

## ULTRAMAFIC ROCKS AND THE UPPER MANTLE

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### ABSTRACT

Cosmic abundances of elements and meteorite analogies indicate an ultrabasic mantle. Detailed review of ultramafic rocks in their varied associations suggests that representative mantle samples are included among oceanic and orogenic peridotites, and nodules in basalts and kimberlites. Extrapolation from rocks to mantle must be made with caution; their petrogenesis is complex and the mantle is heterogeneous. Twenty-seven estimates of mantle composition based on meteorite models, ultramafic rocks, and hypothetical peridotites are tabulated and compared with basalts in chemical variation diagrams. The three groups are separated by their alkali contents and stages of fractionation. Mantle estimates based on ultramafic rocks are low in alkalis to provide common basaltic magmas by simple subtraction, but they can yield high-pressure picritic liquids capable of fractionating during uprise to produce low-pressure basaltic types at the surface. Trace-element contents of ultramafic rocks and the upper mantle are not well determined. Models for the upper mantle are based on mineral facies determined experimentally for appropriate compositions, temperatures and pressures. Review of experimental data for the system peridotite-gabbro-water, correlated with geophysical data, suggests that the upper mantle is composed of peridotite with layers and pods of eclogite and residual or precipitated dunite and peridotite; with increasing depth, feldspathic peridotite (rarely occurring) is transformed to spinel peridotite and this to garnet-peridotite, at 50-75 kms depth. Garnet peridotite persists down to the Transition Zone at 350-400 kms, where pyroxene is dissolved by garnet, and olivine is transformed to a spinel-like phase. Eclogite is abundant from 80 to 150 kms depth. The low velocity zone begins at 75-100 kms where hornblende becomes unstable, producing traces of interstitial hydrous magma of andesitic composition within eclogite; within a peridotite mantle, interstitial hydrous magma may be alkalic (potassic).

### INTRODUCTION

The physical properties of the earth and its interior are becoming closely defined as geophysical methods are improved, but chemical and mineralogical models are less firmly based because we have no direct information about the chemistry of the Earth's interior. There are three main approaches for estimating the composition and mineralogy of the Earth:

1. The use of data from extraterrestrial sources, and the formulation of physical and chemical models for the origin of the solar system and the Earth.
2. Study of the chemistry and mineralogy of igneous rocks derived from the mantle.
3. Comparison of the variations of density and other physical properties measured for the mantle by geophysical methods with experimental values for appropriate materials determined in the laboratory at high pressures and temperatures. This approach involves the results of static compression and shock wave experiments, and the formulation of a seismic equation of state. In this paper we are concerned only with the upper mantle down to a depth of 400 kms and with the relevant static high pressure studies.

Results from the first approach using the chondrite model indicate that the overall composition of the mantle is ultrabasic although consideration of the chemistry of chondrites and achondrites has led to the conclusion that the upper mantle, at least, is gabbroic in composition—eclogite in mineralogy (Lovering, 1958). The second approach, the petrological approach, is concerned with mantle-derived rocks. By analogy with the meteorite models ultramafic rocks are logical candidates, and it has long been known that the physical properties of the mantle are satisfied by ultramafic rocks such as dunite, peridotite, and eclogite. The rocks selected as representatives of the mantle must have concentrations of the heat-producing radioactive elements K, Th, and U, which are adequate to

account for surface heat flow. It is well established that basaltic magmas are the products of partial fusion of mantle material, and the study of gabbros and basalts places limits on the chemistry of the mantle. Extraterrestrial material thus provides estimates for overall mantle composition, and terrestrial rocks provide information about the upper mantle composition and mineralogy.

Press (1969) recently reported the first independent determination of density in a portion of the suboceanic mantle, using a Monte Carlo procedure to find earth models consistent with a range of geophysical data. He compared the density distributions with densities computed by Clark and Ringwood (1964) for two mantle models involving a hypothetical peridotite (pyrolite) or an eclogite and found only the eclogite model was consistent between 80 and 150 kms depth. Above and below this depth interval, either model is acceptable, but the peridotite model is weakly favored in the region near 300 kms depth. Press concluded that between 80 and 150 kms depth, a suboceanic mantle composed of peridotite with about 50 percent eclogite was required. The geophysical data confirm petrological inferences that both ultrabasic and basic material must be considered as constituents of the upper mantle. Given working values for the composition of the mantle, the mineralogy at various depths can be estimated from phase equilibrium studies at high pressures and temperatures using whole rocks, silicate systems, or germanate model systems. The phase changes determined for the materials selected involve changes in physical properties which must correlate with the mantle properties determined by geophysical methods.

