MINERALOGICAL MAGAZINE

VOLUME 44 NUMBER 333 MARCH 1981

The nature of the upper mantle

HALLIMOND LECTURE TO THE MINERALOGICAL SOCIETY OF GREAT BRITAIN AND IRELAND, 1980

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ABSTRACT. The nature of the upper mantle below the ancient cratonic areas can be deduced by study of the xenolith suites in kimberlites. Studies on the proportions of xenoliths, together with their mineralogy and chemistry, suggest an upper mantle containing an upper harzburgite zone and a lower lherzolite zone, with both of these zones containing chemical and mineralogical variants, together with minor rock types such as glimmerites, MARID-suite rocks, pyroxenites, and eclogites. Isotopic studies of the phases in xenoliths have yielded restricted ranges of values for the isotopic composition of hydrogen, carbon, oxygen, and sulphur which are tentatively identified as the true isotopic values for these elements in the upper mantle. In addition, recent discoveries suggest that diamond may be a primary upper-mantle phase.

The textures and fabrics of the xenoliths indicate that plastic deformation has taken place in connection with the intrusive kimberlite event, and also in earlier events unconnected with the kimberlite event. In addition, brittle fracture has been observed, this fracturing often being accompanied by the filling of the ensuing veins and joints by fluids that have crystallized potassium-, titanium-, and water-rich phases; limited metasomatism of peridotite wall rocks accompanies this vein filling and more widespread pervasive metasomatism may also be present.

Although most upper-mantle rocks are now metamorphic, in some rare instances there are relics of earlier rock types that have not been completely obliterated by subsequent metamorphic events; most of these could be attributed to an igneous origin and in most cases during subsequent metamorphism the original rock types have been subjected to increasing pressures and/or lower temperatures.

Although the source of most materials now present at the earth's surface can be directly attributed to some identifiable source within upper-mantle rocks, no source has yet been identified for CO_2, N, P, and the rare gases.

EVERYTHING happening at the surface of the Earth, including such diverse phenomena as the building of mountain ranges, the formation of ocean basins, volcanic activity, and even changes in sedimentation (with their attendant rise or decline of faunal and floral populations) is a response to events taking place within the upper mantle; none the less, relatively little is known of this important part of the Earth. The speed of seismic waves impose certain constraints on the density of materials within the upper mantle, and indirect evidence for the chemistry of parts of the upper mantle can be deduced from studying the major, trace, and isotope chemistry of volcanic rocks erupted in the oceans; in addition, there are small, inthrust, often highly deformed slices of upper-mantle material within some fold belts. However, the widest range of samples of upper-mantle materials are the xenoliths found in certain basaltic rocks and kimberlites; this paper will concentrate upon the xenoliths brought up in kimberlites and the model proposed later can only be applied to the continental shield areas to which kimberlites are confined.

For the purpose of this paper the Earth's upper mantle is defined as a concentric shell of the Earth lying below the seismically determined Mohorovicic discontinuity and above the transition zone (again seismically determined) at some 300 to 400 km depth.

Kimberlite is a volumetrically insignificant rock type occurring in vents, sills, and dykes of very limited size. None the less, in the context of our investigations into the nature of the upper mantle, kimberlite assumes a special importance. Because it contains diamond, which may be used as an upper-mantle stratigraphic marker, we know that kimberlite originates from depths of at least 150 km, which is deeper than the site of generation of most other igneous rock types; and it follows that kimberlite has had the possibility of carrying out a large amount of upper-mantle sampling during its ascent to the Earth's surface.

The exotic upper-mantle blocks, termed 'xenoliths' are set within the host kimberlite, and in general are rounded or discoidal in shape; often they are in the size range of 10–30 cm along their longest axis, though others in excess of 1 m in length are known. Xenoliths generally comprise not more than 2–5 % of the total bulk rock but in exceptional cases, as at the Matsoku Pipe in Lesotho, may comprise up to 20 % of the bulk rock. Generally, the commonest type of xenolith is peridotite; other rock types such as pyroxenite, eclogite, glimmerite (mica-rich rocks) are considerably less common, though it should be noted that in some kimberlites eclogites may comprise a high proportion of the total xenolith population. In addition, there are nodules of high-Ti garnet, pyroxenes, olivines, and ilmenites that are known as the 'megacryst suite'; they are interpreted as high-pressure phenocrysts precipitating from the kimberlite magma, and, since they sometimes form composite nodules, they will be described below together with the other upper mantle rock types. For the sake of convenience in ascertaining the nature of the upper mantle, we can consider (*a*) the physical make-up of the mantle, deduced from the proportions of xenoliths, and their mineralogy and chemistry; and (*b*) upper mantle processes, deduced from the fabrics and textures of the xenoliths.

The physical make-up of the upper mantle

Peridotites

As stated above, the commonest type of upper-mantle xenoliths in kimberlite are rocks comprising more than 50 % olivine. For convenience, these can be broken down into lherzolites, containing clinopyroxene and orthopyroxene in addition to olivine; and the harzburgites which do not contain clinopyroxene. It should be noted, however, that many of the so-called 'lherzolites' in kimberlite do not contain the 5 % of modal clinopyroxene required for them to be termed 'lherzolite' under the IUGS scheme of nomenclature of ultrabasic rocks; however, the term is so well entrenched in the kimberlite and upper-mantle literature that the name will be retained here.

Lherzolites. Lherzolites are one of the major xenolith groups and comprise forsterite, enstatite, chrome-diopside, and often, pyrope garnet. Yet within this major grouping there are considerable variations. The modal percentage of garnet, clinopyroxene, and orthopyroxene vary widely (Table I) and, in some specimens whose bulk composition contains very little alumina, garnet may be absent. Although aluminous spinel is known in reaction coronas around garnet, lherzolite specimens containing primary aluminous spinel are exceedingly rare (Lock and Dawson, 1980). The over-all fabrics of the lherzolites are also highly variable. Most commonly the fabric is coarse (terminology of Harte, 1977), with equant crystals and even distribution of the constituent phases throughout the rock; however, other specimens exhibit strong mineral banding and others show deformation fabrics which will be discussed more thoroughly below. In addition to these variations in phase distribution, phase size, and over-all rock fabric, it is also apparent that there is considerable variation in the chemistry of individual phases from specimen to specimen. In two papers reviewing the chemistry of garnets and pyroxenes in kimberlites, xenoliths, and diamond inclusions, Dawson and Stephens (1975) and Stephens and Dawson (1977) recognized twelve chemically distinct groups of garnet on the basis of their TiO_2, Cr_2O_3, FeO, CaO, and MgO contents; and, on the basis of their TiO_2, Al_2O_3, Cr_2O_3, FeO, MgO, CaO, and Na_2O contents, five groups of orthopyroxene and ten groups of clinopyroxene. Using these groupings it can be shown that a range of garnet lherzolites contains five different types of garnet (titanian pyrope, high-Ti pyrope, chrome pyrope, and titanian uvarovite-pyrope), five different kinds of orthopyroxene (enstatite, Cr-Al enstatite, Na-Ca enstatite, Ti-enstatite, high-Ti bronzite) and five different types of clinopyroxene (sub-calcic diopside, diopside, Ti-Cr diopside, chrome-diopside, and ureyitic diopside), although there is a preponderance of 'normal' specimens containing the combination enstatite, chrome diopside, and chrome pyrope. A suite of specimens containing sub-calcic diopside, Ti-enstatite or high-Ti bronzite, and titanian

TABLE I. *Modes of rocks in the garnet-lherzolite–garnet-pyroxenite group*

	1	2	3	4	5	6
Olivine +serpentine	63 (55–83)	66 (50–75)	20	16	8	—
Orthopyroxene	26 (7–36)	25 (12–34)	25	33	52	4
Clinopyroxene	4 (0.2–15.5)	3 (1–17)	35	18	16	81
Garnet +kelyphite	6 (0.9–15.9)	6 (1–15)	20	31	24	15
*Others	1 (0.2–6.7)	—	—	—	—	—

* Phlogopite, rutile, opaques.

1. Average and range of 10 garnet lherzolites, Lesotho (Dawson, 1960).
2. Average and range of 23 South African garnet lherzolites (Mathias *et al.*, 1970).
3. Garnet-websterite, Dutoitspan (Nixon *et al.*, 1963).
4. Garnet-websterite, Matsoku, Lesotho (Mathias and Rickwood, 1968).
5. Garnet-websterite, BD1160, Bultfontein (Dawson, unpubl.).
6. Garnet-pyroxenite, BD1755, Klipfontein (Dawson, unpubl.).

