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Kimberlite nodules, upper mantle petrology, and geotherms

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The types of nodules erupted in kimberlite and believed to be of upper mantle origin are divided into five types: (*a*) peridotites and dunites, (*b*) garnet–pyroxenites, (*c*) eclogites and grospydites, (*d*) megacrysts, (*e*) amphibole-bearing and mica-rich types. The value of *relative* pressure–temperature estimates in determining the conditions of formation of the nodules is noted and such estimates are used to discuss the distribution of rock types and minerals in the upper mantle sampled by kimberlite. Emphasis is placed on garnet–lherzolites and it is found that there is little evidence for the restriction of deformed or undepleted varieties of garnet–lherzolite to particular depths. Phlogopite, apparently in equilibrium with the host garnet–lherzolite assemblage, appears to be restricted to rocks formed in the lower range of pressure and temperature estimates (less than about 1150 °C). Ilmenite-bearing and other megacrysts of relatively high Fe and Ti and low Cr types appear to have pressure–temperature estimates in the intermediate and higher part of the garnet–lherzolite range. The distribution of phlogopite and megacrysts may be related to melting processes. Present information does not suggest extensive changes in the pressure and temperature of formation of nodules as a consequence of diapiric or other activity associated with kimberlite genesis.

The pressure–temperature gradients of garnet–lherzolites from *individual* pipes do not show any inflexions, but appear to be straight or gently curved. Allowing for the uncertainties in absolute pressure and temperature estimates, the pressure–temperature ranges and gradients of nodule suites are roughly in accord with geotherms based on geophysical data for an upper mantle with a convective circulation.

1. Introduction

The ultramafic nodules carried up to the Earth's surface in magmas (principally of alkali basaltic and kimberlitic type) have long been recognized as important clues to the nature of the Earth's upper mantle and mantle processes (see, for example, Wyllie 1971*a*, chapter 6). Estimates of the temperature and pressure of formation of the nodules prior to eruption provide valuable data in unravelling the original position of the nodules within the upper mantle and reconstructing petrological and physical conditions therein (Boyd 1973). It is the objective of this paper to review these features with respect to the nodules erupted in kimberlite.

Excluding xenoliths believed to originate in the Earth's crust, five general categories of nodules from kimberlite may be recognized as detailed below:

(*a*) *Peridotites and dunites*. The most frequently occurring nodules in kimberlite are olivine-rich rocks usually containing substantial orthopyroxene and with or without some clinopyroxene and garnet. These peridotites are thus classified as harzburgites (olivine–orthopyroxene) and garnet–harzburgites (olivine–orthopyroxene–garnet) and their clinopyroxene-bearing equivalents the lherzolites and garnet–lherzolites. Dunites occur, but are not found abundantly. The most common minor mineral which may be additionally present in all these rock types is a chrome-rich spinel; but primary phlogopite, ilmenite, sulphides, rutile, graphite and diamond (in approximately decreasing order of occurrence) are also known.

(*b*) *Garnet–pyroxenites*. These rocks (also named garnet–websterites) consist essentially of orthopyroxene, clinopyroxene and garnet. Their mineral chemistry is generally related to that of the peridotites and transitional types with variable amounts of olivine link the two categories. Garnet–pyroxenites have not been commonly found.

(*c*) *Eclogites and grospydites*. These nodule types, consisting essentially of clinopyroxene and garnet and with transitional relations to one another, are not as common as the peridotites, but are widely found and exceptionally abundant in some kimberlites. Possible minor minerals in these rocks include: kyanite, corundum, rutile, ilmenite, quartz, orthopyroxene, diamond, graphite, coesite and potash felspar (phlogopite is not uncommon but usually appears to be secondary).

(*d*) *Megacrysts* (*discrete nodules*). These nodules consist predominantly of large (> 1.0 cm) single crystals of clinopyroxene, orthopyroxene, garnet, olivine, ilmenite or phlogopite; these large crystals sometimes carry smaller inclusions of one another. Included in this group are certain well known and occasionally abundant regular intergrowths of clinopyroxene with ilmenite and rarer ones of orthopyroxene with ilmenite.

(*e*) *Amphibole-bearing and mica-rich types*. In addition to the above well characterized groups of nodules, are a diverse collection of polycrystalline types (including nodules dsecribed as glimmerites) which contain amphiboles or are rich in biotite. The amphiboles reported are both hornblendic (Johnston 1973) and richteritic (Erlank 1973, 1976; Dawson & Smith 1977), and the nodules may be related (see, for example, Johnston 1973; Erlank 1976) or largely unrelated (e.g. the Marid suite of Dawson & Smith 1977) to the peridotites. The placing of these rocks in a single category here reflects uncertainty concerning their interrelations, distribution and occurrence, rather than any belief in a common petrogenesis.