### ESTIMATES OF MANTLE COMPOSITION BASED ON EXTRATERRESTRIAL ROCKS

The study of meteorites has yielded a large quantity of data. Comparison of the chemistry, mineralogy, and petrol-

TABLE 1. ESTIMATES OF MANTLE COMPOSITION FROM METEORITE ANALOGIES, AND EARTH ORIGIN MODELS

| Wt.<br>%                       | Mason<br>1966 | Ringwood<br>1966 |
|--------------------------------|---------------|------------------|
| SiO <sub>2</sub>               | 48.09         | 43.25            |
| MgO                            | 31.15         | 38.10            |
| FeO                            | 12.71         | 9.25             |
| Fe <sub>2</sub> O <sub>3</sub> | —             | —                |
| Al <sub>2</sub> O <sub>3</sub> | 3.02          | 3.90             |
| CaO                            | 2.32          | 3.72             |
| Na <sub>2</sub> O              | 1.13          | 1.78             |
| K <sub>2</sub> O               | 0.13          | —                |
| Cr <sub>2</sub> O <sub>3</sub> | 0.55          | —                |
| MnO                            | 0.43          | —                |
| P <sub>2</sub> O <sub>5</sub>  | 0.34          | —                |
| TiO <sub>2</sub>               | 0.13          | —                |
| Total                          | 100.00        | 100.00           |

ogy of meteorites with terrestrial igneous rocks should eventually yield an acceptable model for Earth formation and differentiation, but at present the application of meteorite analogies to calculations of Earth composition remains speculative. The bulk composition of the Earth is estimated from meteorite analogies by assigning compositions to the core and mantle, whose masses are known from geophysical measurements, and combining these in the appropriate proportions. The results obtained thus depend upon the model adopted for the origin and differentiation of the Earth. This is not the place for a review of Earth models but estimated mantle compositions are required for comparison with estimates from terrestrial rocks.

Two estimates of mantle composition are shown in Table 1. Mason (1966a) assumed that the mantle and crust together have the same composition as the silicate portion of bronzite chondrites, and that the core is composed of the nickel-iron portion of these meteorites together with the troilite. Ringwood (1966a) assumed that the composition of the Earth is represented by the non-volatile elements of the carbonaceous chondrites, and an early high-temperature event caused the loss of volatile elements, reduction of oxides, and the formation of the core; the core incorporated silicon. The two estimates of core composition thus differ with respect to the light element incorporated in the core, Mason's model requiring sulfur, and Ringwood's requiring silicon. Because the two models differ, the estimates of mantle composition also differ, but they agree with respect to the following:

1. More than 90 percent by weight of the mantle is represented by the system FeO-MgO-SiO<sub>2</sub>. No other oxides exceed 4 percent by weight of the mantle.
2. More than 98 percent by weight of the mantle is represented by the system FeO-MgO-SiO<sub>2</sub>-CaO-Na<sub>2</sub>O-Al<sub>2</sub>O<sub>3</sub>; together, the components CaO-Na<sub>2</sub>O-Al<sub>2</sub>O<sub>3</sub> lie within the range 6 to 10 percent by weight of the mantle.
3. No other oxide reaches a concentration of 0.6 percent of the mantle. All other oxides together total 1.6 percent

according to Mason, and even less according to Ringwood.

FeO substitutes for MgO in most magnesian minerals, and Na<sub>2</sub>O substitutes for CaO in plagioclase at low pressures; at high pressures, Na<sub>2</sub>O is accommodated along with CaO in jadeitic pyroxenes. Therefore, the mineralogy of this material can be represented in simplified form in the quaternary system CaO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>. The estimated mantle compositions listed in Table 1 are equivalent to ultramafic rocks, and we turn next to consideration of ultramafic rocks exposed at the Earth's surface. However, we should note that the validity of chondritic earth models is debatable. Studies of strontium isotopes, K/Rb ratios, and the abundance ratios of U, Th and K in meteorites and terrestrial rocks indicate that the earth's chemical composition may be closer to that of basaltic achondrites than to chondrites (Gast, 1960, 1965; Tilton and Reed, 1963; Wasserburg *et al.*, 1964), but Hart and Aldrich (1967) have pointed out the problems involved in interpretation of trace element results when the mineralogy of the source region is uncertain.

