PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

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Phil. Trans. R. Soc. Lond. A 1980 **297**, 273-293
doi: 10.1098/rsta.1980.0215

Phil. Trans. R. Soc. Lond. A **297**, 273–293 (1980)
Printed in Great Britain

Chemical variations in upper mantle nodules from southern African kimberlites

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A summary of the rock and mineral nodules erupted with kimberlites is presented. Garnet-peridotites are separated into various categories according to: their relatively depleted and fertile chemical character; their deformed and undeformed nature; and the Ca/(Ca + Mg) ratio of their clinopyroxenes (with associated temperature characteristics). Some bulk chemical similarities are noted between high-temperature, deformed and less depleted peridotites and the wall rocks to minor intrusions.

There is clear evidence of the occurrence of infiltration metasomatism (involving K, Ti and other incompatible elements) in some mantle xenoliths before their incorporation in the kimberlite. These metasomatic effects may be linked with earlier magmatic events. Extreme chemical heterogeneities may be produced by metasomatism, as with restricted partial melting. Attempts to find homogeneous rocks with average or pristine upper mantle compositions are considered unrealistic.

Within the Kaapvaal craton there is no overall pattern of lateral variations in the nodule types from kimberlite, though there is evidence of local heterogeneity and regional changes at craton margins. Boyd & Nixon's models of vertical layering of rocks and minerals in the upper mantle are contrasted with a model in which horizontal variations in temperatures occur after magma intrusion. It is suggested that the high-temperature deformed peridotites and the megacrysts result from the intrusion of high-temperature magmas into overlying cooler mantle, and that xenoliths from kimberlite provide little direct evidence of mantle stratification with depth.

1. INTRODUCTION

Kimberlites erupt through the outermost parts of the Earth, often carrying diamonds and other minerals and rocks which are evidently of deeper origin than inclusions in other eruptive rocks or than any other rocks exposed at the Earth's surface. Much of this material included within kimberlites is now considered not to be cognate to the kimberlites and is thus xenolithic or xenocrystic; however, some may be at least partly related to the kimberlite by petrogenesis. In this paper we are concerned only with material of upper mantle origin, and ignore inclusions probably derived from the crust. The importance of the deep material sample brought up by kimberlites has long been recognized and an excellent early summary of the nodule types found in southern Africa was given by Wagner (1914).

A simple consideration of the variety of xenoliths found in any one kimberlite suggests that the upper mantle must be heterogeneous. This study is directed towards assessing the scale and importance of the chemical variations that exist by consideration of the individual components and of the processes that produce and modify them within the mantle. The data and discussion are essentially limited to southern Africa (figure 1) but it may be noted that closely similar nodules occur in other areas (see, for example, Sobolev 1974; McCallum & Eggler 1976).

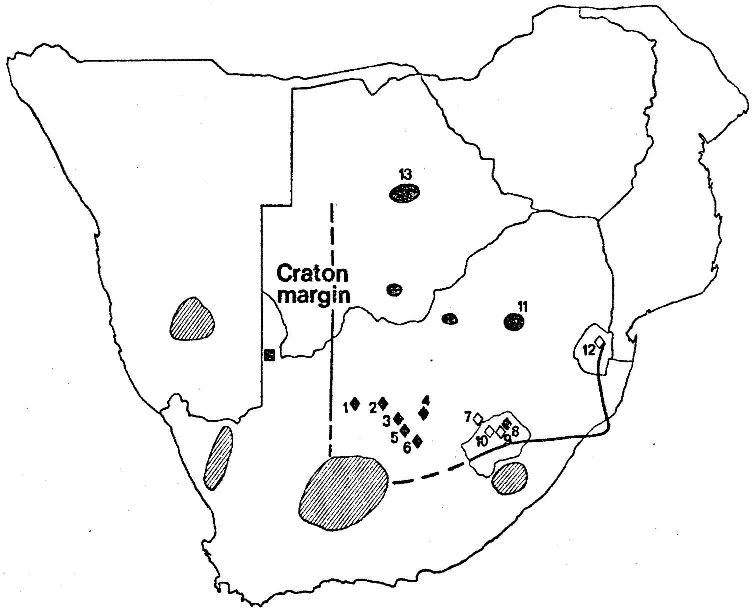

Figure 1. Distribution of some kimberlites in southern Africa. The heavy line indicates the southern margin (largely after Reeves 1977) of the Kaapvaal craton (other lines are boundaries of countries). Shaded areas, with clusters of non-diamondiferous kimberlites; black areas, with clusters of diamondiferous kimberlites; black diamonds, diamond mines; white diamonds, diamondiferous kimberlites; black square, non-diamondiferous kimberlite. Numbered localities: 1, Finsch; 2, Bellsbank; 3, Kimberley; 4, Roberts Victor; 5, Koffiefontein; 6, Jagersfontein; 7, Monastery; 8, Letseng la Terae; 9, Matsoku; 10, Thaba Putsoa; 11, Premier group; 12, Dokolwayo; 13, Orapa group. There is uncertainty in the south central region both with respect to the position of the craton margin and as to whether the large shaded kimberlite area is wholly non-diamondiferous.

2. Principal varieties of upper mantle materials in kimberlites

In a recent paper (Harte 1978) the inclusion suite in kimberlites was divided into five categories: (*a*) peridotites and dunites, (*b*) garnet–pyroxenites, (*c*) eclogites and grospydites, (*d*) megacrysts (discrete nodules) and (*e*) amphibole-bearing and mica-rich types. In this paper we further subdivide categories (*a*), (*d*) and (*e*) to separate particular chemical or petrogenetic associations for which there are reasonable data. In addition some discussion of inclusions in diamonds and of the chemical composition of kimberlite itself is made below. Table 1 indicates the general characteristics of the resultant list of mantle-derived rocks and minerals, and more specific features shown by nodules collected from one location (where the nodule relations are constrained by their possible derivation from a single path through the upper mantle). In this section aspects of the information given in table 1 are amplified, with the particular exception of the metasomatic rocks (categories VIII and IX), which are largely considered separately in §3.

(*a*) *Peridotites, pyroxenites and dunites*

The peridotites of table 1 are lherzolites and harzburgites commonly with some garnet and sometimes with a chromite-rich spinel; rocks with spinels of high Al_2O_3/Cr_2O_3 ratio are extremely rare and are not considered in this review. The common coarse peridotites (category I of table 1) are the major mantle rock type of the depth range generally sampled by kimberlite as judged by the overall xenolith population. Olivine and orthopyroxene are always the principal modal minerals of category I peridotites, with clinopyroxene and garnet generally being less than 5 and 11% respectively (Chen 1971; Cox *et al.* 1973). In general, these rocks (also referred to as granular peridotites: see review of terminology by Harte 1977) show little obvious petrographic evidence of their evolutionary history, but characteristically have depleted† chemical compositions; they are often regarded as residua generated in presumed magmatic events but evidence in favour of this rather than a cumulate origin, for example, is limited and more than one origin may be involved (Gurney *et al.* 1975; O'Hara *et al.* 1975). In addition to restricted bulk chemical compositions, the category I peridotites have mineral compositions indicative of fairly restricted and relatively low temperatures and pressures of formation which are compatible with calculated shield geotherms (see §4 and figure 3); thus, if a magmatic origin is assumed, these rocks appear to have cooled and re-equilibrated metamorphically. The harzburgites with high-Al_2O_3 orthopyroxene (category VI, table I) described by Dawson *et al.* (this symposium) may represent potential common garnet–lherzolites which have not equilibrated to the normal temperature–pressure conditions, but in some cases they may also come from shallower depths than usual.