Initially it may be noted that the immediate, pre-eruptive, history of many peridotites and garnet–pyroxenites is clearly metamorphic rather than magmatic in that they show evidence of penetrative plastic deformation. In less deformed varieties a metamorphic history is occasionally indicated by evidence of coarse exsolution textures. Such textures are also seen in the eclogite–grospydite xenoliths occasionally, and extensive granular exsolution might be inferred to have occurred commonly in this nodule type (Lappin & Dawson 1975; Harte & Gurney 1975*b*). These nodules therefore appear to be of xenolithic rather than cognate origin with respect to kimberlite, and the imprint of the kimberlite magmatism and eruption (as distinct from any diapiric activity associated with the kimberlite) on the nodules appears to be generally small. Characterization of any *original* magmatic origin needs to penetrate the screen of metamorphic features noted above, and must rely on bulk compositional rather than textural features (Gurney, Harte & Cox 1975). With respect to the nodule categories (*d*) and (*e*) listed above, a more direct relation with kimberlite or other magma may commonly exist (Nixon & Boyd 1973*b*; Dawson & Smith 1977).

Pressure and temperature limits of formation of nodules may be set by experimental determination of the stability fields of particular minerals and mineral assemblages (e.g. graphite/diamond; phlogopite; spinel–peridotite/garnet–peridotite) but continuous assessment of pressure and temperature depends on the use of continuously varying mineral compositions in particular assemblages. The most useful assemblage from the viewpoint of sensitivity to the physical parameters coupled with available experimental information is clinopyroxene–orthopyroxene–garnet. Boyd (1973) demonstrated the value of knowledge of pressures and temperatures of formation, which he based on the temperature dependence of the clinopyroxene limb of the clinopyroxene–orthopyroxene solvus and the temperature and pressure dependence of Al_2O_3 in orthopyroxene coexisting with garnet. Although considerable uncertainty remains concerning the experimental calibration of these equilibria and their application to natural compositions (Powell 1978, this volume; Howells & O'Hara 1978, this volume), they will be used in this paper to provide a basis of estimation of relative pressures and temperatures. These latter, for ease of comparison with other published data, have been calculated using the Ca/(Ca+Mg) ratio of the clinopyroxene (Davis & Boyd 1966, experimental data) and the raw Al_2O_3 content (MacGregor 1974, experimental data) of the orthopyroxene (Boyd 1973). Discussion of the likely error in these values is deferred to §3. In §2 their use is essentially relative, and providing that reasonable care over bulk compositions is exercised the *relative* temperatures and pressures are unlikely to be incorrect.

The above equilibria are applicable to the mineral assemblages of the garnet–lherzolites and garnet–pyroxenites in the nodule categories (*a*) and (*b*) above. They may also be applied to the megacrysts (*d*), provided one accepts that the megacrysts were originally in aggregates containing two pyroxenes and garnet (Nixon & Boyd 1973*b*).

Relative temperature estimates may also be made for the eclogites and grospydites using the distribution of Fe and Mg between clinopyroxene and garnet, but experimental calibration is restricted to particular compositions (Hensen 1973; Akella & Boyd 1974; Raheim & Green 1974) which do not correspond to those of the commoner eclogite and grospydite nodules. Pressure estimates for these rocks depend upon the presence of graphite or diamond, or rely upon calibration using temperature and an estimated geotherm (McCallum & Eggler 1976).

In §2 the nodule evidence for petrological variation in the upper mantle as a function of increasing pressure and temperature (and depth) is reviewed; while in §3 the dT/dP gradients shown by the nodules will be compared with geotherms based on geophysical data. Clearly, the importance of knowing the absolute values of pressure and temperature or at least the likely errors in estimated values, is much more critical with respect to §3 than §2.

The dT/dP gradients derived from nodule studies are often termed 'geotherms'. While this name is acceptable, it does often tend to prejudice the view that the dT/dP gradients represent a widespread and long time-period depth section of the upper mantle; and that, despite obvious selective sampling, the kimberlite nodules may be treated like random fragments from a drill core (cf. Boyd & Nixon 1975, p. 432; MacGregor 1975, p. 455; Dawson, Gurney & Lawless 1975, p. 300; Irving 1976, p. 640 *et seq.*). Similar circumstances surround the use of the word 'depth' and discussion of changes with depth by the author is not intended to imply the absence of horizontal petrological variations or influence of diapirism.

2. Petrological characteristics and their variation with pressure and temperature of origin

An elegant and influential summary and interpretation of a large number of data for the kimberlite nodules of Northern Lesotho was presented in a series of papers by Boyd (1973), Boyd & Nixon (1973), Nixon & Boyd (1973*a*, *b*) and Nixon, Boyd & Boullier (1973). A summary and amplification of their data and viewpoints is given in Boyd & Nixon (1975). The work of these authors forms a valuable starting point in the present discussion. The distribution indicated by estimated pressures and temperatures of origin of various petrological features is discussed in turn below with emphasis upon the implications for the Boyd–Nixon model of data available from 1973 onwards. Most of the data discussed come from southern African kimberlites.