TABLE II. *Summary of chemical differences in the phases in the two groups of Thaba Putsoa lherzolite xenoliths*

	Granular 'depleted'	Sheared 'fertile'
Olivine		
% Fo	93	90.5
CaO wt%	<0.03	0.09
Al_2O_3 wt%	<0.03	0.07
Orthopyroxene		
% En	93.6	90.3
TiO_2 wt%	0.05	0.19
Na_2O wt%	0.07	0.31
Clinopyroxene		
Ca/(Ca+Mg)	0.46	0.35
TiO_2 wt%	0.05	0.25
Cr_2O_3 wt%	1.5	0.9
FeO wt%	2.0	5.2
Garnet		
% Pyrope	~ 70	~ 70
TiO_2 wt%	0.17	0.58
Cr_2O_3 wt%	3.47	1.34
Summary		Relatively hot, high in FeO and TiO_2, low in Cr_2O_3

pyrope have been reported, together with the normal kind of garnet lherzolite, from the Thaba Putsoa pipe, Lesotho, where there is an apparent relationship between the texture of the rock and the over-all mineralogy (Table II); the high-Ti and Fe-rich 'fertile' lherzolites being sheared whereas the normal 'depleted' lherzolites are granular and show no signs of deformation (Nixon and Boyd, 1973*a*). The correlation between apparent fertility (in terms of higher Fe and Ti) and deformation seen in the Thaba Putsoa specimens is not consistently found in nodule suites from other localities; for example, both deformed and undeformed xenoliths in the Bultfontein suite show no Fe and Ti enhancement whereas, on the other hand, undeformed nodules containing enhanced amounts of Fe and Ti are found in the nodule suites at Matsoku (Gurney and Harte, 1980) and at the Letseng pipe both in Lesotho (Bishop *et al.*, 1978). In terms of pyroxene geothermometry, the low Ca/(Ca+Mg) ratio of the clinopyroxenes in the deformed fertile Thaba Putsoa xenoliths indicates that they equilibrated at relatively high temperatures, compared with the normal depleted lherzolites; *if* these relatively high temperatures are correlated with increasing depth, then these fertile peridotites equilibrated in the depth range 150-70 km compared with a depth range of 120-40 for the non-deformed depleted lherzolites. In addition, density calculations indicate that the Fe-rich lherzolites are denser than the depleted ones and theoretically, in a steady-state situation, should be found at greater depths than the depleted ones (Jordan, 1979). Hence, there is a possibility that, within the garnet lherzolite zone of the upper mantle extending from depths of 120 km down to approximately 200 km, there is a density and chemical stratification.

Additional less-common primary phases which have been found in garnet lherzolites are phlogopite, sulphides, carbonate, and chrome-rich spinel, the presence or absence of which will impose small-scale inhomogeneites upon the larger-scale inhomogeneity outlined above. Primary phlogopite has been found in diamond-bearing garnet lherzolite (Dawson and Smith, 1975*a*) and so must persist to depths of at least 150 km.

Harzburgites. Until fairly recently harzburgites have been relatively neglected rocks, although they can form a high proportion of the nodule suite at many localities. Current methods of geobarometry and geothermometry cannot be applied to them, although on density grounds it has been assumed that they form a zone of depleted mantle between the garnet lherzolite zone and the Moho (for example, the model of Clarke and Ringwood, 1964). However, the harzburgites show a considerable textural variation ranging from relatively fine-grained varieties, in which the phases are no greater than 1 mm in size, to relatively coarse-grained ones in which crystals may be in the size range of 10-30 mm; moreover, whereas the finer-grained ones tend to have a very even grain size, in the coarser ones large crystals of orthopyroxene tend to be set in a relatively fine matrix of olivine. In their study on kimberlite and xenolith pyroxenes Stephens and Dawson (1977) recognize that most harzburgites contain low alumina enstatite, but that a small number contain enstatites that are relatively high in chromium, aluminium, and lime, thereby containing appreciable garnet and diopside in solid solution. Subsequent work has shown the picture to be even more complex and Dawson *et al.* (1980) recognize (*a*) barren harzburgites, i.e. ones in which the orthopyroxenes contain less than 1 % Al_2O_3, ~ 0.3 % Cr_2O_3, and virtually no CaO and Na_2O; (*b*) fertile harzburgites in which orthopyroxenes contain ~ 3 % Al_2O_3, ~ 0.8 % Cr_2O_3, and ~ 0.8 % CaO but no Na_2O; and (*c*) high-sodium fertile harzburgites, in which the orthopyroxenes have ~ 0.2 wt% Na_2O in addition to Al_2O_3, Cr_2O_3, and CaO contents similar to those in fertile harzburgite. In addition, there are some rare harzburgites that contain amphibole (chrome pargasite)

(Dawson and Smith, 1975*b*), and a relatively few aluminous spinel harzburgites are also known (Smith and Levy, 1976). Hence, recent studies are showing that, within the upper harzburgite zone of the upper mantle, which was formerly regarded as having a rather monotonous mineralogy, considerable textural, mineralogical, and chemical variations are present.

Pyroxenites

These contain less than 50% of modal olivine, the other main constituents being orthopyroxene, clinopyroxene, and, in some cases, garnet; the commonest accessory phase is a chrome-rich spinel. The group embraces olivine pyroxenites and websterites, both garnet-bearing and garnet-free. Some olivine pyroxenites consist of phases that are compositionally similar to those found in the common garnet lherzolites, and hence may be regarded as modal variants of the garnet lherzolites (Table I). However, within the xenolith suite at the Matsoku pipe, Lesotho, are garnet-olivine pyroxenites containing phases consistently richer in Fe than those in the common garnet lherzolites; these have been interpreted as upper-mantle fluids, derived by partial melting of garnet lherzolite, that did not migrate very far from their site of origin and consolidated under approximately the same *PT* conditions as the common garnet lherzolites (Cox *et al.*, 1973; Gurney *et al.*, 1975). In addition to the individual phases being more Fe-rich, the higher modal amounts of pyroxenes and garnets relative to olivine results in these rocks being higher in FeO, CaO, Al_2O_3, Cr_2O_3, and Na_2O than common garnet lherzolites.

Eclogites

Eclogites form only a small proportion of the xenolith population at most kimberlite localities, but in some rare cases, such as at Roberts Victor and Rietfontein (S. Africa), Obnazhennaya (USSR), and Orapa (Botswana), they dominate the xenolith population. Eclogites have received considerable attention particularly from experimental petrologists in connection with the genesis of basalt (e.g. Yoder and Tilley, 1962; O'Hara and Yoder, 1967; Ringwood, 1975) and by seismologists in connection with the nature of the Mohorovicic discontinuity. A word of caution is necessary regarding the nomenclature of eclogites. In the past the word 'eclogite' has been applied to bi-mineralic rocks occurring in the kimberlite pipes; however, recent studies of 'eclogites' have shown some to be modal variants of garnet lherzolite (see Table I) and others that are basic plagioclase-free granulites. Originally, Beck (1899) introduced the word 'griquaite' to distinguish the eclogites occurring in kimberlites from those occurring in fold-mountain belts and regionally metamorphosed terrains; he believed the eclogites to be composed of minerals that had precipitated from the kimberlite magma. The term 'griquaite' has recently been reintroduced into the literature to describe garnet-pyroxene rocks which are believed to have been aggregates of phases precipitating from the kimberlite magma (Nixon and Boyd, 1973*b*); however, these phases are compositionally unlike those found in most eclogites.

Eclogites are rocks consisting essentially of pyrope-almandine-grossular garnet and diopside-jadeite (omphacite) clinopyroxene. Additional minor or accessory phases give us distinctive eclogite types such as kyanite eclogite, corundum eclogite, coesite eclogite, and carbonaceous eclogite (containing diamond and/or graphite); grospydite is a variant of kyanite eclogite, the difference being that the garnet contains > 50% of calcium garnet molecule. Common accessory phases in all types are rutile and sulphides; and primary amphibole and mica have been reported as rare accessories. There is now a tremendous wealth of data on the chemistry of phases in eclogites and their variants and it is impossible to review all the data; fortunately the range of compositions in the eclogites of the Roberts Victor pipe, South Africa, covers almost the entire range of eclogite compositions, so it is sufficient to discuss the work carried out on the eclogites from this pipe with only occasional reference to other localities to underline certain points. Dawson and Stephens (1976) show that eclogites contain garnets of six distinct compositions on the basis of their TiO_2, Cr_2O_3, FeO, MgO, and CaO contents—calcic pyrope almandine, titanian calcic magnesian almandine, magnesian almandine, pyrope-grossular-almandine, and chrome pyrope; most of these are found in eclogites from the Roberts Victor Mine, illustrating the wide range of composition in garnets from eclogites in an individual intrusion; an exception to this are the high-almandine garnets (not found in Roberts Victor) which are found in eclogites in the intrusions of the Colorado Plateau (O'Hara and Mercy, 1966). It is pertinent to note that garnet compositions may vary widely within individual hand-specimens (Chinner and Cornell, 1974; Lappin and Dawson, 1975). In the case of the clinopyroxenes, Stephens and Dawson (1977) distinguish seven chemically distinct varieties of clinopyroxenes in eclogites and grospydites. They vary from chromiferous diopsides (indistinguishable from those in garnet lherzolite) to jadeites and jadeitic diopsides with high sodium and aluminium contents. The clinopyroxenes exhibit only small variations in their Ca:Mg:Fe ratios compared with coexisting garnets, and the main variations are in (Ca + Mg):