Rarely, coarse peridotites are found with more clinopyroxene and garnet and distinctly higher FeO/MgO and CaO, Al_2O_3, TiO_2 contents (discounting metasomatic effects) than the common peridotites. Gradations occur both modally and chemically from these high FeO/MgO peridotites to garnet–pyroxenites, and these rocks are placed collectively in category II of table 1. The fertile compositions of rocks in this category have led to the suggestion that they may represent essentially 'primary' upper mantle material and can be used in computations of 'source' mantle (Chen 1971; Berg & O'Hara 1973; O'Hara *et al.* 1975); but the most extensive data on such rocks from one locality (Matsoku) led to the conclusion that they represent igneous cumulates (Cox *et al.* 1973; Gurney *et al.* 1975) and such an origin, possibly of Karroo age, was supported by isotopic data (Kramers 1977, 1979). Dunites (category III a) with relatively high FeO/MgO also indicate the presence of cumulates (Gurney *et al.* 1975; Boyd & Nixon 1978). Mineral data on category II rocks indicate temperatures and pressures similar to the category I peridotites, but exsolution in pyroxenes from some localities (e.g. Koffiefontein) testifies to a previous higher-temperature history.

Peridotite xenoliths showing significant deformation are divided into two major categories (IV and V). The first group shows a wide range in deformational features with gradations from the coarse (undeformed) rocks through porphyroclastic and mosaic–porphyroclastic types (Harte *et al.* 1975; Dawson *et al.* 1975; Harte 1977), but the modal and chemical (mineral and bulk) characteristics are commonly those of the undeformed category I peridotites; hence category IV peridotites appear to be deformed varieties of the commonest peridotites. Category V rocks, on the other hand, in addition to their deformed characteristics (Boullier & Nicolas 1975; Boyd & Nixon 1975; Harte 1977), have distinct mineral and bulk chemical characteristics. In particular, the clinopyroxenes in category V peridotites are subcalcic diopsides (Stephens & Dawson 1977) with distinctly lower Ca/(Ca + Mg) than the chrome diopsides

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† The terms 'depleted' and 'fertile' as used in this paper refer qualitatively and relatively to the amount of basaltic melt that given rock types might yield under a specific set of conditions; they are not used to imply any parent–daughter relationships between the rock types.

TABLE 1. ROCKS AND MINERALS ERUPTED WITH KIMBERLITE

type		*general features: data from all pipes*	*specific features shown at single pipes*
I	common coarse peridotites	Universal distribution and abundant. Rocks largely of olivine + opx. with little cpx. and garnet. Very small range in bulk and mineral compositions. Depleted characteristics for major and trace elements. Low to medium *T*–*P* range of equilibration. Fo: 91–95. Ca/(Ca + Mg): 43–49.	Range in bulk and mineral compositions, and range in equilibration conditions may be very tightly restricted. Degree of depletion partly characteristic of locality, e.g. Bultfontein, Matsoku, pipe 200, Thaba Putsoa.
II	pyroxenites and peridotites with high FeO/MgO	Widespread but usually very rare. Wide bulk and mineral compositional range. Relatively high Fe/Mg, Ca, Al, Na, Cr and trace elements compared to other peridotite types. Usually little deformed. Some restricted *T*–*P* range as I. Fo: 83–89.	Occasionally fairly common. Koffiefontein: subsolidus exsolution in pyroxenes. Matsoku: spread of compositions at one location indicates cumulate origin.
III	dunites	Widespread, sometimes common. Two varieties: (*a*) Fo: 93–95 (often coarse). (*b*) Fo: 88–90 (usually finer grained).	Both varieties at one location (e.g. Bultfontein, Matsoku). (*a*) are most depleted rocks (cf. I). (*b*) are most magnesian cumulates (cf. II).
IV	deformed peridotites with calcic cpx.	Widespread and common. Frequently modally and chemically indistinguishable from type I. *T*–*P* range also as I.	Bultfontein: *T*–*P* range overlaps type I rocks but displaced towards slightly higher values. Matsoku: Restricted *T*–*P* range indistinguishable from all other rocks.
V	deformed peridotites with sub-calcic cpx.	Abundance highly variable; common in some pipes, absent in others. Fe/Mg of bulk rock and minerals typically higher than types I and IV but lower than type II. Ti enriched in mineral's compared with I; also K in cpx. High *T* of equilibration. Possible association with megacrysts. Fo: 88–92. Ca/(Ca + Mg): 28–38.	Equilibration *T* often very tightly restricted and distinct from those of most type I rocks (e.g. Frank Smith, Premier, Thaba Putsoa). Equilibration conditions similar to ilmenite-free megacrysts. Distinguishing chemical features seen in comparisons at single pipe.
VI	harzburgites with high-Al_2O_3 opx.	Some possibly equivalent to type I rocks with cpx. and garnet not exsolved from opx. Others may be of shallower origin.	No detailed study available.
VII	eclogites and grospydites	Universal distribution; usually rare but occasionally very common. Diamond and graphite accessory minerals; no olivine. Very wide range in bulk rock and mineral compositions. Wide range of equilibration *T*.	Roberts Victor: total range of rock types in chemically coherent series (magmatic origin?) with ancient date of formation. Evidence of sub-solidus re-equilibration.

VIII	xenoliths showing infiltration metasomatism	Probably widespread at least in formation of primary phlogopite; rocks of all above categories may be affected with vein and more extensive effects. Associated minerals: phlogopite, amphiboles, ilmenite, rutile, sulphides, unusual opaque. Involves introduction of: K, Ti, Fe, Cu, Ni Nb, Zr, Hf, Ba, Rb, Sr, Na, S, H_2O, (CO_2).	Metasomatism occurs *in situ* in mantle, and probably not related to kimberlite event. Mineralogy and textures variable, e.g. Bultfontein: richterite association. Matsoku: phlogopite–ilmenite–rutile association.
IX	xenoliths showing diffusion metasomatism	Quite commonly seen in eclogitic xenoliths. May be important in peridotitic rocks on scale larger than common xenoliths.	Matsoku: some peridotites–pyroxenites demonstrate extensive exchange across rock contacts for all elements except Cr. Roberts Victor: eclogite–grospydite xenoliths show marked mineral composition gradients in one specimen.
X	megacrysts or discrete nodules	(*a*) Cr-poor variety: widespread, occasionally very common. Wide range in chemistry and equilibration temperature. Cooler megacrysts intergrown with ilmenite. Hotter megacrysts indicate possible relation to V. Little deformed. High Fe/Mg and Ti, low Cr. Fo: 84–89. (*b*) Cr-rich variety: uncertain distribution and overall features. Mineral compositions overlap those in type I rocks.	(*a*) Wide range of T of equilibration at single pipe. Monastery: full range of types and compositions showing gradual and coherent chemical variation which is related to wide T range of equilibration; small P range; magmatic origin. Kimberlitic inclusions in some olivines. (*b*) No detailed study available.
XI	diamond inclusions	Peridotite minerals and eclogite minerals probably cogenetic with diamond. Peridotitic inclusions have restricted and depleted chemistry (Fo: 93–94) and low–medium equilibration T. Eclogite minerals have wide range in chemistry and equilibration T.	Each pipe has both peridotitic and eclogitic inclusions in proportions which vary widely and do not correlate with xenolith abundances. Eclogite inclusions show full range of xenolith compositions at one locality.
XII	Marid suite	Widespread, sometimes common. Rocks consisting of: mica, amphibole, rutile, ilmenite, diopside, and zircon. Related to kimberlite and of relatively shallow origin.	No detailed study available.
XIII	kimberlite	Very variable in composition with high concentration of incompatible elements, CO_2 and H_2O.	Several varieties (intrusions) often at one pipe. A single pipe may show all types (I to XII) of inclusions.

opx., orthopyroxene; cpx., clinopyroxene; Fo, MgO/(MgO + FeO) molecular percentage in olivines; Ca/(Ca + Mg), atomic percentage in clinopyroxenes. 'Deformed' refers essentially to rocks with porphyroclastic and mosaic–porphyroclastic textures (Harte, 1977).