#### PERIDOTITE MINERALOGY, FIELD ASSOCIATIONS AND PETROGENESIS

The terms "ultramafic" and "ultrabasic" relate, respectively, to mineralogical and chemical classification of rocks, but both have been used rather loosely. Ultramafic rocks are defined as those with color indices of more than 70, and ultrabasic rocks are those containing less than 45 percent SiO<sub>2</sub> (Williams *et al.* 1954). Most ultramafic rocks are ultrabasic and most ultrabasic rocks are also ultramafic, but there are exceptions; and an ultrabasic magma (compositional usage), if it existed, could not be ultramafic (mineralogical usage). Therefore, both terms have specific uses and should be retained.

There have been several recent publications providing a background for this review. The two-volume treatise *Basalts* (Hess and Poldervaart, 1967, 1968) surveys in detail the occurrence and petrogenesis of the liquid extract from the mantle. Wager and Brown (1967) published a comprehensive work, *Layered Igneous Rocks*, which covers ultramafic cumulates from basaltic magmas. I edited a book on *Ultramafic and Related Rocks* (Wyllie, 1967a) which presents a framework for classification and discussion, and surveys the petrogenesis of ultramafic rocks in all of the associations where they occur. A session on "The origin of basaltic and ultramafic rocks" was included in the symposium on *Deep-seated Foundations of Geological Phenomena* at the 23rd International Geological Congress in Prague, August 1968. A review paper on ultramafic rocks (Wyllie, 1969a) was submitted in April 1968 and distributed by the symposium chairman H. Kuno to den Tex, Maxwell, Challis, Peters, Jackson, Rost, and Forestier, who were invited to present discussion papers at the symposium. The review and discussions were published together in 1969, except for Forestier's discussion, which ap-

peared only in abstract form in the Program (Forestier, 1968).

*Mineralogy of peridotites.* The main features of the mineralogy of ultramafic rocks are illustrated in Table 2. Each major mineral can form an essentially monomineralic rock, and combinations of the minerals in various proportions produce members of the peridotite clan. For upper mantle studies we are most interested in the peridotites composed largely of olivine and pyroxenes. The alumina in these rocks is distributed among the pyroxenes and accessory minerals such as plagioclase, spinel, and garnet. Small amounts of water are likely to be accommodated in amphiboles under upper mantle conditions with serpentine being produced in the presence of water under crustal conditions at lower temperatures. Feldspathoids and phlogopite are among the accessory minerals in the alkalic ultrabasic rocks.

It was noted by Green and Ringwood in 1963 that peridotites with sufficient alumina displayed four distinct mineral assemblages: (1) olivine+Al-poor pyroxenes+plagioclase (2) olivine+aluminous pyroxenes±spinel (3) olivine+Al-poor pyroxenes+pyrope-rich garnet and (4) olivine+amphibole. These different assemblages are exhibited by rocks of quite similar compositions. Examples cited have CaO ranging from 2.03 to 2.97 percent, and Al<sub>2</sub>O<sub>3</sub> ranging from 2.41 to 3.99 percent (Ringwood, 1966b; compare Table 1). They concluded that the first three mineral assemblages were controlled by the pressure-temperature conditions of their formation and not by differences in bulk composition. This implied that large-scale mineralogical zoning could be expected in the upper mantle if the chemistry was similar to that indicated by chondrite models (see Table 1 for chondritic 1966 models), and experimental studies were designed to test this possibility. The existence of peridotite mineral facies in which the aluminous accessory mineral changes from plagioclase to spinel to garnet with increasing pressure was subsequently established as shown in Figure 9.