(Na + Al) (i.e. variable jadeite to diopside). In particular, the pyroxenes in kyanite eclogites and grospydites are very similar, being characterized by high amounts of CaO, Na_2O, and Al_2O_3 and relatively small amounts of FeO and MgO; a constant feature in these pyroxenes is the presence of excess amounts of Al^{vi} which may be regarded as kyanite in solid solution (Sobolev *et al.*, 1968). In the case of diamond-bearing eclogite the garnets are indistinguishable from those in non-diamond-bearing rocks. Although Sobolev and Lavrentev (1971) reported small amounts of Na in the garnets of diamond-bearing eclogites, this being attributed to high pressures, a later study by Bishop *et al.* (1978) indicates that garnets from a large number of non-diamondiferous eclogites contain equally large amounts of Na. These latter authors suggest that the presence of Na is not necessarily due to the high pressures and can be explained by coupled substitution with P and Ti; there is no need to invoke sixfold coordination of silicon, as postulated by Sobolev and Lavrentev (op. cit.). However, the clinopyroxenes from diamond-bearing eclogites contain little if any Al^{vi}, which contrasts with those in other eclogites, and the Fe is mainly in the ferrous state (Sobolev, 1977).

In the case of their bulk compositions, eclogites can be matched in the broad sense with some of the numerous variants of basalt on the basis of their SiO_2, CaO, MgO, FeO, and Al_2O_3 contents. In addition, the broad chemical equivalence of basalt with eclogite, and high-alumina basalt with kyanite, eclogite, and grospydite is assumed in experimental work examining the basalt→basic granulite→ eclogite transition (Yoder and Tilley, 1962; Green, 1967; and Khitarov *et al.*, 1971). However, it has been pointed out (Dawson, 1980) that most kyanite eclogites and grospydites are both more calcic and more aluminous than even high-alumina basalts. In addition, there are two other major discrepancies—the virtual absence of both K_2O and TiO_2 in eclogites compared with most basalts; although, it is true that some eclogites do contain significant amounts of K_2O, this is in the form of secondary mica and/or amphibole and there is very little K_2O in unaltered eclogite. And, although there are some basaltic rock types that contain low amounts of K, most low-K basalts have higher MgO/CaO and MgO/Al_2O_3 ratios than eclogites. In summary, although on the basis of most oxides, eclogites can be broadly matched with some of the wide variety of basaltic rocks, this is not the case when K_2O and TiO_2 are examined. The reason for this is not clear, though there are at least three possibilities: (*a*) eclogites originally may have crystallized from melts of basaltic composition, but during crystallization pressures were too high to allow formation of potassic phases; this raises the problem of where the incompatible elements migrated and in what form?; (*b*) basalts and eclogites originate from similar melts but during their ascent most basalts become enriched in K and Ti and have undergone some degree of fractionation; or (*c*) eclogites and basalts result from chemically distinct melts perhaps reflecting small differences in the mineralogy of the upper mantle where these melts originate.

Conditions of formation of peridotites, pyroxenites, and eclogites

The pressures and temperatures of formation of rocks of the *peridotite-pyroxenite* suite may be ascertained from the stability fields of individual minerals or assemblages based upon current experimental data. The methods most commonly used are: the solubility of orthopyroxene in clinopyroxene coexisting with orthopyroxene for obtaining temperatures; and the solubility of Al_2O_3 in orthopyroxene coexisting with garnet for pressure estimates. For reviews of these methods see Boyd (1973) and Harte (1978). Current methods are in the process of being refined and in extrapolating from a simple experimental system to naturally occurring mineral assemblages, it should be appreciated that difficulties do exist and that there is a fundamental assumption that both the experimental and the natural assemblages are equilibrium assemblages. The parameters used, which are instructive for calculating comparative equilibrium conditions of different nodules, are the Ca/(Ca + Mg) ratio of clinopyroxene and the Al_2O_3 content of orthopyroxene. Using these parameters, whereby it is possible to estimate temperatures and pressures of formation of garnet-pyroxene-bearing assemblages, the garnet lherzolites equilibrate as a group over a wide temperature range between 900 and 1400 °C. However, the individual nodule suites from different kimberlite intrusions have equilibrated over different, and sometimes narrow, temperature ranges; for example those from the Bultfontein Mine equilibrate over a very narrow temperature range 950–1050 °C (Dawson *et al.*, 1975) whereas the nodule suite from Thaba Putsoa equilibrates from approximately 900 °C to nearly 1400 °C, with deformed, fertile peridotites equilibrating over the higher part of the range. If the temperatures are taken at their face value, then garnet lherzolites as a group have been derived from depths of between 120 and 210 km; however, Gurney and Harte (1980) have questioned this assumption and proposed that the higher-temperature peridotites are formed within a high-temperature aureole around an ascending upper-mantle diapir, being derived from the same depths as the cooler ones, the latter being distant from, and thereby unaffected by, the mantle diapir, i.e. in the region of 120 to 170 km depth. The greater depths of 170–210 km will only be true if it is assumed that there is no perturbation in the geothermal gradient during the kimberlitic activity. In the case of the rare lherzolites containing diamond, these must have equilibrated on the high-pressure side of the diamond/graphite equilibrium curve, i.e. at depths > 150 km.

The more Fe-rich garnet websterites and garnet-pyroxenites are believed to have originated by partial fusion of garnet lherzolite to form liquids relatively rich in Fe; it is proposed that these liquids did not migrate far from their site of origin and consolidate under exactly the same *PT* conditions as the rocks from which they are derived, i.e. at temperatures of 950 to 1100 °C and 120 to 160 km depth (Gurney *et al.*, 1975).

The sequence garnet lherzolite–garnet harzburgite–harzburgite–dunite has been linked together as being a sequential series; the garnet lherzolite is the most fertile and the dunite is the most barren in the sense that dunite is the end-member after complete extraction of basaltic liquid from parental garnet lherzolite (O'Hara *et al.*, 1975). In the sequence, it is envisaged that elements such as CaO, Al_2O_3, Na_2O, and FeO are removed from lherzolite in the form of basaltic liquid, leaving behind a residium increasingly depleted in these elements; higher Mg/Fe and Cr/Al ratios are characteristic of the residua. With reference to the garnet lherzolites in particular, it has been noted above that there are relatively fertile and relatively barren garnet lherzolites. Whereas this concept of fertile *v.* barren certainly holds for the major elements, Shimizu (1975) has pointed out that deformed xenoliths (which are allegedly fertile) have flat, chondrite-like rare-earth element patterns, whereas coarse xenoliths (allegedly depleted in mobile elements such as K, Rb, etc.) show a relatively high amount of *RE* and, moreover, have a relatively steep, negative slope; there is thus an apparent clash between the major- and trace-element evidence as to which type of garnet lherzolite is the more depleted. One possible explanation is like that postulated by Frey and Green (1974) and Frey *et al.* (1978) to explain similar apparent anomalies in spinel lherzolite suites. The xenoliths that are fertile with respect to their major elements are representative of the undepleted upper mantle, whereas the others that are more barren in terms of major-element analyses have been subjected to a certain degree of melting and depletion relative to the fertile lherzolites; their relatively high *REE* and minor-element concentrations are due to a metasomatic event that has been superimposed upon the xenoliths following the earlier, major-element depletion event. This would not, however, explain why the fertile xenoliths are invariably sheared and to explain this Dawson (1980) has proposed that the sheared, fertile peridotites are not equilibrium assemblages but have been subjected to a certain amount of infiltration metasomatism and perhaps have changed their bulk and individual phase composition during their deformation and recrystallization as a result of diffusive massive transport in narrow and localized upper mantle deformation zones. In this context, Allègre *et al.* (1978) have reported evidence of isotopic disequilibrium in deformed garnet lherzolites.