Literature references for pipes noted: Bultfontein (Kimberley), Dawson *et al.* (1975), Erlank & Rickard (1977), Boyd & Nixon (1978), Lawless (1978); Frank Smith, Boyd (1974); Matsoku, Cox *et al.* (1973), Harte *et al.* (1975), Harte & Gurney (1975*a*), Harte *et al.* (1977); Monastery, Jakob (1977), Gurney *et al.* (1979); Pipe 200, Carswell *et al.* (1979); Premier, Danchin & Boyd (1976), Danchin (1977); Roberts Victor, Macgregor & Carter (1970), Lappin & Dawson (1975), Hatton (1978); Thaba Putsoa, Nixon & Boyd (1973*a*, *b*), Boyd & Nixon (1975).

of all other peridotite xenoliths. This feature indicates a higher temperature origin for category V deformed peridotites (see §4 and figure 3). In addition these peridotites often have FeO/MgO, Na_2O, CaO, Al_2O_3 and TiO_2 values (percentages by mass) which are higher than those of the common category I peridotites (excluding rocks obviously affected by metasomatism), but partly lower than those of category II peridotites and pyroxenites. The category V peridotites are therefore distinguished on a combination of mineral and bulk chemical characteristics in addition to their deformation–recrystallization textures. The less depleted chemical features of the category V peridotites have led to the suggestion (Nixon & Boyd 1973*a*; Boyd & Nixon 1975) that they have suffered less melt extraction than the common peridotites (category I), and this has been supported by trace element studies (Shimizu 1975*b*; Shimizu & Allègre 1978). In §4 we suggest that these less depleted characteristics are largely a product of diffusion (*not* infiltration) metasomatism between the peridotites and an adjacent magma body.

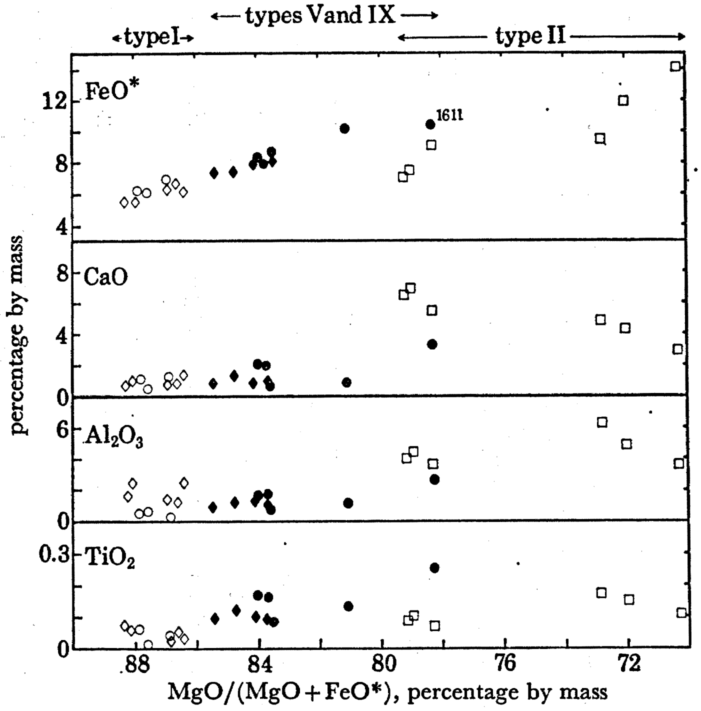

FIGURE 2. Bulk compositional relations of periodotite nodules from adjacent northern Lesotho kimberlites: Thaba Putsoa and Mothae (Nixon & Boyd 1973*a*, table 19) and Matsoku (Cox *et al.* 1973, table 26; Harte *et al.* 1977 and unpublished). Open circles (Thaba Putsoa and Mothae) and open diamonds (Matsoku) are common coarse peridotites (category I). Solid circles (Thaba Putsoa and Mothae) are deformed high-temperature peridotites (category V). Solid diamonds (Matsoku) are category I peridotites modified by diffusion metasomatism to become category IX. Open squares (Matsoku) are high FeO/MgO peridotites–pyroxenites (category II). FeO* is total Fe as FeO. 1611 is an exceptionally fertile category V peridotite from Thaba Putsoa (see text).

Figure 2 summarizes some of the chemical characteristics of the various peridotite types by using data for nodules from northern Lesotho kimberlites. The compositions shown are for rocks that lack reported evidence of infiltration metasomatism, but they include data for some rare coarse peridotites affected by diffusion metasomatism at the margins of intrusive magmatic sheets (Harte *et al.* 1977) and which show compositional similarities to the deformed high temperature peridotites (category V). The brackets indicated for peridotite types in figure 2 are not suggested to apply universally. Data for nodules from the Premier pipe (Danchin 1977) show more overlap between the category I and category V peridotites and indicate that category V peridotites with the high FeO/MgO of rock 1611 (figure 2) are very unusual.

(*b*) *Eclogites and grospydites*

This category of xenoliths (VII in table 1) is represented at virtually all kimberlite localities, but is usually much less abundant than the category I peridotites. By comparison with the peridotitic range of xenoliths in which diamond or graphite are extremely rare, the eclogitic xenoliths contain diamond and/or graphite relatively frequently (see, for example, Reid *et al.* 1976). Mineral compositional variations are more frequent in individual xenoliths of eclogite than peridotite (see §3), but many specimens are reasonably homogeneous. Overall, the eclogites and grospydites show a very wide range of bulk and mineral chemical compositions. At Roberts Victor mine, where these rocks are most abundant, this range of composition is coherent and gradational in a series of homogeneous samples (Hatton 1978). The most magnesian rocks (with approximate bulk mass percentages MgO 19.0, CaO 4.0, Al_2O_3 8.0) have high Cr_2O_3 contents (up to 1.6%) and may contain chromite or orthopyroxene; they are linked by a continuous chemical series to Cr_2O_3-poor, Na_2O-rich kyanite-bearing rocks (with approximate bulk mass percentages MgO 6.0, CaO 14.0, Al_2O_3 26.0). Hatton (1978) suggests that this coherent suite of Roberts Victor xenoliths is the product of differentiation processes associated with a single magmatic event. A common and ancient time of origin of the Roberts Victor suite is indicated by Kramers (1977) on the basis of lead isotope ratios.

Rare specimens show contacts between peridotite and eclogite. Hatton (1978) describes a chrome-rich eclogite in contact with altered peridotite at Roberts Victor; we have found a magnesian orthopyroxene-eclogite in contact with peridotite in a specimen from Newlands.

Estimates of the temperatures of formation of eclogite minerals (§4, figure 3) cover a range of approximately 400 °C, and similarly large ranges are found for eclogite xenoliths from single localities (Roberts Victor (Hatton 1978); Orapa (Shee 1978); Premier (R. V. Danchin, unpublished, and Gurney *et al.* 1980). The validity of these estimates is very uncertain but it seems unlikely that eclogite bodies such as that postulated by Hatton (1978) extend over depths equivalent to 400 °C on a normal geotherm. Also, exsolution features are quite often found in eclogites and, together with mineral composition heterogeneities, these indicate incomplete metamorphic reconstitution after a higher temperature, presumably igneous, event (Harte & Gurney 1975*b*).

(*c*) *Megacrysts*

Single crystals of large size (commonly exceeding 1.0 cm) form widespread and sometimes abundant inclusions in kimberlite; they are referred to both as megacrysts and discrete nodules. Smith *et al.* (1976) recognize two separate groups of kimberlite megacrysts: Cr-poor (category X*a*, table 1) and Cr-rich (category X*b*). Though the latter are recorded (e.g. by Shee 1978) little is known concerning their overall nature and distribution in southern Africa, and the following notes concern only the Cr-poor type.