(a) *The garnet–lherzolites*

General distribution

These rocks are both relatively abundant and amenable to temperature and pressure estimation by the methods presently available. In terms of temperatures and pressures (calculated in the simple manner described above) these nodules occur with pressure–temperature coordinates ranging approximately from 900 °C at 40–65 kbar† to 1250–1450 °C at 70 kbar. Their frequency of occurrence in this spectrum appears markedly uneven. Lower pressure and temperature varieties (temperature range 900–1100 °C with corresponding depths of approximately 125–165 km) are widely abundant. Garnet lherzolites with higher pressure and temperature estimates are more restricted in occurrence and usually fall in the range 1200–1450 °C with corresponding depths of 185–220 km (figure 2). The tendency for garnet–lherzolites to show two groupings of pressure–temperature estimates is often more pronounced if individual pipes are considered and was well demonstrated for the northern Lesotho pipes by Boyd & Nixon (1975). Similar features have been found for other pipes (Johnston 1973; Boyd 1974; Danchin & Boyd 1976) though a high temperature group has not been found at all pipes (Gurney *et al.* 1975; Dawson *et al.* 1975; Boyd & Nixon 1976).

Bulk compositional variations

The majority of garnet–lherzolites are rich in olivine and orthopyroxene and poor in clinopyroxene and garnet, and have low bulk rock Al_2O_3, FeO, TiO_2, CaO, Na_2O, and K_2O (e.g. the 'common peridotites' or 'CP' of Cox, Gurney & Harte 1973, and Gurney *et al.* 1975). Similar rocks were described as 'depleted' by Boyd (1973) and Nixon & Boyd (1973*a*) on the basis of bulk rock and mineral compositions. These authors found that depleted peridotites formed their lower pressure–temperature group (as described above) of garnet–lherzolites; while the higher pressure–temperature group were characterized by less depleted types. Data from other pipes do not show this simple pattern. The Matsoku pipe (Cox *et al.* 1973; Gurney *et al.* 1975), in northern Lesotho, yields garnet–lherzolites with a closely restricted range of relatively low pressure–temperature estimates and a compositional range outreaching the total variation found by Boyd & Nixon. Data from the Frank Smith pipe (Boyd 1974, p. 288) suggest the possibility of wider compositional variation amongst high pressure and temperature nodules. In other cases (Danchin & Boyd 1976), mineralogical compositions indicate the same pattern as found in some pipes in northern Lesotho by Boyd & Nixon (1976).

Considering that the garnet–pyroxenites generally show low pressure–temperature estimates (see below) it appears that there is probably much more diversity in bulk rock compositions coming from the lesser depth range than the greater depth range. Boullier & Nicolas (1973, p. 66) have suggested that the tendency towards overall less-depleted compositions in the higher pressure–temperature xenoliths is a result of mechanical mixing of different rock types, during deformation. Clearly, more data, particularly of a modal and bulk rock compositional nature are needed in order to establish any 'stratification' of garnet–lherzolite rock compositions.

Deformation

A considerable amount of information has accumulated in recent years concerning the deformational textures of ultramafic nodules erupted in kimberlite and basalt (e.g. Boyd 1973;

† 1 kbar = 10^8 Pa.

Nixon & Boyd 1973*a*; Boullier & Nicolas 1973, 1975; Cox *et al.* 1973; Dawson *et al.* 1975; Harte, Cox & Gurney 1975; Mercier & Nicolas 1975). A variety of names have been applied to the various textural types; these are summarized in Harte (1977) and the names used here will be those proposed in the same publication.

Boyd (1973) and Nixon & Boyd (1973*a*) drew a distinction between little deformed ('granular') and strongly deformed ('sheared') types. Their first type essentially corresponds to the *coarse* textural type of the presently adopted classification; while their second type is equivalent to the *porphyroclastic* and *mosaic–porphyroclastic* textural types. Boyd & Nixon (1976) found that the lower pressure–temperature garnet–lherzolites were coarse while the higher pressure–temperature lherzolites were always porphyroclastic or mosaic–porphyroclastic. Nixon *et al.* (1973) and Boyd & Nixon (1975) proposed a progressive increase in degree of deformation with depth. Other data have shown that this straightforward correlation between textural type and depth of origin is not substantiated. In particular, xenoliths of relatively low temperature and pressure origin from Matsoku (Cox *et al.* 1973; Harte *et al.* 1975), Kimberley (Dawson *et al.* 1975; Boyd & Nixon 1976) and Colorado–Wyoming (McCallum & Eggler 1976) show an extensive range of coarse, porphyroclastic and mosaic–porphyroclastic types. However, most data support the concept that high pressure–temperature nodules are porphyroclastic or mosaic–porphyroclastic, although nodules from the Frank Smith kimberlite include some relatively little-deformed types (Boyd 1974).

The characteristics of the deformed xenoliths strongly indicate a dominant power law creep process (see summary in Carter 1976, pp. 351–353). The extent to which this results from pervasive mantle creep rather than some transient diapiric activity associated with kimberlite formation is difficult to assess (Boullier & Nicolas 1975, pp. 472–475; Harte *et al.* 1975, pp. 490–494). The limited recovery, recrystallization and grain growth shown by some specimens suggests extrusion rapidly following the deformation. Goetze (1975) has argued on the basis of dislocation studies that some mosaic–porphyroclastic xenoliths show high-stress and strain rate deformation which occurred immediately prior to eruption. More generalized evidence, which might support the association of the deformation textures with a transient phenomenon connected with the generation of the kimberlite, is the paucity of granuloblastic nodules and dominant lack of evidence of repetitive cycles of deformation (Harte 1977).