Eclogites and their variants can theoretically equilibrate over a considerable *PT* range within the upper mantle. The high-level boundary is where plagioclase disappears from rocks of broadly basaltic composition, this being governed by two reactions (*a*) forsterite + anorthite → garnet; and (*b*) albite → jadeite + quartz. Assuming that eclogites are broadly chemically equivalent to various types of basalt and basic granulite, Ringwood and Green (1966) have shown that the bulk composition has some effect on the basic granulite → eclogite transition; for example, at a temperature of 1100 °C the transition from garnet granulite (garnet + pyroxene + plagioclase) to eclogite (garnet + pyroxene + quartz) occurs at 15 kb for rocks of alkali-poor olivine tholeiitic composition, at 18–20 kb for alkali olivine basaltic composition and quartz tholeiitic composition, and at approximately 24 kb for rocks of high-alumina basalt composition. Upper-mantle eclogites have equilibrated at depths below this zone of transition, and attempts to determine the conditions of equilibration are currently one of the more interesting aspects of geothermometry and geobarometry; due to constant refining of the techniques, it is apparent that estimates will be subject to constant revision. For example, one currently used method of geothermometry is based upon a distribution coefficient K_D = (FeO/MgO) garnet/(FeO/MgO) clinopyroxenes for coexisting garnets and clinopyroxenes (Råheim and Green, 1974) which increases inversely with temperature; Roberts Victor eclogites have K_D values ranging from 2.19 to 4.92 (Bishop *et al.*, 1978) which, assuming pressure of 30 kb will give temperatures ranging from 800 to 1200 °C, or assuming a pressure of 20 kb, a range of 800–1100 °C. However, if the Ca content of the garnets is also taken into consideration, as proposed by Ellis and Green (1979) the temperature range of most Roberts Victor eclogites is restricted to a much narrower range of 1017–1160 °C (D. A. Carswell, pers. comm.); this assumes a pressure of 45 kb and all iron as Fe^{2+} and applies also to the diamondiferous eclogites from the Roberts Victor Mine described by Reid *et al.* (1976). This more restrictive range of temperatures, if correct, suggests the Roberts Victor eclogites, including diamondiferous ones, originate from a relatively restricted upper-mantle zone; this makes considerably more sense than previous assumptions that kimberlite ascending through very restricted fissures in the upper mantle could sample a whole series of eclogite pods at various levels.

The pressures are more difficult to determine, though calculated temperatures may give an approximate depth estimate assuming a steady-state geothermal gradient beneath the shields, such as that proposed by Clark and Ringwood (1964). A minimum pressure can be assumed for some eclogites containing distinctive phases; for example, those containing coesite equilibrated on the high-pressure side of the quartz → coesite transition curve (Boyd and England, 1960) at a minimum depth of 100 km, and the use of the experimentally determined curve of anorthite → garnet + kyanite + quartz in the system $CaO-Al_2O_3-SiO_2-H_2O$ (Boettcher, 1970) would give the lowest stability limit for kyanite eclogites and grospydites as ~ 50 km. In addition, eclogites containing diamond equilibrate on the high-pressure side of the graphite/diamond equilibrium curve which, beneath the continents, give depths in excess of 150 km.

Glimmerites and MARID-suite rocks

Nodules containing large proportions of phlogopite occur in many South African and Russian kimberlite pipes, and have been referred to loosely as 'glimmerites'; although the mineralogy is usually dominated by phlogopite, nodules consisting of various combinations of phlogopite, amphibole, clinopyroxene, ilmenite, rutile, apatite, olivine, and zircon are also present. Considerable variations are present in the chemistry of the phases from these mica-rich nodules in South African kimberlites, though Dawson and Smith (1977) do recognize the MARID-suite consisting of variable proportions of compositionally distinct phlogopite, potassic richterite, rutile, ilmenite, and diopside; zircon was considered a possible additional phase of the MARID-suite by Dawson and Smith (1977) and has subsequently been found in MARID-suite xenoliths (J. J. Gurney, pers. comm.). The MARID-suite nodules are characterized by foliation or lineation or both, as a result of alignment of phlogopite and amphibole crystals, and banding is present in some specimens due to different concentrations of the phases; most specimens are deformed, the deformation being evident in the bending of amphibole crystals and undulose extinction of phlogopites. The characteristics of the phases in the MARID-suite are shown in Table III from which it will be noted that the bulk rocks are characterized by high concentrations of K, Ti, H_2O, S, Nb, Zr, Sr, and Ba, reflecting the phases' major- and trace-element concentrations. Dawson and Smith (1977) have proposed that the MARID rocks crystallized under high oxidation conditions from a magma chemically similar to kimberlite within the higher parts of the upper mantle. The presence of amphibole restricts the depth to less than approximately 100 km.

TABLE III. *Characteristics of phases and bulk rocks in the MARID-suite*

MICA; *mg* 0.82–0.88, low Cr_2O_3, low Al_2O_3 (10%), low Na_2O (0.2%)
AMPHIBOLE; *mg* 0.87–0.91, K = Na, (Na + K) > Ca. Richterite
RUTILE; up to 3 wt% Fe_2O_3
ILMENITE; MgO 8–12%, low Al_2O_3 (< 0.2%)
DIOPSIDE; *mg* 0.85–0.89, Ca/(Ca + Mg) ~ 0.45, Fe^{3+} required
Overall; low Al_2O_3, high TiO_2, high Fe^{3+}, low Cr_2O_3, high K/Na

Bulk rock; high concentrations of K, Ti, H_2O, S, Nb, Zr, Sr, Ba.

$mg = Mg/(Mg + Fe_{total})$.

The megacryst suite

Within most kimberlites there occur large crystals of olivine, orthopyroxene, clinopyroxene, garnet, ilmenite, and mica; they are of large size (commonly greater than 2.0 cm) and thereby are inferred not to have been derived by fragmentation of the commoner types of upper-mantle xenoliths. These large single crystals have been referred to in the literature both as 'megacrysts' (the terminology adopted here) and 'discrete nodules' (Nixon and Boyd, 1973*b*); the two terms are synonymous. Two series of megacrysts are known, the first in which the phases are relatively rich in chrome, and a second in which the phases are relatively deficient in chrome.

The chrome-rich megacrysts are much rarer and have been described from the pipes of the Colorado plateau (Eggler and McCallum, 1976) and also from the Excelsior Mine, South Africa (Boyd and Dawson, 1972). In the case of the specimens from the Excelsior Mine, the garnet megacryst is a chrome-rich (8.5 wt% Cr_2O_3) and low calcium (2.7% CaO) garnet, intergrown with serpentine, representing pseudomorphs after ?olivine or ?orthopyroxene; the composition of this pyrope is similar to that of many garnets occurring as inclusions in diamond and also occurring as isolated crystals in some kimberlite pipes (e.g. Finsch Mine—Gurney and Switzer, 1973). The megacryst suite at Excelsior also contain high-Ti pyrope and magnesian ilmenite (Dawson, unpublished data); similarly, the megacryst suite of the Sloan kimberlite diatreme, Wyoming, contains examples of both Cr-rich and Cr-poor megacrysts of pyrope, clinopyroxene, and orthopyroxene. Consistent features are higher Mg and Cr in one suite compared with higher Ti and Fe in the other; in addition, chromite and olivine occur as inclusions in the high-chrome megacrysts whereas magnesian ilmenite may be regarded as an additional member of the low-chrome megacryst suite.

In most kimberlites, however, the main megacryst group is the chrome-poor one and, in addition to containing the large single crystals of olivine, orthopyroxene, clinopyroxene, garnet, ilmenite, and mica, there are many cases where small inclusions of one mineral species are found within a single large crystal of another, and in some cases intergrowths occur. The most extensive study of low-chrome megacrysts from a single locality is that by Gurney *et al.* (1979) on the megacrysts from the Monastery Mine, South Africa; the megacrysts show a substantial range in chemical variation particularly in their Mg/Fe ratios. Olivines fall into two groupings, those within a compositional range Fo_{84-8} with an NiO content of approximately 0.3 %, whereas the other groups have compositions of Fo_{72-8} and with NiO less than 0.12 wt %. Orthopyroxene forms large homogeneous megacrysts, sometimes forming irregular intergrowths with ilmenite, and they tend to have variable Mg/Fe ratios though the Ca content is relatively constant; only the more Fe-rich ones may form intergrowths with ilmenite. Clinopyroxenes vary from subcalcic diopsides through titanian chrome-diopsides to low-chrome diopsides (terminology of Stephens and Dawson, 1977) and more calcic ones may form regular lamellar intergrowths with ilmenite. Garnets define a trend of changing Mg/Fe ratios at constant Ca content (rather similar to the orthopyroxenes) with the more Fe-rich ones sometimes containing inclusions of magnesian ilmenite; in all cases they are low in chrome though the more magnesian ones are high-Ti pyropes or titanian pyropes (terminology of Dawson and Stephens, 1976). Ilmenites range widely in composition particularly with respect to Mg/Fe, but those coexisting as intergrowths with clinopyroxene and orthopyroxene and those existing as inclusions in garnet are all very similar in composition, having the lowest Mg/Fe ratios. Of the silicate-oxide intergrowths, the most spectacular are the lamellar intergrowths between low-Cr diopside and magnesian ilmenite. In these cases, fine lamellae of ilmenite grow parallel to the long axes of large host clinopyroxenes which may be up to 20 cm in diameter (Dawson and Reid, 1970); although previously regarded as rarities, they have now been described from numerous kimberlite localities in South Africa and Lesotho, and others are now known from kimberlite intrusions in the USSR and the USA. The less spectacular irregular intergrowths of orthopyroxene and ilmenite are less common, but none the less occur at several localities in southern Africa and one locality in the USSR.