The Cr-poor megacryst group includes large crystals of orthopyroxene, clinopyroxene, garnet, olivine and ilmenite. Smaller inclusions of one of these minerals may be found within a large single crystal of another, and in some cases regular lamellar intergrowths occur, especially of clinopyroxene and ilmenite and to a lesser extent of orthopyroxene and ilmenite. The megacrysts show a substantial range of chemical variation, including FeO/MgO. The compositions are largely different from those of peridotite xenoliths, but the more magnesian megacrysts show some similarities to minerals in category V peridotites (Nixon & Boyd 1973 *b*). Chemical considerations in addition to inclusion and intergrowth relations indicate that the various crystalline species of the megacryst group have often been in equilibrium with one another (Nixon & Boyd 1973 *b*; Jakob 1977; Gurney *et al.* 1979).

The overall chemical features, large size, and association with ilmenite, seen in Cr-poor megacrysts have led to the view that they are phenocrysts or other crystals physically associated with magma; and that differentiation of this magma is responsible for the chemical variations of the megacrysts (Nixon *et al.* 1963; Nixon & Boyd 1973 *b*; Gurney *et al.* 1979). Some inclusions within megacrysts suggest that the magma may be related to or may become kimberlite (Haggerty & Boyd 1975; Jakob 1977; Rawlinson & Dawson 1977; Gurney *et al.* 1979; Pasteris *et al.* 1979).

Temperature estimates for the formation of megacrysts cover a wide range (§4, figure 3), even when taken for a single kimberlite locality, and there is a notable association of Mg-rich megacrysts with higher temperatures, and Fe-rich megacrysts (including those containing ilmenite) with lower temperatures (Nixon & Boyd 1973 *b*; Boyd & Nixon 1973; Gurney *et al.* 1979), these features again being linked to a differentiating magma body. Pressure estimates for the megacrysts are more restricted, and the most extensive study of abundant megacrysts from a single location (Monastery Mine) indicated that all megacrysts might have formed at the same depth (Gurney *et al.* 1979).

(*d*) *Inclusions in diamonds*

Natural diamonds contain inclusions of minerals that are considered to be in equilibrium with each other and the host diamond (Sobolev 1974). Recent studies on ten sources in southern Africa (Hawthorne *et al.* 1980) and a similar study on Yakutian diamonds (Yefimova & Sobolev 1977) shows that more than 99% of all inclusions in diamonds may be assigned to either an eclogitic or a peridotitic suite. Often peridotitic inclusions greatly predominate in diamonds from one kimberlite pipe, even at Roberts Victor where xenoliths are largely eclogites. However, eclogitic suite inclusions appear to be always present and are common at some localities (e.g. Premier) or even predominate (e.g. at Orapa).

The eclogitic mineral inclusions show a wide range in compositions, equivalent to the range shown by the eclogite xenoliths, from Cr-rich magnesian compositions to kyanite-bearing assemblages. Temperatures of formation of eclogitic inclusions cover a broad range (figure 3) also similar to the xenoliths. It has been demonstrated (Gurney *et al.* 1980; Hawthorne *et al.* 1980) that the general range in compositions shown by minerals in the xenoliths is shown by the inclusions, and that 15 gt/cpx. inclusion pairs at Premier have a range of Fe/Mg K_D ratios equivalent to a possible range in temperature of 300 °C (such temperature ranges from minerals in diamond may be more reliable than those from xenoliths because Fe^{2+}/Fe^{3+} variations are likely to be minimized). Thus eclogitic suite inclusions may have originated in a broadly similar manner to eclogite xenoliths, and this accords with the relatively common occurrence of diamonds in the eclogite xenoliths.

Peridotitic suite inclusions in diamond have restricted chemical compositions which overlap but tend to be more Mg-rich, Cr-rich and Ca-poor than the minerals in category I peridotites. Temperature and pressure estimates for these inclusions are limited by the fact that olivine and orthopyroxene are the most commonly occurring minerals. In general, estimates fall in the lower part of the total peridotite field (figure 3). In this respect and in their major element compositions, the peridotitic suite inclusions appear distinct from and unrelated to the megacrysts and high temperature deformed peridotites (category V). We consider that they most probably originate by crystallization from (kimberlitic) magmas generated by partial melting of category I peridotite in the presence of CO_2 (Harte *et al.* 1980).

(*e*) *Marid suite and kimberlite*

The Marid suite has been described by Dawson & Smith (1977) and includes unusual and distinctive rocks consisting of varying proportions of mica, amphibole, rutile, ilmenite, diopside, and zircon. The last named phase was predicted but not observed by Dawson & Smith; its presence has been substantiated in more than 50 Marid suite xenoliths from Kimberley by Bayly *et al.* (1980). Dawson & Smith suggest that the Marid suite has an igneous origin by crystallization from a kimberlitic magma within the higher parts of the upper mantle. Polymict peridotites, consisting of a wide variety of mantle rock and mineral clasts cemented by phlogopite, ilmenite, rutile, and sulphides (described by Lawless *et al.* 1977), are probably related rocks which, since amphibole has not been recorded, may have a slightly deeper origin.

The kimberlites which transport the various nodule types to the surface show wide variations in chemistry (see, for example, Gurney & Ebrahim 1973). The high concentrations of K, Ti, other incompatible elements and volatiles in kimberlite contrast with their concentrations in the upper mantle xenoliths (excepting to a limited extent rocks affected by infiltration metasomatism – §3) erupted with the kimberlite. In this respect the kimberlite matrix (and the probably closely related Marid suite nodules) testifies to the generation of extreme chemical heterogeneities within the upper mantle.

3. MANTLE METASOMATISM

Categories VIII and IX in table 1 differ from the other categories in that they do not refer to particular petrological types but to rocks of various types (largely corresponding to I–VII in table 1) that show clear evidence of having been modified by metasomatic processes. The two categories are separated from one another according to whether the metasomatic process involved appears to be one of largely infiltration or diffusion (Korzhinskii 1970). In the latter case, diffusive transfer across boundaries in upper mantle rocks has largely resulted in modifications of pre-existing mineral chemistry. In the former case, movement of a fluid appears to be involved, and the mantle rocks have been modified more particularly by the introduction of elements which lead to the formation of new minerals, though there may also be changes in pre-existing mineral chemistry. Diffusion metasomatic effects may be highly significant in the formation of specific rocks or the evolution of limited volumes of magma, but it is infiltration metasomatism that generates the most marked chemical heterogeneities in the nodule suite from kimberlites.

(*a*) *Xenoliths showing infiltration metasomatism*

Many of the rock inclusions within kimberlite yield evidence of their penetration by fluids that leave in the rock a record of mineral development distinct from that of magma (though this does not exclude the emanation of the fluids from magmas). At least two groups of infiltration metasomatic events must be distinguished:

(*a*) Those that have occurred while the rocks were *in situ* within the upper mantle, i.e.

before inclusion in kimberlite. Phlogopite appears to be the most abundant and widespread mineral formed as a consequence of such metasomatism in both peridotites and eclogites. Also documented as metasomatic minerals in peridotites are: potassic richterite, ilmenite, rutile, sulphides, a Ba- and Zr-rich opaque mineral, and carbonate (Cox *et al.* 1973; Harte & Gurney 1975*a*; Erlank & Rickard 1977). In eclogites, Shee (1978) records pargasitic hornblende and edenite in addition to the widespread phlogopite.

(*b*) Those of late stage origin, associated with the incorporation of nodules into kimberlite and the upward movement of kimberlite towards the surface. Phlogopite associated with serpentine and kelyphite are common examples of material of this origin.