However, these arguments are by no means conclusive or necessarily applicable to all deformed xenoliths. If an upper mantle model involving small-scale convective circulation (McKenzie & Weiss 1975; see also §3) is accepted, the garnet–lherzolite nodules are probably derived from the thermal boundary layer and regions immediately adjacent to it. In this situation the high pressure–temperature nodules may be derived from near the base of the thermal boundary layer and therefore involved in the small scale circulation and thus usually strongly deformed. In the upper part of the thermal boundary layer and in the base of the overlying plate the extent of deformation will be less and more variable as a function of the extent of coupling between the plate and thermal boundary layer (McKenzie & Richter 1976).

(*b*) *Phlogopite and megacrysts*

The occurrence of phlogopite in kimberlite nodules is highly important because of its geochemical implications and the influence of H_2O on melting temperatures (Lambert & Wyllie 1968; Wyllie 1971*b*). Boyd (1973) and Boyd & Nixon (1973) noted a restriction of phlogopite to garnet–lherzolites with relatively low pressures and temperatures of equilibration.

Much of the phlogopite seen in these xenoliths appears to be obviously secondary because of its association with kelyphite, but some appears to be primary (Carswell 1975). Boyd (1973) suggested that phlogopite in peridotites was generally metasomatic because of its association with depleted xenoliths. Cox *et al.* (1973) and Harte *et al.* (1975) noted the presence in Matsoku xenoliths of phlogopite in probable textural equilibrium with other (often recrystallized) silicates and which had often been introduced into the rocks in association with ilmenite, rutile and sulphides. These authors described such phlogopite as 'primary–metasomatic' in distinction to the common 'late–secondary' material. Harte & Gurney (1975*a*) provided electron microprobe data supporting the equilibration of 'primary–metasomatic' phlogopite with the host garnet–peridotite silicates and its formation under the ambient (around 1050 °C) mantle conditions shown by garnet–lherzolites from the Matsoku pipe.

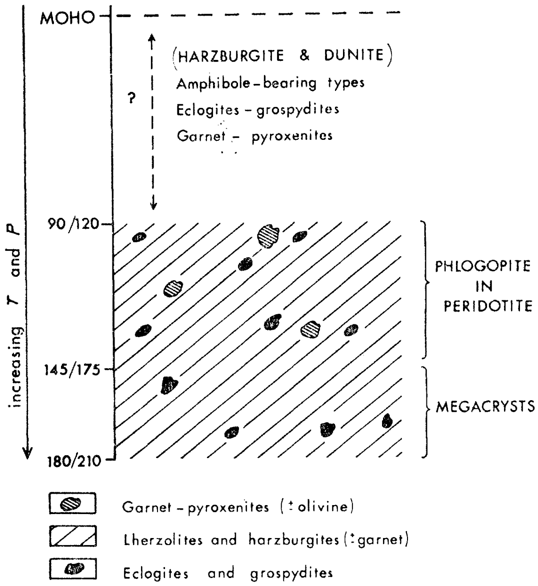

FIGURE 1. Schematic illustration of the distribution of rocks and minerals in the upper mantle as indicated by kimberlite nodules. The numbers give depths equivalent to pressure estimates (see p. 496).

No phlogopite, other than of kelyphitic association (Boyd 1974, p. 288) has been described from high pressure–temperature nodules. Thus, providing we can make the restriction of considering only 'primary' phlogopite, present evidence strongly supports the view that phlogopite only occurs in the lower pressure–temperature part of the broad zone of origin of garnet–lherzolites. Furthermore, such phlogopite appears to be commonly of metasomatic origin.

Boyd & Nixon (1973, pp. 266–268), following Lambert & Wyllie (1968) and Wyllie (1971*b*), suggested that phlogopite would break down to be replaced as an H_2O-bearing phase by silicate melt at greater depths. This transition was linked by Boyd & Nixon (1973) with the upper mantle low-velocity zone and the distribution of megacrysts (discrete nodules), particularly ilmenite-bearing varieties and others relatively rich in Fe and Ti and poor in Cr. Boyd & Nixon (1973) and Nixon & Boyd (1973*b*) suggested that these megacrysts originated as phenocrysts in crystal-mush magmas of the low-velocity zone. The megacrysts give pressures and temperatures of formation, on the basis of certain assumptions (Boyd & Nixon 1973, pp. 262–266; Nixon & Boyd 1973*b*, pp. 73 *et seq.*), in the deeper part of the garnet–lherzolite depth range.

Pressure–temperature estimates therefore support the link between phlogopite and megacrysts via the formation of silicate melt. However, the connection of these processes with the low-velocity zone has been made tenuous because of the lack of geophysical evidence for a possible melting zone beneath shields (Knopoff 1972, p. 512). Eggler & McCallum (1976) have sought to maintain the links between megacrysts and melts by invoking the presence of diapirs within which melting occurs. These interesting suggestions obviously need further clarification with respect to the distribution of different types of megacrysts (e.g. Eggler & McCallum 1976) and the interrelations of nodule pressure–temperature estimates, phlogopite stability limits, peridotite melting temperatures and the abundance of H_2O and CO_2 (see, for example, Boyd 1973, p. 2542; Modreski 1972; Mysen & Boettcher 1975).