As noted above, some megacrysts occasionally contain small inclusions of another species of the megacryst suite and the tie-lines between coexisting garnets and pyroxenes or garnets and orthopyroxenes are in excellent accord with theoretical tie-lines calculated on the results of experimental systems. In addition, tie-lines between garnet and olivine suggest that the most subcalcic clinopyroxene and the most magnesian garnets, orthopyroxenes and olivines have formed in equilibrium with each other, and that gently rotating concordant tie-lines may be drawn for successively more Fe-rich compositions. As a result, Boyd and Nixon (1975) and Gurney *et al.* (1979) broadly agree that these trends indicate that the megacrysts represent a single differentiation sequence and that the megacrysts are cumulate minerals that precipitate from a magma, possibly the host kimberlite. Coexisting garnets and ilmenites have been synthesized in the pressure range 21 to 40 kb and the temperature range of 900–1100 °C from olivine basanite compositions (Green and Sobolev, 1975); the compositions of the synthesized ilmenites and garnet are very similar to those in kimberlite and, in view of the Mg/Fe ratios and amounts of Al_2O_3 and Cr_2O_3 in their synthesized garnets, Green and Sobolev (1975) suggest that natural garnet/ilmenite pairs crystallized within these experimental temperature ranges but under varying pressures and oxygen fugacity conditions.

Another well-investigated megacryst suite is that from the kimberlites of northern Lesotho (Boyd and Nixon, 1975); the phases and range of composition are very similar to those from the Monastery Mine, but there is an additional point of great interest. From the Thaba Putsoa intrusion Boyd and Nixon (1975) describe composite nodules formed of subcalcic diopsides and titanian pyropes, the composition of which can be matched with individual megacrysts; they are interpreted as aggregates of megacrysts which have crystallized from the kimberlite magma, and as such conform to the original definition of griquiate (Beck, 1899). However, it must be stressed that the composition of these phases are quite unlike those found in other kimberlite eclogites, despite the fact that the terms 'griquaite' and 'eclogite' are often used synonymously.

As noted above, there is an aggregation of phases into composite rocks; it has been suggested that this may take place in an upper-mantle magma-chamber (Gurney and Harte, 1980), and Haggerty *et al.* (1979) suggest that the pyroxene-ilmenite intergrowths may reflect formation in upper mantle pegmatites.

Summary

In summary, it can be said that the upper mantle is mineralogically complex and extremely heterogeneous. An upper harzburgite zone, persisting to a depth of 100-120 km, is underlain by a garnet lherzolite zone but within these two major bodies variations exist since both barren and fertile variants of harzburgite and garnet lherzolite are known. These rock-types are consistent with the material demanded by seismic constraints. Within these two major zones are variants such as lherzolite in the garnet lherzolite zone, the presence of which may be due to low alumina, and there are rare examples of spinel lherzolite, the stability field of which is within the harzburgite zone; also in the upper harzburgite zone there is a possibility of heterogeneous bodies of glimmerite and MARID-suite rocks. In addition, if estimates of the equilibration range of eclogites is correct, small bodies of eclogite (themselves heterogeneous) will be found throughout both the harzburgite and garnet lherzolite zones of the upper mantle.

Hydroxyl-bearing species are amphibole and high-Ti phlogopite in the harzburgite zone, and low-Ti phlogopite in the garnet lherzolite zone. Experimental data suggest that calcite will be the main CO_2-bearing species at higher levels, giving way to dolomite and magnesite at increasing depths (Wyllie, 1979).

Despite the other chemical variations within the upper mantle, modern data suggest that the stable isotope values of various elements within the upper mantle are reasonably uniform; these values are summarized in Table IV. Current data suggest the following isotopic ranges for phases derived from the upper mantle: δD -40 to -80 ‰; $\delta^{13}C$ -2 to -9 ‰; $\delta^{18}O$ 2 to 8.5 ‰; $\delta^{34}S$ 1 to 2 ‰. The reasons for certain departures of some values in upper mantle phases from these preferred ranges (e.g. the high-δD values for richterites and the high $\delta^{13}C$ values for some diamonds) is not well understood, although in the case of diamond it is a possibility that isotopic fractionation may have occurred or the values been modified by interaction with C-rich fluids during the kimberlite event. The S isotope values are extremely close to those of chrondites. There are as yet insufficient data to say whether samples in kimberlite can give us any idea as to changes in the isotopic values of the mantle throughout geological time, though there is some evidence of slight but distinct differences in the isotopic composition of the C in diamonds from the Precambrian Premier Mine kimberlite compared with those in the Late-Cretaceous South African kimberlites (Deines, in press).

TABLE IV. *Stable isotope values of upper-mantle rocks and phases*

		Values ‰	
Hydrogen δD	Phlogopites, glimmerite nodules, Dutoitspan and Wesselton	$-43.2--77.4$	Kuroda *et al.* (1975), Sheppard and Dawson (1975)
	Richterites, glimmerite nodules, Dutoitspan and Wesselton	$-103.1--132.5$	Kuroda *et al.* (1975)
	Phlogopite, peridotite nodule, Bultfontein	-36.2	Kuroda *et al.* (1975)
	Richterite, peridotite nodule, Bultfontein	-79.4	Kuroda *et al.* (1975)
	Phlogopite megacrysts, Monastery kimberlite	-65 ± 2	Sheppard and Dawson (1975)
	Phlogopite megacrysts, kimberlites, USA and India	$-48--71$	Sheppard and Epstein (1970)
Carbon $\delta^{13}C$	Graphite, garnet peridotite, Mir	-5.9	Kropotova and Fedorenko (1970)
	Diamond, eclogite, Mir	-8.5	Kropotova and Fedorenko (1970)
	Kimberlite carbonate (various sources)	$-2--9$	Sheppard and Dawson (1975)
	Primary carbonatites (various sources)	$-2--9$	Sheppard and Dawson (1975)
	Diamonds (various sources)	$-2--32.3$ (mainly $-3--9$)	Deines (in press)
Oxygen $\delta^{18}O$	Garnets and cpx., gt. lherzolite	5.1-5.8	Garlick *et al.* (1971)
	Eclogites, Roberts Victor	2-8	Garlick *et al.* (1971)
	Garnet, grospydite, USSR	6-6.3	Garlick *et al.* (1971)
	Carbonatites (primary)	6-8.5	Sheppard and Dawson (1975)
	Phlogopite megacrysts in kimberlite	4-8	Sheppard and Dawson (1975)
Sulphur $\delta^{34}S$	Gt. pyroxenite, Obnazhennaya	$-0.5-+2.1$	Grinenko and Ukhanov (1977)
	Eclogite, Obnazhennaya	0.5-1.8	Grinenko and Ukhanov (1977)
	Eclogites, Roberts Victor and Premier	0.2-2.1	Tsai *et al.* (1979)

Processes in the upper mantle

Upper-mantle movements. It has been mentioned above that many peridotites, especially garnet lherzolites, show deformation fabrics akin to those found in crustal cataclasites and mylonites. The textures range from those in which the only sign is marginal recrystallization of olivine to others in which pyroxene neoblasts are strung out into lamellae, and more resistant garnets such as spinel and garnet have been disrupted and strung out into chains of comminuted crystals; these textures, classified by Harte (1977), have been attributed to super-plastic flow (Boullier and Gueguen, 1975). That these fabrics are the result of differential movement within the upper mantle is generally accepted, but the site of this deformation and its relationship, if any, to major earth movements has been the subject of considerable debate. For example, Boyd (1973) proposed that the deformation fabrics in xenoliths from South African Cretaceous kimberlites were due to shear at the base of the subcratonic lithosphere during the breakup of Gondwanaland, with the implication that they were due to strong horizontal movements. Conversely, other workers (e.g. Green and Gueguen (1974) and Gurney and Harte (1980)) have argued that the textures were formed at a higher level in the upper mantle around an ascending diapir. However, thermal considerations indicate that annealing would obliterate the textures in the thermal regimes pertaining at the base of the lithosphere (Goetze, 1975), and it is now generally accepted that the textures were formed during the kimberlite event, the textures having been frozen in as a result of rapid transportation to the surface of the xenoliths by the host kimberlite.