For brevity, minerals belonging to group (*a*) might be referred to as *primary* and those of group (*b*) as *secondary*, but it must be noted that since group (*a*) minerals have been introduced into a pre-existing rock they are also secondary in one sense even when they are in equilibrium with the other minerals in the same rock. To avoid any confusion in this situation Harte *et al.* (1975) referred to group (*a*) as *primary-metasomatic* and group (*b*) as *late-secondary*. More recently, Boyd & Nixon (1978) have used the terms 'mantle-derived' and 'late-stage' to distinguish phlogopites formed respectively before and after inclusion in kimberlite. The primary metasomatic mantle suite is clearly the more important of the two groups from the viewpoint of upper mantle processes, compositions and heterogeneities, and it therefore forms the principal subject of further discussion.

A petrographic and chemical distinction between primary and secondary phlogopite in nodules from the Kimberley area has been made by Carswell (1975) and Boyd & Nixon (1978). The occurrence of both types in depleted garnet–peridotites provides circumstantial evidence for their metasomatic origin (Boyd 1973, p. 2541), while the deformation of primary phlogopite indicates its formation before inclusion within kimberlite (Boyd & Nixon 1978, p. 1368).

More extensive and direct evidence concerning the primary-metasomatic nature of phlogopite and other minerals has come from studies on xenoliths from Matsoku (Cox *et al.* 1973; Harte *et al.* 1975; Gurney *et al.* 1975; Harte & Gurney 1975*a*) and studies of richterite-bearing xenoliths (Erlank 1976; Erlank & Rickard 1977). At Matsoku the primary-metasomatic mineral suite includes phlogopite, ilmenite (with spinel exsolution), rutile (with spinel exsolution) and sulphides (largely pyrrhotite, pentlandite and chalcopyrite). These minerals may be disseminated throughout the matrix of deformed and recrystallized nodules, but they also occur in veinlets, narrow zones and diffuse patches in both coarse and finer-grained peridotites and pyroxenites (Cox *et al.* 1973; Harte *et al.* 1975; Gurney *et al.* 1975; Harte & Gurney 1975*a*). The frequent concentration of the primary-metasomatic minerals in veinlets or narrow zones within depleted peridotites (categories I and IV, table 1), together with the recrystallization or other textural modification of the common peridotite minerals (olivine, orthopyroxene, clinopyroxene, garnet) in such zones, is considered to be strong evidence for the formation of the primary-metasomatic minerals as a result of fluid infiltration into a pre-existing rock. Unequivocal evidence in support of this is seen where an unfoliated vein carrying primary-metasomatic minerals cross-cuts the foliation of a flaser (porphyroclastic) peridotite (Harte *et al.* 1975, p. 497).

Chemical as well as textural features indicate that the Matsoku primary-metasomatic minerals are not the product of straightforward contamination of nodules by their host kimberlite. The primary-metasomatic minerals show restricted and distinct mineral compositions (Harte & Gurney 1975*a*), and the phlogopites have similar compositions to the primary phlogopites of Carswell (1975) and Boyd & Nixon (1978). The normal peridotite minerals (olivine, orthopyroxene, clinopyroxene, garnet) recrystallized in association with the metasomatic veining show closely similar compositions to those found in rock volumes free from metasomatic effects. Overall it is apparent that the metasomatic veins and rocks were in a state of reasonable chemical and textural equilibrium at the same T–P conditions as Matsoku rocks unaffected by metasomatism (Gurney *et al.* 1975; Harte & Gurney 1975*a*), and thus that the metasomatism pre-dated incorporation within the kimberlite. The possibility remains, however, as tentatively suggested by Harte *et al.* (1975), that the fluid responsible for the metasomatism did derive, at some distance, from the developing kimberlitic magma before incorporation of the xenoliths within the erupting kimberlite. More recent evidence suggests the probability that the metasomatic mineralization seen at Matsoku is connected with a magmatic event associated with the intrusion of clinopyroxene-rich sheets (Harte *et al.* 1977). Isotopic evidence indicates that this event may be related in age to the Karroo magmatism (Kramers 1977, 1979) and is thus distinctly earlier than the kimberlite.

With respect to the peridotite nodules containing potassic richterite, Erlank & Rickard (1977) note that the evidence for formation of the richterite together with phlogopite, opaques and calcite by metasomatism includes the occurrence of poikiloblastic and replacement textures as well as vein structures. The occurrence of this metasomatism before inclusion in kimberlite is shown by the deformation of richterite and phlogopite in a porphyroclastic peridotite. In addition, Sr isotope evidence (Erlank & Shimizu 1977) indicates that the metasomatic event pre-dates kimberlite emplacement and is isotopically unrelated to the kimberlite.

From the nature of the metasomatic minerals noted, it is evident that they must represent veritable treasure trove with respect to those elements in which most xenoliths from kimberlite are described as depleted; for example, Erlank & Rickard (1977) report a richterite-bearing peridotite with 2.1% K_2O by mass; while a rutile-rich primary-metasomatic vein from Matsoku contains 18.7 wt % TiO_2 by mass. Rutile in the Matsoku metasomatic rocks (Harte & Gurney 1975*a*) also yields high ZrO_2 (0.3–0.7) and Nb_2O_5 (0.2–1.1); and Erlank & Rickard (1977) report an opaque phase with BaO 9.1; ZrO_2 7.4; SrO 2.4; and Nb_2O_5 0.5 (all percentages by mass).

The infiltration metasomatic phenomena are important in that they indicate the presence and independent movement within upper mantle rocks of fluids whose chemistry and motion are capable of leading to the formation of minerals: rich in Fe, with appreciable K, Ti, H_2O, and sometimes with significant S, Nb, Zr, Hf, Sr, Cu, Ni, Rb, Ba, Na and CO_2. The formation and migration of such fluids testifies to the occurrence of processes of intense segregation of elements, generating marked chemical heterogeneities, in the rocks constituting the lower lithosphere. It is possible, considering the restricted view of spatial relations given by kimberlite sampling, that the metasomatic fluids are closely related to and derived from an adjacent magma body – i.e. that they indicate activity essentially akin to pneumatolysis. Thus we have not suggested nor intended to imply that mantle metasomatism, as seen in nodules, indicates some large-scale process involving nebulous ichors drifting wantonly through an unsuspecting mantle. It must nevertheless be recognized that even if the metasomatism is limited to a close association with magmatic phenomena the possible extent of magmatism in the evolution of the lithosphere allows the metasomatism considerable scope in creating widespread lithological heterogeneities. Furthermore, the rarity of hydrous and carbonate minerals in the deeper part of the upper mantle does suggest the possible pervasive occurrence of H_2O- and CO_2-bearing fluids that have been linked to the existence of a low-velocity zone in many regions (Lambert & Wyllie 1968; Wyllie & Huang 1975; Eggler 1976). These fluids may or may not be magmatic, but in either case it seems likely that some metasomatism will be associated with them, and that the resultant metasomatic rocks will be distributed across extensive areas.

(*b*) *Xenoliths showing diffusion metasomatism*

Diffusive transfer of elements is evidently an important factor in the attainment of local chemical equilibrium within rocks generally. The efficacy of diffusion in mantle rocks, especially peridotites, is shown by the homogeneity of mineral compositions in many xenoliths from kimberlite (see, for example, Boyd & Finger 1975); and such homogeneity even applies to rocks showing modal layering (Gurney *et al.* 1975). However, large chemical gradients have been documented in eclogite clinopyroxenes showing exsolution (Harte & Gurney 1975*b*). Diffusion metasomatism across contacts between units of differing mineral chemistry in single xenoliths has been shown most particularly in eclogite xenoliths. Marked chemical gradients have been recorded in both chrome-rich and kyanite-bearing eclogites by Chinner & Cornell (1974), Lappin & Dawson (1975) and Hatton (1978).