(*c*) *Synthesis*

The distributions of minerals and rock types as a function of pressure–temperature estimates based on the assemblage clinopyroxene–orthopyroxene–garnet have now been largely reviewed. Further data bearing upon petrological variation with depth are listed below:

(*a*) Peridotites and eclogites containing either diamond or graphite have been reported (Boyd 1973; Dawson & Smith 1975*b*; McCallum & Eggler 1976; Reid *et al.* 1976); thereby demonstrating a sampling by kimberlite of material on both sides of the graphite–diamond reaction curve (Bundy, Bovenkerk, Strong & Wentorf 1961).

(*b*) The wide spectrum of Fe/Mg distribution coefficients for clinopyroxene–garnet in eclogites (e.g. McCallum & Eggler 1976; Reid *et al.* 1976) suggests the presence of such rocks over a broad depth range, though some consideration must be given to the cooling history of some eclogites–grospydites (Harte & Gurney 1975*b*; Lappin & Dawson 1975).

(*c*) The garnet–pyroxenites generally yield relatively low pressure–temperature estimates (Cox *et al.* 1973; Boyd 1974; McCallum & Eggler 1976).

(*d*) The amphibole-bearing nodules are restricted to the lower part (shallower depth) of the garnet–lherzolite pressure–temperature range (Johnston 1973; Erlank 1976) and even lower temperature environments (Dawson & Smith 1977).

(*e*) Harzburgite nodules are relatively common in kimberlite but their relative pressures and temperatures of origin are uncertain. Some probably come from the same depth range as garnet–lherzolites (see especially Danchin & Boyd 1976). Spinel, when present in the harzburgites, is usually chromiferous and therefore not necessarily associated with shallow depths (Dawson & Smith 1975*a*, p. 347; Boyd & Nixon 1975, p. 442). A lack of garnet–lherzolites with pressure–temperature estimates corresponding to the uppermost part of the upper mantle creates a 'void' which it is tempting to fill with harzburgites and dunites. The occurrence of partial melting with the generation of relatively low density harzburgite and dunite also suggests the abundance of these rock types in the uppermost mantle (O'Hara, Saunders & Mercy 1975, p. 579; Oxburgh & Parmentier 1978, this volume).

The data presented in this section are summarized in a generalized way in figure 1. The numbers at the left-hand margin of figure 1 give two estimates of depth in kilometres; the larger number in each pair corresponds to that found using the basis of pressure–temperature estimation adopted herein (§1) for comparative purposes. The smaller numbers are preferred values (see §3).

Although figure 1 is presented as a depth section, it must be emphasized that the figure is highly stylized and, furthermore, that it cannot be assumed that all the features shown are simple functions of depth in some average subcontinental section. The broad distribution over extensive ranges of pressure and temperature of peridotitic and eclogitic nodules and garnet–pyroxenites must be depth related, though the detailed heterogeneity both vertically and horizontally is uncertain – individual pipes show distinctive features betokening much detailed upper mantle heterogeneity. In other cases the indications of the impact of significant kimberlitic or other localized activity cannot be ignored. Dawson & Smith (1977) note the possible links between certain amphibole-bearing nodules and kimberlite magma, while the presumed magmatic origin of megacrysts perhaps in association with diapirism has also been noted. In addition the metasomatic nature of phlogopite (and some amphibole – see Erlank 1976) may be emphasized and its possible, though admittedly tenuous, links with the kimberlite noted (Harte *et al.* 1975, p. 501). An association of nodule characteristics with the kimberlite and its envelope is strengthened if the deformation of the nodules is viewed as part of a diapiric process related to kimberlite genesis. Thus the significant questions may be such as how the formation of phlogopite is related to temperature and water distribution within a diapir or the kimberlite envelope, rather than at what depth in the mantle widely dispersed phlogopite gives way to silicate melt. Some further aspects of the influence of diapirs are discussed in the following section, but it must be noted here that the available evidence (Boyd & Finger 1975; Boyd 1975; Gurney *et al.* 1975; Harte & Gurney 1975*a*) suggests little if any modification of temperatures and pressures of formation in association with deformation or primary metasomatic processes.

3. Geotherms

(*a*) *Non-inflected geotherms for garnet–lherzolites from individual pipes*

By using the pressure–temperature estimates from a group of kimberlite nodules, a pressure–temperature range and $\mathrm{d}T/\mathrm{d}P$ may be determined which may be presumed to relate to the upper mantle section sampled by the kimberlite(s). In a pioneering study, Boyd (1973) interpreted such estimates for northern Lesotho as displaying a 'fossil geotherm' for the late Cretaceous upper mantle, and as showing an inflection to higher $\mathrm{d}T/\mathrm{d}P$ in the deeper part of the mantle section. Boyd (1973) suggested that the inflection was a product of stress-heating. Various authors have doubted the feasibility of the stress-heating hypothesis and have reinterpreted the inflection using mantle diapir or plume models (Green & Gueguen 1974; Parmentier & Turcotte 1974), or suggested that the inflection is an artefact of the method of calculation (Mercier & Carter 1975, p. 3359) or faulty experimental calibration (Howells & O'Hara 1978, this volume).

