Boyd, F. R. & Nixon, P. H. (1979). The ilmenite association at the Frank Smith mine, R.S.A. In: Boyd, F. R. & Meyer, H. O. A. (eds) *The Mantle Sample*. Washington, DC: American Geophysical Union, pp. 256–278.
- Pearson, D. G. (1999). The age of continental roots. *Lithos* **48**, 171–194.
- Pearson, D. G., Davies, G. R. & Nixon, P. H. (1993). Geochemical constraints on the petrogenesis of diamond facies pyroxenites from the Beni Bousera peridotite massif, North Morocco. *Journal of Petrology* **34**, 125–172.
- Pearson, D. G., Boyd, F. R., Hoal, K. E. O., Hoal, B. G., Nixon, P. H. & Rogers, N. W. (1994). A Re–Os isotopic and petrological study of Namibian peridotites: contrasting petrogenesis and composition of on- and off-craton lithospheric mantle. Extended Abstract. *Mineralogical Magazine* **58**, 703–704.
- Richardson, S. H., Gurney, J. J., Erlank, A. J. & Harris, J. W. (1984). Origins of diamonds in old enriched mantle. *Nature* **310**, 198–202.
- Rogers, N. W., Hawkesworth, C. J., Parker, R. J. & Marsh, J. S. (1985). The geochemistry of potassic lavas from Vulcini, central Italy, and implications for mantle enrichment processes beneath the Roman region. *Contributions to Mineralogy and Petrology* **90**, 244–255.
- Schulze, D. J. (1987). Megacrysts from alkaline volcanic rocks. In: Nixon, P. H. (ed.) *Mantle Xenoliths*. Chichester: Wiley, pp. 433–452.
- Smith, C. B. (1983). Pb, Sr and Nd isotopic evidence for sources of southern African Cretaceous kimberlites. *Nature* **304**, 51–54.
- Smith, C. B., Gurney, J. J., Skinner, E. M. W., Clement, C. R. & Ebrahim, N. (1985). Geochemical character of S. African kimberlites: a new approach based on isotopic constraints. *Transactions of the Geological Society of South Africa* **88**, 267–280.
- Spriggs, A. J. (1988). An isotopic and geochemical study of kimberlites and associated alkaline rocks from Namibia. Ph.D. thesis, University of Leeds.
- Tankard, A. J., Jackson, M. P., Eriksson, A. K. A., Hobday, D. K., Hunter, D. R. & Minter, W. E. L. (1982). *Crustal Evolution of Southern Africa*. Berlin: Springer.
- Thirlwall, M. F. (1982). A triple filament method for rapid and precise analysis of rare earth elements by isotope dilution. *Chemical Geology* **35**, 155–166.
- Van Heerden, L. (1984). Lower crustal xenoliths from the Gibeon Kimberlite Province, S.W.A.: an investigation. B.Sc. thesis, University of Cape Town.
- Weaver, B. L., Wood, B. J., Tarney, J. & Joron, J. L. (1986). Role of subducted sediment in the genesis of ocean-island basalts: geochemical evidence from S. Atlantic ocean islands. *Geology* **14**, 275–278.
- Wilson, M., Rosenbaum, J. M. & Dunworth, E.-A. (1995). Melilitites; partial melts of the thermal boundary layer? *Contributions to Mineralogy and Petrology* **119**, 181–196.
- Wood, B. J., Blundy, J. A. & Robinson, A. C. (1999). The role of clinopyroxene in generating U-series disequilibrium during mantle melting. *Geochimica et Cosmochimica Acta* **63**, 1613–1620.
- Zindler, A. & Hart, S. R. (1986). Chemical geodynamics. *Annual Review of Earth and Planetary Sciences* **14**, 493–571.