Some rare xenoliths show contacts between magmatically intruded clinopyroxene-rich sheets and a host rock with the modal and textural characteristics of common coarse peridotite (category I). Despite this intrusive relation, the mineral compositions across the contacts are the same for most elements (with the particular exception of Cr), and the host rock minerals show notably higher Fe/Mg and Ti than those usual for category I peridotite (Harte *et al.* 1977). These features are attributed to diffusion metasomatism between the intrusive sheets and the host peridotites. The bulk chemical compositions of the latter rocks show 'less depleted' characteristics, similar to those of high-temperature deformed peridotites (category V, table 1) as shown in figure 2.

4. DISTRIBUTION OF THE ROCK TYPES WITHIN THE UPPER MANTLE

(*a*) *Lateral and age-related variations*

Evidence of broad regional variations in the nature of inclusions contained in southern African kimberlites is limited to changes across the margins of the Kaapvaal craton (figure 1):

(1) diamondiferous kimberlites are essentially confined to the craton;

(2) within the craton, the kimberlites near the craton margins show less economic potential for diamond mining than those in central regions;

(3) the Al_{En} content of orthopyroxenes in coarse garnet–peridotite xenoliths is in general lowest for kimberlites on the craton, increases towards craton margins, and is highest for xenoliths from kimberlites off the craton;

(4) considering the predominant kimberlites of post-Karroo age, the younger intrusives lie off the craton, while the majority of diamond mines are in kimberlites with intrusive ages close to 90 Ma (Davis 1977).

Aspects of these features have been noted and discussed by Clifford (1966), Dawson (1970) and Nixon *et al.* (1973), and they are thought to relate principally to differences in lithospheric thickness and heat flow.

Within the Kaapvaal craton (figure 1), clusters of kimberlites occur and these clusters are distributed over an extensive geographic area. All categories of nodules listed in table 1 can be found in each cluster, while in localities such as Jagersfontein, Frank Smith, Premier or the Kimberley Mines, most categories may be found in a single pipe. At the same time large differences in dominant nodule types occur between closely spaced pipes such as Roberts Victor and Wesselton, Jagersfontein and Koffiefontein, Bellsbank and Finsch, or Thaba Putsoa and Matsoku. Clearly these features do not indicate the presence of broad lateral variations in the rock types present in subcratonic mantle; but might, depending upon the circumstances at depth governing the kimberlite sampling, indicate much relatively localized upper mantle heterogeneity with respect to the proportions of rock types present.

It should be noted that the Precambrian diamondiferous kimberlite of Premier has an inclusion suite closely similar to that shown by the dominant approximately 90 Ma old diamondiferous kimberlites (Danchin & Boyd 1976), and thus provides no evidence of variation of mantle lithology over the time period concerned (*ca.* 1000 Ma).

(*b*) *Vertical or depth-related distribution and variations*

The *T–P* conditions of origin of the rocks and minerals found included in kimberlite may potentially be determined by using experimentally calibrated mineral equilibria. Such information may also be used to construct *T–P* profiles that might represent mantle geotherms (see, for example, Boyd 1973). The principal mineral data used have been: (1) the Ca/(Ca + Mg) ratio of clinopyroxene in equilibrium with orthopyroxene; (2) the Al content of orthopyroxene in equilibrium with garnet; and (3) the Fe/Mg distribution between garnet and clinopyroxene. The first two of these equilibria have been applied particularly to the garnet–lherzolite and garnet–websterite nodules to yield both temperature and pressure estimates (Boyd 1973); they have also been applied to megacrysts on the assumption (substantiated by consideration of mineral compositions) that such crystals were part of a clinopyroxene–orthopyroxene–garnet assemblage (Nixon & Boyd 1973*b*; Gurney *et al.* 1979). The third equilibrium has been predominantly applied to eclogites to give temperature estimates at some assumed pressure (see, for example, Reid *et al.* 1976). All methods involve uncertainties concerning experiments, application of experimental data for restricted chemical systems to complex natural rock systems, and appropriate thermodynamic modelling especially in view of limited experimental and crystal chemical data. Further discussion of the methods and reference lists concerning their application may be found in Wells (1977), Howells & O'Hara (1978), Mori & Green (1978), Powell (1978), O'Hara & Yarwood (1978) and Fraser *et al.* (1978). In general, pressure estimates for xenoliths appear to be much more liable to substantial error than temperature estimates.

Temperature and pressure estimates based on such silicate mineral equilibria may be compared with other *T–P* indicators, such as the experimentally determined curves for diamond–graphite equilibrium or melting of peridotite, or they may be compared with geophysically estimated geotherms showing temperature variation with depth. Again there are uncertainties. Wyllie (1978) and Eggler (1978) demonstrate the wide variation in possible melting temperatures as a function of the content of 'volatiles' (principally H_2O and CO_2). Geotherms calculated from measurements of heat flow, radiogenic element abundances, etc., are dependent not only on the numerical values estimated for such parameters but also upon the relative importance attached to conduction, convection and other types of heat transport, or to specific *T–P* estimates (see, for example, Clark & Ringwood 1964; Tozer 1967; Pollock & Chapman 1977). Harte (1978) constructs an average shield geotherm whose deepest section is fixed around 200 km by the occurrence of mantle convection as modelled by McKenzie *et al.* (1974), McKenzie & Weiss (1975) and McKenzie & Richter (1976). Such a model obviously conflicts with Jordan's (1978) assertion of the existence of thick (400 km) continental tectosphere.

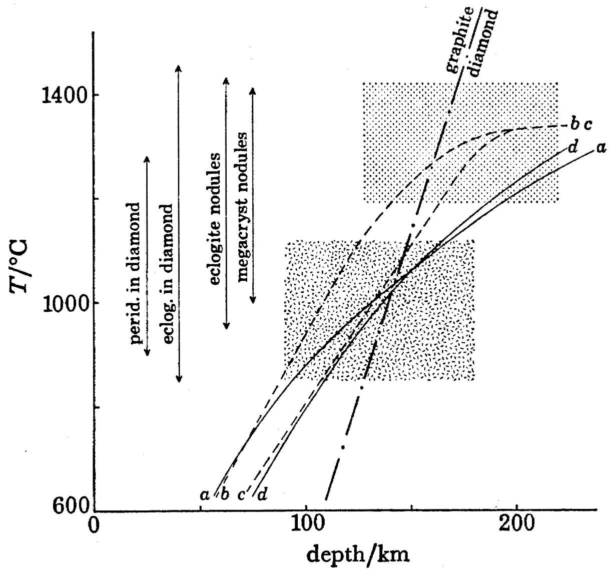

FIGURE 3. Temperature and pressure (≡ depth) estimates relevant to inclusions in kimberlite. Lines *a*, *b*, *c* and *d* are geotherms: *a*, Clark & Ringwood (1964) shield geotherm for 0.042 heat flow; *b* and *c*, shield geotherms for 0.044 and 0.040 heat flows after method of Harte (1978); *d*, Pollock & Chapman (1977) shield geotherm for 0.040 heat flow (all heat flows in watts per square metre). Graphite–diamond reaction line from Kennedy & Kennedy (1976). The ornamented fields give approximate *T–P* ranges for category V peridotites (dotted) and for other (category I–IV) peridotites (dashed). The arrowed lines give probable *T* ranges of formation for peridotitic and eclogitic mineral inclusions in diamonds, as well as for eclogite and megacryst nodules. Principal nodule/inclusion data sources: Boyd (1973, 1974); Boyd & Nixon (1973, 1975, 1978); Nixon & Boyd (1973*a*, *b*); Cox *et al.* (1973); Danchin & Boyd (1976); Dawson *et al.* (1975); Gurney *et al.* (1979); Hatton (1978); Hawthorne *et al.* (1980); Johnston (1973); Lawless (1978); MacGregor (1975); Meyer & Tsai (1976).