*Ultramafic and ultrabasic rock associations.* Ultramafic and ultrabasic rocks occur in a variety of field and petro-

TABLE 2. MINERALOGY OF ULTRAMAFIC ROCKS

| Major Minerals | Rock Types   | Accessory Minerals |
|----------------|--------------|--------------------|
| Olivine        | Dunite       | Garnet             |
| Clinopyroxene  | Pyroxenite   | Spinel             |
| Orthopyroxene  |              | Feldspar           |
| Hornblende     | Hornblendite | Feldspathoids      |
| Biotite        | Biotitite    | Phlogopite         |
| Serpentine     | Serpentinite | Melilite           |

TABLE 3. CLASSIFICATION OF MAIN ULTRAMAFIC ROCK ASSOCIATIONS

| Ultramafic Associations                                      | Tectonic Environments                            | Source of Rock                       |
|--|--|--------------------------------------|
| 1. Layered intrusions  | Nonorogenic<br>Varied                            | Basalt magma from mantle             |
| 2. Alkalic complexes, Kimberlites                            | Cratonic   | Magma from mantle                    |
| 3. Alpine-type intrusions<br>(a) ophiolites<br>(b) root-zone | Orogenic<br>Island arcs                          | Magma and solids from mantle         |
| 4. Oceanic rocks   | Ocean trenches<br>Ocean floors<br>Oceanic ridges | Mantle or oceanic crust              |
| 5. Nodules<br>(a) alkali olivine-basalts<br>(b) kimberlites  | Varied   | Mantle, crust, or magma concentrates |

graphic associations, and we may expect, therefore, that there is a variety of processes involved in their origin and emplacement. The petrological problem is to determine which ultramafic rocks, if any, represent mantle material and to deduce the processes which led to their emplacement into the crust. Although interpretation of processes may require extensive study, rock associations are more easily distinguished, and this is an essential first step. In recent reviews I have outlined eleven petrographic associations, some with subdivisions, as a working classification (Wyllie, 1967b, 1969a, 1969b). I also noted that for petrogenetic discussions relevant to the composition of the upper mantle it was convenient to consider four groups of associations which are summarized in Table 3. These are:

1. Layered, stratiform and other intrusions involving gabbro or diabase together with accumulations or concentrations of mafic minerals. These occur in varied tectonic environments, but they are rarely affected by contemporaneous orogeny.

2. The alkalic rocks of stable continental regions, including kimberlites, mica peridotites, members of ring complexes, and ultrabasic lava flows. These usually occur in stable or fractured continental regions, and their distribution in belts appears to be controlled by deep-seated tectonics with linear trends.

3. The several serpentinite-peridotite associations of the orogenic belts that have been classified together as alpine-type intrusions. These include large and small bodies distributed along deformed mountain chains and island arcs, usually along with gabbros or basic volcanic rocks. Relationships among the rocks of this association are often complicated by metamorphism and metasomatism.

4. Serpentinites and peridotites of the oceanic regions. These have been identified on mid-oceanic ridges, within deep ocean trenches, and where small scarps expose material below the cover of basalts forming the ocean floor.

In the present context, with the nature of the upper mantle as our objective, a fifth group must be included:

5. Peridotite and eclogite nodules in alkali olivine-basalts and kimberlites.

Criticisms of my working classification of associations (Wyllie, 1969a), and suggested alternatives, were presented by den Tex (1969), Peters (1968a), and Forestier (1969). The main criticism involved my treatment of peridotites in the orogenic belts, grouped together above under (3). I subdivided them into six classes on the basis of the metamorphic grade of the country rock, or specific features of the igneous bodies, and including a class of metasomatic or metamorphic rocks. Forestier (1968) and den Tex (1969) considered that the primary subdivision of the orogenic peridotites should be into the ophiolitic, or truly Alpine-type, and the "root-zone" peridotites, occurring in crystalline rocks. I have adopted this sub division as groups 3(a) and 3(b), respectively, in Table 3. Forestier (1968) and Rost (1969) objected to application of the term "alpine-type" to rocks older than the Alpine orogenesis, often to rocks of Precambrian age. However, the term has been so widely used by non-Alpine geologists to describe the ultramafic rocks in orogenic zones that it is probably firmly established in this context.

*Petrogenesis: process involved in the origin and emplacement of ultramafic rocks.* Before attempting to extrapolate back to the presumed mantle sources of any of these groups

of rocks, we should consider the variety of processes that can be involved in their origin and emplacement. Figures 1, 2, and 3 illustrate in diagrammatic fashion some aspects of the associations of ultramafic rocks listed in Table 3 and serve as a basis for a review of processes.