In figure 2, pressure–temperature estimates for garnet–lherzolites from a number of kimberlite pipes are plotted. The estimates are those adopted for relative purposes (§ 1) and thus use the principal method adopted by Boyd (1973, figs. 5 and 7A) and Boyd & Nixon (1973). The data for northern Lesotho (figure 2*a*) show that *individual pipes* yield $\mathrm{d}T/\mathrm{d}P$ gradients *without* inflexions and indeed with very little, if any, curvature. The difference in interpretation of essentially the same data points as compared with Boyd (1973, see especially fig. 7*a*) is that the latter author has drawn a $\mathrm{d}T/\mathrm{d}P$ gradient through data points from several pipes and also such that it lies parallel to the Clark & Ringwood (1964) shield geotherm at shallow depths. In figure 2*a* the nodules from Thaba Putsoa indicate an approximately linear $\mathrm{d}T/\mathrm{d}P$ gradient involving higher temperatures at particular pressures than other northern Lesotho pipes. This accords with the results of Mercier & Carter (1975, fig. 9) and is also evident on comparison of figs 1 and 2 of Boyd & Nixon (1975), even though both sets of authors have calculated their data points by other methods.

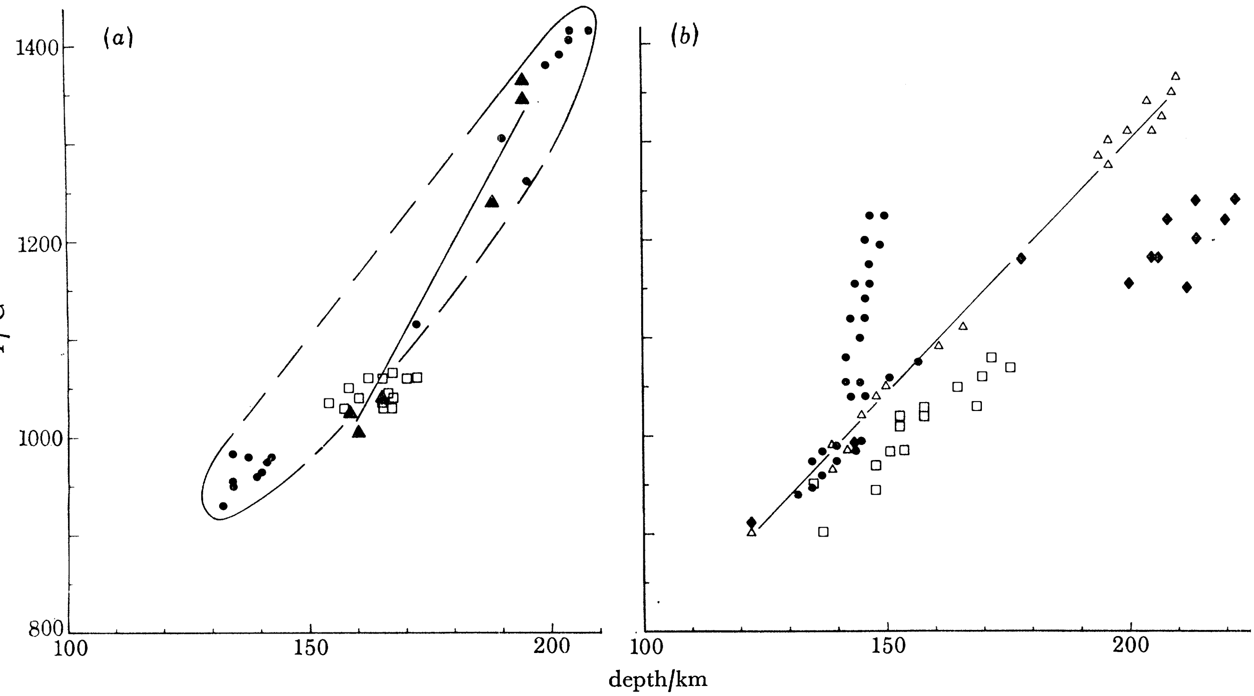

FIGURE 2. Estimated temperatures and pressures (depths) of formation of garnet–lherzolite xenoliths from various kimberlite pipes. (*a*) From northern Lesotho: dots, Thaba Putsoa; triangles, Mothae; squares, Matsoku; the loop surrounds Thaba Putsoa data points and the straight line connects Mothae data points. (*b*) From S & SW Africa and Udachnaya (U.S.S.R.): triangles, Premier (data points connected by straight line); squares, Kimberley; dots, Louwrencia; diamonds, Udachnaya. A tight clustering of data points occurs for Matsoku and highest temperature rocks from Premier and some of these data have been omitted to prevent overcrowding. Data are taken from: Boyd & Nixon (1973*a*, 1975); Boyd, Fujii & Danchin (1976); Cox *et al.* (1973); Danchin & Boyd (1976); Dawson & Smith (1975); Dawson *et al.* (1975); MacGregor (1975).