Figure 3 shows some estimates of nodule formation and geotherms together with the diamond–graphite reaction curve. The temperature estimates for peridotitic rocks and minerals (including megacrysts) are based on the Ca/(Ca + Mg) ratio of clinopyroxene calibrated against the experimental determination by Davis & Boyd (1966) of the pyroxene solvus. These data are used because they are simply determined, have been widely used, and for most peridotites from kimberlites yield estimates reasonably similar (± 50 °C) to those obtained by using more complex thermodynamic models or other experimental data. The use of other experimental data (see, for example, the compilation in figure 4 of Howells & O'Hara 1978) would tend to increase intermediate- and high-temperature estimates, and more particularly decrease low-temperature estimates: the latter effect might spread the garnet–peridotite sample through to significantly shallower mantle depths. Pressure limits shown in figure 3 for peridotites and pyroxenite nodules are little more than guesses of the likely limits given all the uncertainties noted above. The temperature estimates for eclogitic materials assume a restricted range of pressure conditions as suggested by the diamond–graphite curve or geotherm estimates.

Accepting figure 3 at its face value, it is clear that there is extensive overlap in the conditions of origin of peridotite xenoliths, eclogite xenoliths, megacrysts and diamond inclusions within a temperature range of approximately 850–1450 °C. However, the peridotites are separated into two groups. There is a lower temperature group with a predominant clinopyroxene Ca/(Ca+Mg) atomic percentage range of 43–49, which encompasses the large number and wide variety of nodules belonging to categories I–IV (table 1). The other, higher temperature, group consists only of the category V peridotites with their distinctive subcalcic clinopyroxenes (Ca/(Ca+Mg) predominantly 28–38).

A *T–P* separation between granular peridotites (≡ category I peridotites) and sheared peridotites (≡ category V peridotites) was first noted and discussed in a series of papers by Boyd & Nixon (Boyd 1973; Nixon & Boyd 1973*a*, *b*; Boyd & Nixon 1973, 1975). These authors considered the *T–P* distribution to indicate depth layering in the upper mantle, and associated the sheared peridotites with the inflected (high d*T*/d*P*) segment of a mantle geotherm in a hypothesis whereby plate movements were responsible for the deformation of the deeper peridotites and, as a consequence, for stress-heating which resulted in the geotherm inflection.

Other nodule characteristics were incorporated in their upper mantle model by Boyd & Nixon. The large range of temperatures shown by megacrysts from one pipe locality and the

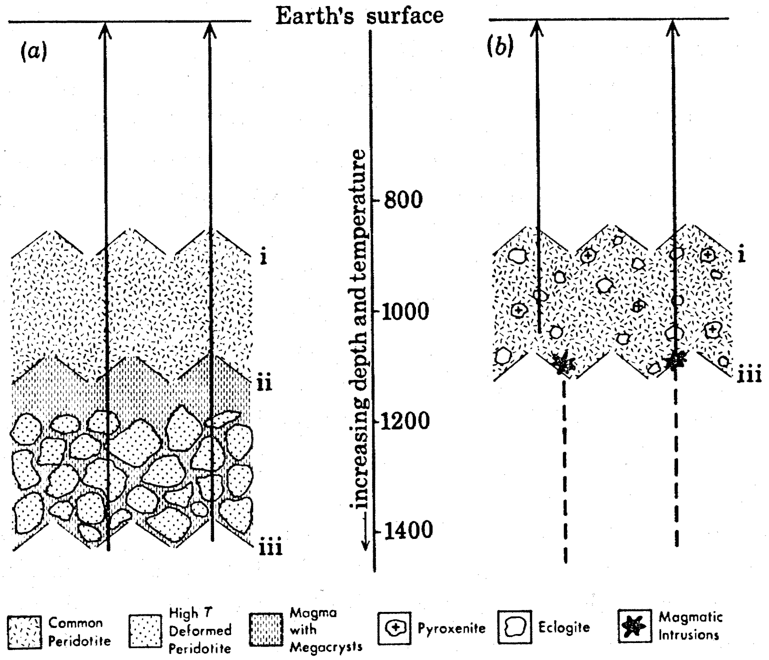

FIGURE 4. Schematic vertical cross sections illustrating models of the upper mantle sampled by kimberlites. (*a*) After Boyd (1973), Nixon & Boyd (1973*a*, *b*) and Boyd & Nixon (1973); (*b*) as proposed by the authors. The common peridotite of (*a*) was originally considered by Boyd & Nixon to be granular, depleted, and undeformed peridotite (category I, table 1) and the high *T* deformed peridotite (category V) was referred to as sheared peridotite. The common peridotite of (*b*) includes categories I, III*a* and IV of table 1, while pyroxenite encompasses all category II and probably category III*b* rocks. The magmatic intrusions in (*b*) indicate the sites of formation of megacrysts and category V peridotites (see text). The boundaries are: (i) upper limit of extensive mantle sampling as shown by garnet peridotites; (ii) lower limit of granular peridotite and phlogopite and chromite in peridotite, as well as top of low-velocity zone and inflexion point on geotherm (Boyd & Nixon 1973, 1975); (iii) approximate lower limit of extensive mantle sampling. The boundaries are zig-zagged to indicate uncertainties in *T* limits and corresponding depths. The long heavy arrows indicate the paths of kimberlites; the broken heavy lines indicate the movement path of the postulated megacryst magmas that cause intrusions at depth in model (*b*). Central numbers indicate rough ambient *T* (in degrees celsius) with increasing depth (temperature perturbation by intrusions at depth in (*b*) is ignored); corresponding depths may be roughly estimated from figure 3.

probably magmatic origin of the megacrysts (see §2) led Nixon & Boyd (1973*b*) to suggest that these crystals formed in crystal-mush magmas dispersed over depths of several tens of kilometres. The correlation of temperature estimates with chemical and 'mineral assemblage' characteristics of the megacrysts (§2) was inferred to indicate that the crystal-mush magmas were differentiated upwards from Mg-rich to Fe-rich and Ti-rich. The overall temperature range estimated for the megacrysts led to their association with the high-T, inflected, limb of the geotherm, and this was supported by the small range of megacryst pressure estimates (Nixon & Boyd 1973*b*). It was also suggested, and supported with chemical and T–P data, that the high-T deformed peridotites (category V) formed solid material between which the crystal–mush magma was dispersed at the greater depths. The upper limit of the magma and the inflexion point of the geotherm were considered to coincide with the top of the upper mantle low velocity zone (Nixon & Boyd 1973*b*; Boyd & Nixon 1973, 1975). The lower depth limits of phlogopite and chromite in peridotite xenoliths were also linked with the top of the low velocity zone.

These views of Boyd & Nixon (essentially dating from 1973) imply a strong zonation in upper mantle lithology with depth and are partly summarized in figure 4*a*. Some aspects of their model (especially concerning deformation, stress heating, and the existence of a subcratonic low-velocity zone) have been revised, but the occurrence of a megacryst-bearing magma dispersed over several tens of kilometres of depth with upward differentiation and enrichment in Fe and Ti remains as a prominent feature of their model (Boyd 1976; Boyd & Nixon 1978; Pasteris *et al.* 1979). Inflected geotherms also continue to be advocated, and as an alternative to the stress heating hypothesis, Boyd (1976) and Boyd & Nixon (1978) suggest the slow upwelling of partly melted material as the cause of the inflexions. Harte (1978) showed that T–P estimates on peridotite nodules from individual pipes do not provide clear evidence for inflected geotherms, and, while the clustering or limited T–P spread of such estimates for individual pipes does not in general preclude the possibility of inflexions, it does throw the burden of evidence for their existence on the interpretation of the megacrysts. From this viewpoint it should be noted that more recent experimental data do not provide evidence for broadening the small P range indicated by megacrysts – indeed, the data of Howells & O'Hara (1978) imply the opposite.