Although these spectacular deformation fabrics can be interpreted as being connected with the kimberlite event, there are, none the less, some textures that were imposed upon the rocks in a deformation event prior to the one connected with the magmatism that transported them to the surface. These textures are seen in coarse-grained peridotites in which there is strong preferred orientation in the olivines and the orthopyroxenes (the so-called tabular granuloblastic texture of Harte, 1977), and Boullier and Nicolas (1975) illustrate convincing examples of rocks, formerly exhibiting laminated and disrupted textures, in which the olivines are now annealed and enclose small grains of disrupted garnet and lamellar bands of pyroxene. At the Matsoku pipe, Harte *et al.* (1975) have recognized certain xenoliths which have been deformed to varying degrees and which have been subsequently annealed. Other peridotites (described in the later section) consist of large, sheared orthopyroxene crystals surrounded by finer-grained but, none the less, coarse grains of olivine, clinopyroxene, and garnet; these are the annealed equivalents of former porphyroclastic-textured rocks in which exsolution of garnet and clinopyroxene from orthopyroxene took place during the earlier deformation event and Dawson *et al.* (1980) have emphasized the importance of directed pressure in catalizing garnet and clinopyroxene exsolution.

There is also a limited amount of evidence for brittle fracture within the upper mantle. Fractured orthopyroxenes in peridotites have already been referred to above, the inference being that although the olivine was able to recrystallize, the strain rate was greater than could be accommodated by recrystallization of orthopyroxene; this in itself is an illustration of the anisotropic response of an individual rock type towards external strain. By contrast, some eclogites are inequidimensional in shape, their long axes being often at a high angle to any observed banding; this can be interpreted as being due to tightly spaced jointing being imposed upon the eclogite just prior to its incorporation into the kimberlite, the fracturing possibly being due to the hydraulic pressures of ascending kimberlite. Further examples of brittle fracture may be seen in veined and metasomatized peridotites in which veinlets of mica, amphibole, diopside, and ilmenite have been injected along the planes of fracture within the nodules; angular blocks bounded by planar surfaces coated with these minerals are interpreted as being fragmented veined peridotites (Dawson, 1979).

Upper-mantle metasomatism. In many xenoliths exhibiting both the superplastic and brittle-deformation textures noted above, the deformation is accompanied by metasomatism. In many deformed peridotites, the presence of intergranular serpentine, phlogopite, and ilmenite testifies to the introduction of fluids rich in water, K_2O, and TiO_2 which have produced new minerals by interaction with existing olivine, garnet, and pyroxene. The most spectacular examples of this combination of deformation and metasomatism are the polymict peridotites described by Lawless *et al.* (1979); strongly deformed clasts of both peridotitic and eclogitic rocks (disequilibrium assemblages) are cemented by ilmenite (phlogopite), rutile, and sulphides. This group of metasomatic minerals are by far the most common combination of phases produced during infiltration metasomatism; other rarer species are richterite and carbonate. Some of the best-studied metasomatic peridotites are those from the Matsoku pipe, Lesotho (Harte and Gurney, 1975); here the metasomatic minerals include the four named above and both ilmenite and rutile show spinel exsolution. The minerals that are disseminated through a matrix of deformed and recrystallized nodules, can occur very distinctly in veinlets, narrow zones, and patches in coarse peridotites and pyroxenites, and it is possible that the veinlets represent original planar fractures. Other good examples of metasomatized peridotites are described from the Bultfontein by Erlank and Rickard (1977) and Dawson (1979). In some cases metasomatism is found alongside veins containing potassium richterite, ilmenite, sulphides, phlogopite, and ilmenite; in the wall-rock peridotite on either side of the vein, garnet and pyroxenes are replaced by richterite, phlogopite, and ilmenite though the fact that these minerals may develop in host harzburgites indicates that original garnet and clinopyroxene is not a prerequisite for development of the metasomatic phases as suggested by Erlank and Rickard (1977). Potassic richterite and phlogopite may also be developed in other peridotites in which there is no obvious veining, but whether this is due to the small scale of the specimens available and the metasomatism is due to diffusion and infiltration from previously adjacent veinlets, or whether this is due to a larger metasomatic event must remain open to question. The available evidence shows that all the metasomatic minerals are extremely rich in elements which are generally depleted in most kimberlite xenoliths. For example, some of the richterite-bearing peridotites contain large amounts of K and rutile-rich primary veins from Matsoku contain greater than 18% TiO_2. Over all, the infiltration metasomatic effects indicate the presence and migration within the upper mantle of fluids capable of generating phases rich in Fe, K, Ti, H_2O, S, Nb, Zr, Hf, Sr, Rb, Ba, Na, and CO_2. It may be noted that the combinations of phases in veined peridotites are very similar to those found in the MARID-suite xenoliths, the formation of which has been attributed to the consolidation in the upper mantle of fluids of chemistry very similar to kimberlites (Dawson and Smith, 1977).

Thermal metamorphism. It has been noted earlier that some of the sheared xenoliths, for example those from the Thaba Putsoa kimberlite pipe, have equilibrated at relatively high temperatures—in the temperature range 1300–1400 °C. In earlier models it was suggested that these high temperatures were due to their derivation from greater depths combined with the effects of shear heating at the base of some cratonic lithospheric plates (Boyd, 1973). It has been subsequently demonstrated that there is no over-all correlation between increased deformation and increased temperature of equilibration of peridotites (Dawson *et al.*, 1975) and various authors (e.g. Green and Gueguen, 1974; Mercier, 1979) suggest that the high temperatures are due to further metamorphism of normal peridotite in the envelope of an ascending mantle diapir. A later model, that of Gurney and Harte (1980), proposes that the deformation, metasomatic, and high-temperature features seen in the Thaba Putsoa type of sheared xenolith have been imposed upon normal peridotites in the envelope of an ascending upper-mantle diapir.

In summary, it can be said that certain areas within the upper mantle have been subjected to magma intrusion which was possibly accompanied by thermal metamorphism, shearing, and metasomatism. The magma injection may have been accompanied by brittle and plastic deformation and the brittle deformation in some cases may have opened channel ways for injection of fluids that consolidated as upper-mantle veinlets, some of which may have imposed limited metasomatic effects upon their wall rocks.

Diamonds in the upper mantle

Because kimberlite is by far the most abundant source of terrestrial diamond, it has been taken for granted for many years that kimberlites and their associated diamonds are somehow genetically connected, and it was assumed that the diamond was a precipitate from the kimberlite magma. More recently, however, this hypothesis has been challenged by workers postulating that diamond is xenocrystal, having been derived by fragmentation of diamondiferous eclogite and peridotite. This has been brought about, to some extent, by the fact that diamondiferous eclogites and peridotites are turning out to be commoner than was formerly supposed; it is now known that at kimberlite localities where eclogites form the largest proportion of the xenolith suite, e.g. at Roberts Victor and Bellsbank area (S. Africa), Orapa, Botswana, and Mir, USSR, diamondiferous eclogites are relatively common. Furthermore, diamond has now been reported in garnet lherzolite—a potentially much more abundant source rock for diamond than eclogite; diamonds have been reported in garnet lherzolites from Mothae (Lesotho) (Dawson and Smith, 1975*a*); from a kimberlite pipe in southern Wyoming (McCallum and Eggler, 1976), from the Udachnaya pipe (Pokhilenko *et al.*, 1977), and more recently from the Finsch Mine, South Africa (J. B. Hawthorne, pers. comm.). It may be noted that the phases in these rocks containing diamonds are of fairly normal composition (e.g. chrome pyrope, chrome diopside, enstatite, olivine Fo_{90-2}) and are unlike the rather unusual compositions of phases in diamondiferous garnet serpentinite or dunite xenoliths from some Russian intrusions in which the garnets contained between

8 and 15% Cr_2O_3 (Sobolev, 1977; Pokhilenko *et al.*, 1977); in addition it has been argued (Robinson, 1978) that the chemical similarity between mineral inclusions in diamond and those in the eclogites and peridotite xenoliths points to their derivation by fragmentation of xenoliths, as proposed by Sobolev (1977). Another feature favouring a xenocrystal origin is the fact that some diamonds have been plastically deformed, and have had a long and complex growth history (Frank, 1969). In addition, the included material in some diamonds in Cretaceous kimberlites has yielded Precambrian ages (Kramers, 1979). People favouring the xenocrystal theory propose that kimberlite is merely a medium whereby diamond is transported from the upper mantle to the surface, and postulate that the reason for diamonds being virtually restricted to kimberlite is that only kimberlite originates at sufficient depths within the mantle to sample diamond-bearing mantle.