In figure 2*b* individual pipes indicate linear or gently curved dT/dP slopes. The Louwrencia data pattern is uncertain but the marked inflexion shown by MacGregor (1975) is seen to be absent when the extremely uncertain spinel–lherzolite data points (Wilshire & Jackson 1975) are removed (cf. MacGregor 1975, fig. 4). The other inflected geotherms shown by MacGregor (1975) are based on data from several pipes. With respect to the Udachnaya data (from Boyd, Fujii & Danchin 1976, fig. 12) it must be emphasized that very low Al_2O_3 contents in orthopyroxenes make the absolute pressure estimates even more tentative than usual (Borley 1975, p. 490).

(*b*) *Comparison of 'geophysical' and 'nodule' geotherms*

In comparing the pressure–temperature ranges estimated from nodule suites with geotherms based on geophysical estimates and models, it is important to consider the likely direction of error in the estimates. The pressure–temperature coordinates for the nodules shown in figure 2 may be displaced by using alternative calculation procedures and/or different equilibria and/or different sets of experimental data (see Powell 1978, this volume; Howells & O'Hara 1978, this volume). With respect to temperature estimation, it is noteworthy that more recent determinations of the diopside–enstatite solvus (Lindsley & Dixon 1976) do not substantially modify estimates based on the solvus of Davis & Boyd (1966). The errors involved in the application of such data to natural assemblages are less certain (Wood & Banno 1973; Mercier & Carter 1975; Powell 1978, this volume).

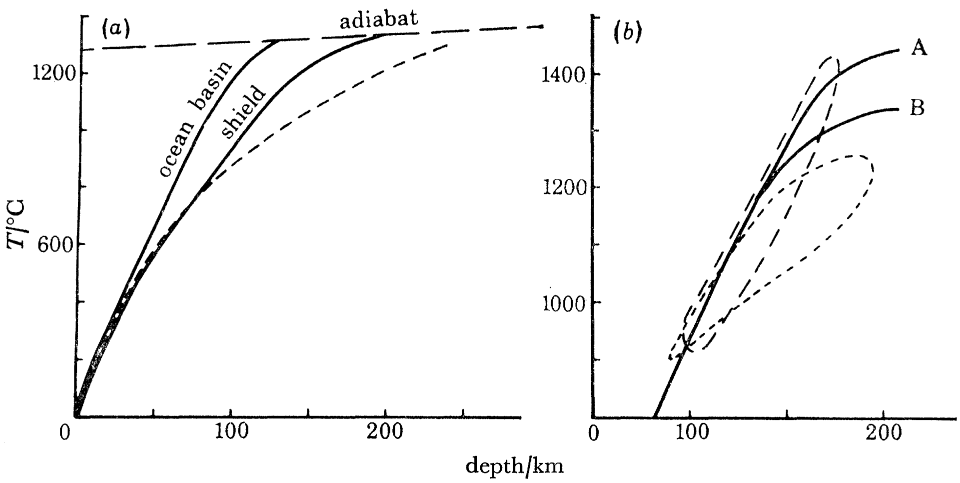

Figure 3. (*a*) Geotherms (solid lines) for the deep ocean basin and continental shield constructed on the assumption of upper mantle convection (see appendix) and compared with the Clark & Ringwood (1964) continental shield geotherm (short dashes). (*b*) The convection-related shield geotherm (B) and the same geotherm with a 100 °C diapiric increment (A) compared with estimated temperature and pressure fields for the Udachnaya (short dashes) and Thaba Putsoa (broken line) kimberlite nodules.

With respect to pressure (or depth), use of the Boyd & England (1964) experimental data increases the estimates of figure 2, but otherwise the evidence favours the lowering of these estimates. Calculations using the Wood & Banno (1973) and Wood (1974) formulations in combination with MacGregor's (1974) experimental data, lower the estimated depths by approximately 20 km. Use of the Al_2O_3 solubility data of Howells & O'Hara (1978) lowers the estimates of figure 2 by perhaps 30–45 km. Calculations using the exchange reactions of Powell (1978) decrease the estimated depths for high-temperature nodules by 60–65 km. The only fresh diamondiferous garnet–peridotite (Dawson & Smith 1975*b*) has a depth of origin, as estimated in figure 2, approximately 30 km in excess of the graphite–diamond reaction curve (Bundy *et al.* 1961; Kennedy & Kennedy 1976). Data for a graphite-bearing peridotite (Boyd 1973; Boyd & Nixon 1975; Danchin & Boyd 1976) suggest that the diamondiferous peridotite has formed at conditions close to the graphite–diamond transition.

In figure 1, two sets of depth estimates are shown. The higher set refer to the data of figure 2, while the lower set are based on a 30 km decrease in the values of figure 2. Such a generalized lowering of the pressure estimates is obviously arbitrary, but is thought to be a more likely indication of the real values given the available information, and is also used below (figure 3).