The models of Boyd & Nixon place emphasis on the potential vertical distribution of features shown by kimberlite nodules. Harte (figure 1, 1978) followed this emphasis, but noted that some aspects of the distribution, particularly with respect to the distribution of megacrysts, phlogopite and some deformed peridotites, may be more directly associated with magma or diapir 'aureoles' than depth. In contrast to Boyd & Nixon's interpretation of the megacryst suite, Gurney *et al.* (1979) suggest, on the basis of the study of the extensive megacryst suite from a single location (Monastery Mine), that their wide temperature range reflects the occurrence of differentiation processes at essentially constant pressure in a magma body of limited extent. Following from this we propose a model for the formation of megacrysts and category V peridotites involving short-term horizontal temperature gradients.

We suggest that relatively localized melting occurs at depth within the upper mantle at temperatures a little above those indicated by the highest-temperature megacrysts, and that the magma so generated then intrudes at a higher level (whose ambient temperatures are consistent with low-temperature megacrysts), thus generating a horizontal temperature gradient at that level. The wide temperature and chemical range of the megacrysts result at this higher level from cooling and differentiation of the magma, with possible variations in cooling rate as a result of localized magma movement and injection. Pegmatite bodies may form and later yield some megacrysts to kimberlite on disaggregation. There is some evidence (Haggerty & Boyd 1975; Jakob 1977; Rawlinson & Dawson 1977; Gurney *et al.* 1979; Pasteris *et al.* 1979) that the late-stage liquid generated by differentiation of the megacryst magma is kimberlitic; thus the kimberlite carrying the megacrysts and other nodules to the surface may also originate in some cases from the same magma that initially gives rise to the megacryst suite.

It is further suggested that the deformed high-temperature peridotites (category V, table 1) form the immediate envelope to the intruded megacryst magma and that they are formed from other peridotite types, with their deformed and high-temperature features being a product of the magma intrusion. The distinctive chemical and less depleted characteristics of category V peridotites (§2, figure 2) may result from metasomatic exchange between the magma and its wallrock, the latter process being analogous to that described for some peridotites (see §3*b* and figure 2) adjacent to intruded clinopyroxene-rich sheets (Harte *et al.* 1977). Metasomatism of this nature might also explain some of the inconsistencies in treating the common peridotites (category I, table 1) as residua derived by melting the category V peridotites (Shimizu 1975*a*, *b*; Shimizu & Allègre 1978 – these authors refer to the equivalent peridotite types as groups II and I). Deformation may also play some part in generating the chemical characteristics of category V peridotites (Boullier & Nicolas 1973, p. 66).

Our alternative upper mantle model is schematically summarized in figure 4*b*, in which the sites of magmatic intrusions are those of generation of megacrysts and category V peridotites (note that the depth temperatures shown on figure 4 do not include the increased temperature associated with the magmatic intrusions). In addition to accounting for the coincidence of high temperatures, deformation and particular chemical characteristics in the category V peridotites, the proposed wallrock origin for these rocks may explain the occurrence of isotopic disequilibrium within them (Allègre *et al.* 1978), and the tendency of the *T*–*P* estimates for category V peridotites from specific localities to cluster tightly and separately (§2, table 1) from other peridotites. The other categories of peridotites (table 1) are believed to be derived from outwith the aureoles of the magma intrusions at the same or higher mantle levels, and from a depth zone (on a normal geotherm) coincident with their *T*–*P* estimates and into which the (megacryst) magma intrusions penetrate. A tendency for these intrusions to penetrate to a similar level may result from the variation of thermal and mechanical properties across the thermal boundary layer between lithosphere and asthenosphere (McKenzie & Weiss 1975).

Clearly, in our model, the *T*–*P* profiles (and inflexions therein) constructed by using megacrysts and category V peridotites cannot be regarded as geotherms in any normal sense. At the same time, the inflected profiles are a product of heat convection in this model as they are in the models of Green & Geuguen (1974), Parmentier & Turcotte (1974), and more recently Boyd & Nixon (1978). In our model, the convection is considered as involving relatively discrete bodies carrying substantial liquid, though the melting giving rise to these bodies at depth may be caused by convection involving essentially solid diapirs or plumes. Clearly a range of possibilities with variable amounts of solid and liquid in the convective agency and variations in the overall scale of the process are possible. Our model emphasizes local magmatic phenomena without the creation of widespread regional depth zones. Thus in some respects figure 4*a* and *b* illustrate extreme viewpoints in interpretation of the peridotite/megacryst nodules from kimberlite. Resolution and mediation between these viewpoints depends upon obtaining accurate estimates of T and P for nodules from individual pipes, though the possibility of variations in this respect between individual pipes must also be expected.

An argument may be made for the sampling of mantle material as nodules at greater depths than indicated in our model (figure 4*b*) on the basis of the high temperatures indicated by some eclogites and eclogitic inclusions in diamonds (figure 3). At present, our inclination is to interpret these high temperatures as reflecting relict temperatures associated with previous magmatic events in which the eclogitic material was initially formed. That such relict temperatures may be preserved is shown by the eclogite nodules themselves – see §2 and the compositions found in single eclogite xenoliths (see, for example, Harte & Gurney 1975*b*).

It is therefore suggested that the nodules in kimberlite provide little direct evidence of mantle petrology below the ornamented zone shown in figure 4*b*. Within this zone, nodule abundances and other characteristics (§2) indicate that coarse common peridotite (category I) is the principal rock type and that within it occur bodies of eclogites–grospydites and pyroxenites (together with high FeO/MgO coarse peridotites). These bodies are schematically shown as 'pips' (figure 4*b*) because the nodule T–P estimates do not suggest any consistent gross layered distribution of these rock types. The size of these bodies is highly uncertain and may well vary considerably; the marked variations in nodule proportions between adjacent kimberlite pipes may indicate that their lateral extension is usually less than a few tens of kilometres. At the lower limit, the homogeneity in mineral compositions in individual peridotite and pyroxenite xenoliths even when layered (Gurney *et al.* 1975; Harte *et al.* 1977) must indicate that pyroxenite and high FeO/MgO peridotite bodies in common peridotite are distinctly larger than xenolith sizes of about 30 cm. Although no gross lithological layering is shown in figure 4*b*, the detailed mineralogical characteristics of all rock types must vary as functions of depth.

5. CONCLUSIONS

Nodules from southern African kimberlites indicate that the mantle from which they derive consists predominantly of peridotite together with some eclogite (including grospydite) and pyroxenite. Marked chemical and textural variations occur within these rock types. To these heterogeneities may be added those developed by the occurrence of mantle metasomatism in which Ti, Fe, K, and other incompatible elements are introduced into mantle rocks. These metasomatic heterogeneities are seen developed on a very small scale in the mantle nodules, but their presence testifies to the existence of fluids (possibly derived from magmas) and the occurrence of processes that create marked chemical heterogeneities. The strong segregation of many elements between minerals and silicate liquid (magma) is manifest and is illustrated by the chemical nature of kimberlite itself. If an origin and evolution for the continental mantle lithosphere involving magmatism (almost certainly with some associated peripheral metasomatism if there is any H_2O or CO_2 involved) is accepted, then the existence of large-scale as well as small-scale heterogeneities may be expected from the basic nature of the processes involved. These factors also make it extremely unlikely that any homogenous rock sample from kimberlite represents the average upper mantle chemical composition.

The distribution of nodule types across the Kaapvaal craton indicates that similar rock types occur throughout the underlying mantle sampled by the kimberlites, but there are extensive local variations in the proportions of different nodule types. It is suggested that the present evidence may be interpreted to indicate extensive sampling of the mantle by kimberlites over only a restricted depth zone and that there is little evidence of any vertical or depth layering of principal rock types within the zone sampled. High-temperature peridotites and megacrysts are interpreted to be associated with discrete (magma) intrusions into this depth zone of common sampling.

We wish to thank Dr F. R. Boyd for helpful consultation and for critically reading the manuscript.

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