Proponents of the phenocrystal theory point out that, whilst acknowledging the existence of diamondiferous eclogites and peridotites, an unusually large proportion of these diamondiferous xenoliths would have had to be preferentially fragmented (compared with non-diamondiferous xenoliths) in order to shed the number of diamonds that are now found in kimberlite; although this argument may be countered by arguing that such deep-seated xenoliths may have undergone preferential expansion, thereby causing possible self-fragmentation, observed diamondiferous xenoliths are by no means more friable than non-diamondiferous ones. Theories for the phenocrystal origin also rely to a certain extent on diamond surface features and also upon evidence from internal inclusions. Delicate growth features which have been recorded on the surface of diamonds could only have been achieved by crystallization of diamond from a liquid or precipation of late diamond on to an existing surface from a fluid (Sunagawa *et al.*, 1978), but there is no good reason why this overgrowth should not have been a relatively late-stage effect. Likewise, the presence in diamonds of submicroscopic liquid inclusions approximating to carbonated picritic composition (Fesq *et al.*, 1975) are taken to indicate that diamonds could only have been derived by crystallization from a liquid; however, there is no evidence to suggest that the included liquid can be unambiguously equated with the host kimberlite and it has also been shown that delicate growth features similar to those found on kimberlite diamonds have also been found on diamonds in eclogite xenoliths (Robinson, 1979), being due to the growth of diamond from a much older liquid unconnected with the kimberlite event.

A further feature for consideration in this debate is the significance and interpretation of the inclusions in diamonds. Broadly speaking, the inclusions can be broken down into two broad groupings—(*a*) those of eclogitic composition; and (*b*) those belonging to an ultrabasic paragenesis. (For reviews, see Sobolev, 1977; Harris and Gurney, 1979). It is certainly correct to say that the garnet and pyroxene compositions in the 'eclogitic' diamond inclusion paragenesis can be to a large extent matched with the pyrope-almandine garnets and omphastic pyroxenes in eclogite xenoliths: and hence it can be argued that the diamonds containing phases of similar compositions must be xenocrysts. However, although a very few of the phases in the diamond-inclusion 'ultrabasic' paragenesis can be matched with those in 'normal' garnet lherzolites, most of them have subtle compositional differences from those in normal garnet lherzolite rocks; for example, compared with the phases in normal garnet lherzolites, the diamond-inclusion garnets tend to be low in Ca but relatively rich in Cr, the olivines tend to be more forsteritic (Fo_{94} *v.* Fo_{90-2}) and also contain a certain amount of divalent Cr, the enstatites are more magnesian (En_{95} *v.* En_{92-4}), clinopyroxenes are extremely rare and chromites are extremely rich in Cr_2O_3—generally greater than 60% Cr_2O_3 (Sobolev, 1977; Hervig *et al.*, in press). In short, although some of these diamond-inclusion phases can be matched with phases in ultrabasic xenoliths (Dawson and Stephens, 1975; Stephens and Dawson, 1977), it would have to be argued that, if the ultrabasic diamond inclusion suite is typical of the regime whence xenocrysts are derived, xenoliths derived from this chemical regime within the upper mantle are extremely rare amongst the xenoliths found in kimberlite pipes; this is in sharp contrast to the 'eclogite-suite' inclusions. In addition to the compositional differences, Boyd and Finnerty (in press) and Hervig *et al.* (in press) show that the equilibration temperatures of most inclusion pairs fall mainly below 1100 °C, suggesting that most diamonds may have formed in a subsolidus event.

As an alternative explanation for the high-Cr, low-Ca nature of the peridotitic inclusions, Harte *et al.* (1980) have proposed that this inclusion suite is formed during partial melting of normal lherzolite, when preferential partition of Ca and Al into a high-carbonate kimberlite liquid results in relict phases with the low-Ca, high-Cr characteristics of the olivine, garnet, orthopyroxene, and chromite commonly found as the ultrabasic diamond inclusion paragenesis; the crystals are still xenocrystal but they have been stripped of their Ca during melting, and the diamond can precipitate from the liquid and overgrow and include these modified xenocrysts.

Another hypothesis, that of Mitchell and Crockett (1971), is that diamond seed nucleii may form in a kimberlite magma within the diamond stability field but this is then followed by a stage of overgrowth of metastable diamond on the seed nucleii during the ascent. An extension of this hypothesis is that during the stage of metastable diamond overgrowth, the isotopic composition of the diamond might change (Mitchell, 1975*a*). Another theory has been put forward by Dawson and Smith (1975) following their discovery of small irregular diamond in garnet lherzolite; recognizing the fact that, during the industrial synthesis of diamond, the main problem is one of nucleation (a problem solved by introducing small amounts of diamond seed crystals), these authors propose that, during a mantle-melting event resulting in the formation of kimberlite magma, small xenocrysts of diamond are shed from fragmented diamondiferous peridotite (as pointed out earlier, a much more widespread source of xenocrystal diamond than eclogite) and act as nucleii for diamond precipitating from the kimberlite.

Whichever of the various hypotheses for diamond formation proves to be correct, three facts seem to be fairly well established. The first is, that judging from the presence of diamond in eclogite of reasonably wide composition (varying from 'normal' eclogite to grospydite compositions), and if the ultrabasic-paragenesis inclusion suite is any guide, then diamond must form in a wide spectrum of chemical environments; in view of the wide range in the chemistry of kimberlite magmas, this spectrum is widened even further if diamond can be proved to have precipitated from kimberlites. Second, relative to the widespread garnet lherzolites, diamond is apparently preferentially concentrated within the very much smaller bodies of eclogite, suggesting that modification of original mantle material is a feature favouring diamond formation, possibly as a result of concentrating volatile matter. Finally, there is an anomaly in hypotheses proposing that both diamond and the megacryst suite of minerals are precipitated from kimberlite. The megacryst-suite phases are richer in FeO and TiO_2, and poorer in Cr_2O_3, than equivalent phases in the diamond-inclusion suite and are quite distinct (Table V), hence, it appears that diamond cannot be regarded as having any genetic link with the megacryst suite.

TABLE V. *Chemical differences between selected diamond-inclusion phases and equivalent phases from the megacryst suite*

	Diamond[1] inclusions	Megacrysts[2]
Olivines		
TiO_2	<0.01	0.02-0.03
Al_2O_3	<0.02	0.04-0.17
Cr_2O_3	0.03-0.07	0-0.03
% Fo	92-4	78-88
Orthopyroxene		
TiO_2	0-0.03	0.08-0.20
Al_2O_3	0.44-0.77	0.75-1.30
Cr_2O_3	0.18-0.55	0.02-0.27
Mg/(Mg+Fe)	93-4	83-94
Garnets		
TiO_2	0-0.03	0.64-0.88
Cr_2O_3	5.53-10.1	0-2.09
CaO	0.59-1.36	4.34-5.06
Mg/(Mg+Fe) × 100	88-90	68-81

[1] Data from Harris and Gurney (1979).

[2] Data from Gurney *et al.* (1979).

Relicts from the early upper mantle?

The assemblages we now see in most upper-mantle xenoliths are metamorphic; even eclogites and glimmerites, generally regarded as frozen liquids, now have metamorphic fabrics. However, in a few rare tantalizing cases we can see fortunate survivors of earlier rock types which have not been completely eradicated by subsequent metamorphic events. Study of these survivors, in which reactions have not gone to completion, are enabling us to deduce something of the origin of garnet lherzolite, a major rock-type in the upper mantle, the presence of which is taken for granted, and the origin of which is rarely questioned. For example, Dawson *et al.* (1980) describe garnet lherzolite specimens in which large fractured crystals of orthopyroxene are set within a fine-grained recrystallizing matrix of olivine; garnet and clinopyroxene have exsolved along kink bands in the orthopyroxene megacrysts around which the exsolved garnet and diopside tend to cluster. By combining evidence from various specimens they propose: (*a*) the presence of an ultra-coarse-grained harzburgite precursor; and (*b*) the generation of specific garnet lherzolites (containing < 5 % modal garnet and < 3 % modal clinopyroxene) by deformation of the harzburgite precursor with attendant exsolution of garnet and clinopyroxene from the earlier Ca- and Al-rich orthopyroxene. Generation of garnet lherzolite in yet another way is that described by Haggerty (1979) in which garnet lherzolite is being formed from former spinel lherzolite, presumably by an

TABLE VI. *Examples of metamorphic upper-mantle assemblages and their inferred precursors*