The dT/dP gradients shown by the kimberlite nodules suites have generally been compared with the shield geotherm calculated using a combination of geophysical and geological data by Clark & Ringwood (1964). Nodule suites indicate temperatures which range from close to this geotherm to values at least 200 K in excess of it. This temperature difference is accentuated if the lower pressure estimates of nodule formation are accepted and presumably must indicate

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the strong influence of mantle diapirs or plumes in the genesis of the nodules if the Clark & Ringwood geotherm is adopted. As shown at the end of §2, the available nodule evidence does not presently favour large-scale temperature perturbations.

The evidence of plate tectonics strongly indicates the presence of a convective circulation in the mantle (McKenzie & Richter 1976) and the existence of such convection seriously modifies the Clark & Ringwood geophysical model (Tozer 1967). Attempts to model the convective circulation (Richter 1973; McKenzie, Roberts & Weiss 1974; McKenzie & Weiss 1975; Richter & Parsons 1975; McKenzie & Richter 1976) indicate the presence of a small-scale circulation in addition to the large-scale circulation determined by plate sizes. Some further observational support for the small-scale circulation is given by recent detailed gravity data (Marsh & Marsh 1976). The constraints of the models of small-scale convective circulation in the upper mantle under both oceans and continents have been used in conjunction with thermal data, largely from Sclater & Francheteau (1970), to construct geotherms for the deep ocean basins and continental shields; see figure 3*a* and the appendix for details of the data used.

The pressure–temperature coordinates of the presently estimated geotherms (figure 3*a*) are subject to variation in the estimated depth and adiabatically controlled temperature at the base of the thermal boundary layer. In addition, variations in heat flow and the geochemical and petrological characteristics of the lithosphere above the thermal boundary layer will affect the position of the geotherms in pressure–temperature space. Within the model of the small-scale convective circulation, diapiric activity is essentially equivalent to regarding the convection as time dependent (McKenzie & Richter 1976); such activity is unlikely to generate temperature perturbations much in excess of 100 K (D. P. McKenzie, personal communication).

In figure 3*b* the convection-related shield geotherm of figure 3*a* is shown together with a similar geotherm in which diapirism is assumed to cause a 100 K temperature increment at the base of the thermal boundary layer (see also Gass, Chapman, Pollack & Thorpe 1978, this volume). These 'geophysical geotherms' are compared with the relatively low-temperature and relatively high-temperature 'nodule geotherms' from Udachnaya (Boyd *et al.* 1976) and Thaba Putsoa (Boyd 1973; Nixon & Boyd 1973*a*). The nodule pressure–temperature fields in figure 3*b* are placed approximately 30 km below their position in figure 2 (see above). The correspondence of the 'geophysical' and 'nodule' geotherms in figure 3*b* is far from exact; but, allowing for the obvious uncertainties in the estimates of both types of geotherm, figure 3*b* does indicate that a fit of both sets of data is possible.

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Appendix

In calculating the convection-dependent geotherms of figure 3*a*, the models of two scales of convection of McKenzie *et al.* (1974), McKenzie & Weiss (1975) and McKenzie & Richter (1976) have been adopted. The small-scale circulation has a thermal boundary layer against the overlying mechanically strong plate, and the lithosphere–asthenosphere junction may be placed at either the top or bottom of this thermal boundary layer depending upon whether a mechanical or thermal definition of this junction is used. The temperature at the base of the thermal boundary layer is adiabatically controlled and may be fixed using estimates of magma temperatures from oceanic ridges. This temperature has been taken as 1275 °C by following the data of Hodges & Bender (1976). The oceanic geotherm in figure 3*a* applies to the deep ocean basins with crust older than about 80–100 Ma. For this case the base of the thermal boundary layer has been fixed at a depth of 130 km and the thermal boundary layer taken as approximately 50 km thick with a 300 °C temperature difference across it (McKenzie *et al.* 1974, §6). In the case of the shield geotherm the base of the thermal boundary layer has been taken to be at a depth of 200 km. An adiabatic gradient of 0.3 °C km^{-1} (Sclater & Francheteau 1970) has been assumed.

At shallower depths than the thermal boundary layer the calculations assume the temperature distribution to be governed by conduction. In making these calculations, uncertainties arise from the lack of detailed geophysical and geochemical information for particular crust–mantle sections. For simplicity, the set of relatively standard and average heat flow (0.046 and 0.044 W m^{-2} for ocean and shield respectively) and geochemical data of Sclater & Francheteau (1970, fig. 11) has been adopted with the exception that all ultrabasic layers have been given a heat contribution of 1.0×10^{-8} W m^{-3}. Reasonable values of thermal conductivity compatible with the thermal boundary conditions were assigned by following Clark (1966, §21) and Kawada (1966). The data used are summarized in table A 1 (the thicknesses of the ultrabasic layers include the thermal boundary layer).

Table A1

layer	thickness/km	heat production 10^{-8} W m^{-3}	thermal conductivity W m^{-1} K^{-1}
shield – upper crust	8	125	2.7
shield – lower crust	32	25	2.5
shield – ultrabasic	160	1	3.5
ocean – crust	5	50	2.0
ocean – ultrabasic	125	1	3.5

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