Existing assemblage	Inferred earlier assemblage	Locality	Reference
1. Ca-garnet + olivine (+cpx and spinel) Garnet dunite	Cpx + spinel (? + olivine) Wehrlite	Frank Smith	Dawson (unpubl.)
2. Gt + ol + cpx + opx Gt lherzolite	Ol + opx Harzburgite	Monastery Mine	Dawson *et al.* (1980)
3. Gt + ol + di + opx (+sp) Gt sp lherzolite	Ol + di + opx + sp Sp lherzolite	Jagersfontein	S. E. Haggerty (unpubl.)
4. Gt + ol + opx Gt harzburgite	Ol + opx + sp Sp harzburgite	Liqhobong (Lesotho)	Dawson *et al.* (1978)
5. $Gt_2 + cpx_2$ Eclogite	$Ky + gt_1 + cpx_1$ Kyanite Eclogite	Roberts Victor	Lappin and Dawson (1975)
6. $Gt_2 + cpx_2$ Eclogite	Gt + Al-rich cpx Gt. clinopyroxenite	Roberts Victor	Harte and Gurney (1975)

Note: In cases 1, 3, 4, and 5, new phases have formed at the expense of pre-existing ones, relicts of which may still be present (in brackets). In cases 2 and 6 new phases arise by exsolution from aluminous pyroxenes.

increase in pressure, during which original spinel is replaced and overgrown by garnet. Other examples of spinel being overgrown and replaced by garnet are noted in lherzolite from the Obnazhennaya pipe (Sobolev and Sobolev, 1964), in olivine pyroxenite from the Slyudanka pipe (Frantsesson, 1970), in lherzolite from Green Knobs, New Mexico (Smith and Levy, 1976), and in lherzolite from the Somerset Island kimberlite, Canada (Mitchell, 1978). Evolution of eclogite from a possibly igneous precursor is described by Harte and Gurney (1975) in which garnet is exsolving from an aluminous clinopyroxenite to form what is now a fairly normal eclogite.

A list of inferred earlier assemblages with their present-day mineralogy is given in Table VI. One of these specimens, the wehrlite from the Frank Smith Mine, will be discussed further. In this specimen, BD 2014/1, garnet is being formed at the interface between clinopyroxene and a chrome-rich spinel, the reaction being approximated by the formula:

$$2\,\underset{\text{Diopside}}{Ca(Mg,Fe)Si_2O_6} + \underset{\text{Cr-spinel}}{(Mg,Fe)(Cr,Al)_2O_4} \rightarrow$$

$$\underset{\text{Uvarovitic garnet}}{Ca_2(Mg,Fe)(Cr,Al)_2Si_3O_{12}} + \underset{\text{Olivine}}{(Mg,Fe)_2SiO_4}$$

Compositions of the phases are given in Table VII; attention is drawn to the unusual composition of the garnet which is unusually high in both CaO and Cr_2O_3. Garnets of this composition are rare, and two examples, one from a wehrlite xenolith in a Russian kimberlite, another included in diamond, are given in Table VII. BD 2014/1 is regarded as the intermediate step in the requilibration of the wehrlitic paragenesis and, had the reaction between chrome spinel and diopside gone to completion, the resulting garnet would have been richer in CaO and Cr_2O_3, thereby resembling the uvarovite-rich garnets from the kimberlites of the Newlands Mine, described by Clarke and Carswell (1975). These authors attributed the origin of these uvarovite garnets to prograde metamorphism of

TABLE VII. *Chemistry of phases in wehrlite BD2014/1 (Frank Smith Mine) and garnets of similar composition*

	Olivine	Diopside	Cr-spinel	Garnet	X	Z
SiO_2	41.1	54.9	0.06	40.0	41.0	40.3
TiO_2	0.02	0.03	0.36	0.12	0.46	0.44
Al_2O_3	0.03	1.43	11.5	16.8	16.2	13.9
Cr_2O_3	0.04	1.51	55.4	8.26	8.00	8.89
FeO	7.39	1.69	20.3	6.41	4.66	7.61
MnO	0.09	0.06	0.28	0.37	0.30	0.36
MgO	51.0	17.3	12.9	14.8	15.2	13.2
NiO	0.45	0.07	0.18	0.02	n.d.	n.d.
CaO	0.02	22.8	0.0	12.7	13.5	14.1
Na_2O	0.01	1.14	0.0	0.01	0.07	n.d.
	100.00	100.8	100.9	99.5	99.39	99.8

Analyst: J. V. Smith.

X—Garnet from gt-ol-di inclusion assemblage in diamond, Udachnaya pipe (Sobolev *et al.*, 1970).

Z—Garnet from wehrlite D-332, Dalnaya pipe, Yakutia (Sobolev *et al.*, 1973).

an earlier wehrlite assemblage, an interpretation that is given some support by the reaction relations observed in the Frank Smith specimen.

Whereas most of the reactions listed in Table VI are relatively uncommon, there is one subsolidus reaction which is relatively widespread. This is the reaction known as 'kelyphitization', as a result of which garnets are surrounded by a reaction rim of secondary minerals; phlogopite is the most common and occasionally within the phlogopite there is a brown spinel. In a study of reaction coronas developed by reaction between garnet and olivine in peridotites from the Lashaine volcano, Tanzania, Reid and Dawson (1972) suggested that the reaction coronas around garnets in kimberlite peridotites results from a two-stage process—first an isochemical reaction between garnet and olivine to give a two-pyroxene-spinel assemblage, followed by metasomatism during which the high aluminous pyroxenes are replaced mainly by phlogopite. Quite recently, garnet lherzolites with fresh reaction rims, resembling those in the Lashaine peridotites, have been found in the Letseng kimberlites in northern Lesotho (Lock and Dawson, 1980); the secondary spinels are highly aluminous, the orthopyroxenes contain up to 10% Al_2O_3, and in some cases reaction has gone to completion with garnet being completely replaced by the pyroxene/spinel assemblages which have been smeared out and interspersed with recrystallizing olivine neoblasts; it is quite feasible that upon annealing of such specimens the resulting rocks would be peridotites containing high alumina spinel which would be extremely similar to the aluminous-spinel facies peridotites which are believed to exist in the upper parts of the upper mantle (Lock and Dawson, op. cit.). Hence, some Al-spinel lherzolites may have a garnet lherzolite precursor.

Concluding remarks

The above remarks are an all-too-brief summary of our present state of knowledge of the make-up of the sub-cratonic upper mantle and the processes operating there. Although our knowledge of the upper mantle has increased considerably over the past twenty years, this summary must still be regarded as an interim or progress report, since much remains to be learnt about the upper mantle. For example, although there is now a plausible source in known upper-mantle materials for most elements that are transferred from the upper mantle to the Earth's crust by various types of igneous activity, there are some important elements for which we have as yet no known source. One of these elements is phosphorus, and although most plausible calculations (e.g. Frey *et al.*, 1978; Smith *et al.*, in press) suggest that P together with halogens and many other compatible elements should have a primary source in upper-mantle apatite, there are as yet no documented examples of primary apatite being found in upper-mantle source rocks. Furthermore, although there are known sources of S and C, in the form of sulphides and diamond within the upper mantle, the amounts of both diamond and sulphides in known upper-mantle materials are too small to account for the amounts of oxidized C and S which are discharged at the surface of the Earth during volcanic eruptions. In addition, although it is generally acknowledged that CO_2 is an important volatile species in the initiation and propagation of ultrabasic igneous activity (Wyllie, 1979), the amount of carbonates observed in upper-mantle rock types is quite trivial. Moreover, the only recognizable upper-mantle source for N, an important element in the atmosphere and biosphere, is as platelets or aggregates within diamond; again we have a mass balance problem.

It is now generally believed that upper-mantle phlogopite is a source of the P, Ti, and H_2O but if, as postulated by Frey *et al.* (1978), the phlogopite is the product of a metasomatic event imposed upon earlier depleted lherzolites, then an even deeper source for these elements must be found. From this particular aspect and in the context of migration of fluids in the upper mantle in general, the study of the veined metasomatized peridotites will be particularly important. Other exciting aspects will be the study of whether chemical equilibrium has been attained in upper-mantle assemblages; and further studies on opaque phases which may be possible sources of P and N in upper-mantle rocks if analogies can be drawn with phosphides and nitrides in meteorites. In studies on isotope equilibrium in particular, we are fortunate that technology has now handed us a new powerful tool in the form of the ion microprobe. Just as the electron microprobe enabled us to make major advances in mineralogy in the past two decades, so the advent of the ion probe will inevitable lead to a new wealth of data. It can be confidently predicted that petrological and geochemical studies on upper-mantle xenoliths will make a major contribution to the new IUGS International Project 'The Lithosphere—the Frontier of the Eighties'.

Acknowledgements. My work on xenoliths has benefited greatly by discussion with many friends and colleagues; in particular I should like to acknowledge J. J. Gurney and J. V. Smith. The manuscript of this paper was typed by M. Townsend and S. Forster, to whom I should like to express my thanks.

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[*Manuscript received 9 June 1980*]