Figure 1 summarizes the major processes involved in groups 1, 3, and 4 of Table 3: the petrogenesis of associations including peridotite and gabbro, picrite and diabase, or picrite-basalt and basalt in nonorogenic and orogenic regions. The general model adopted for magma generation corresponds to that described by Green and Ringwood (1967a). In both environments, the process begins with diapiric uprise of mantle (aluminous peridotite) to a level where partial melting produces interstitial basaltic magma. The crystal mush continues to rise through the mantle. In the left-hand diagram, representing nonorogenic regions, magma separation from the mush is depicted within the mantle, and basaltic magma rises into a large reservoir, or magma chamber, within the crust. Crystal settling from this basic magma produces a series of ultramafic cumulates, and a stratiform intrusion (Jackson, 1967). The magma flows over the cumulates and upwards out of the chamber, producing dykes, sills, volcanic necks, volcanoes, and lava flows (Brown, 1956; Challis, 1965a; Irvine and Smith,

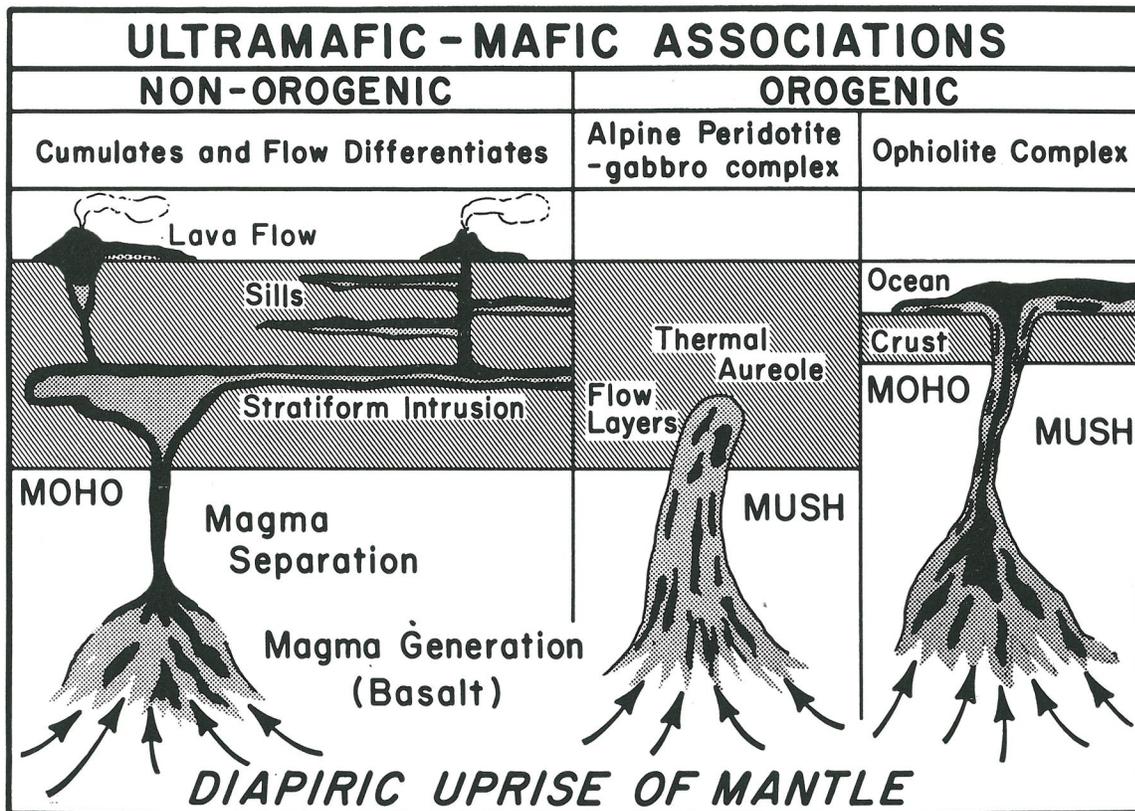


FIG. 1. Schematic representation of the processes involved in the origin and emplacement of associations of ultramafic and mafic rocks. See text for explanation. Stipple: crystalline ultramafic material; black associated with stipple: interstitial basic liquid or crystalline basalt or gabbro.

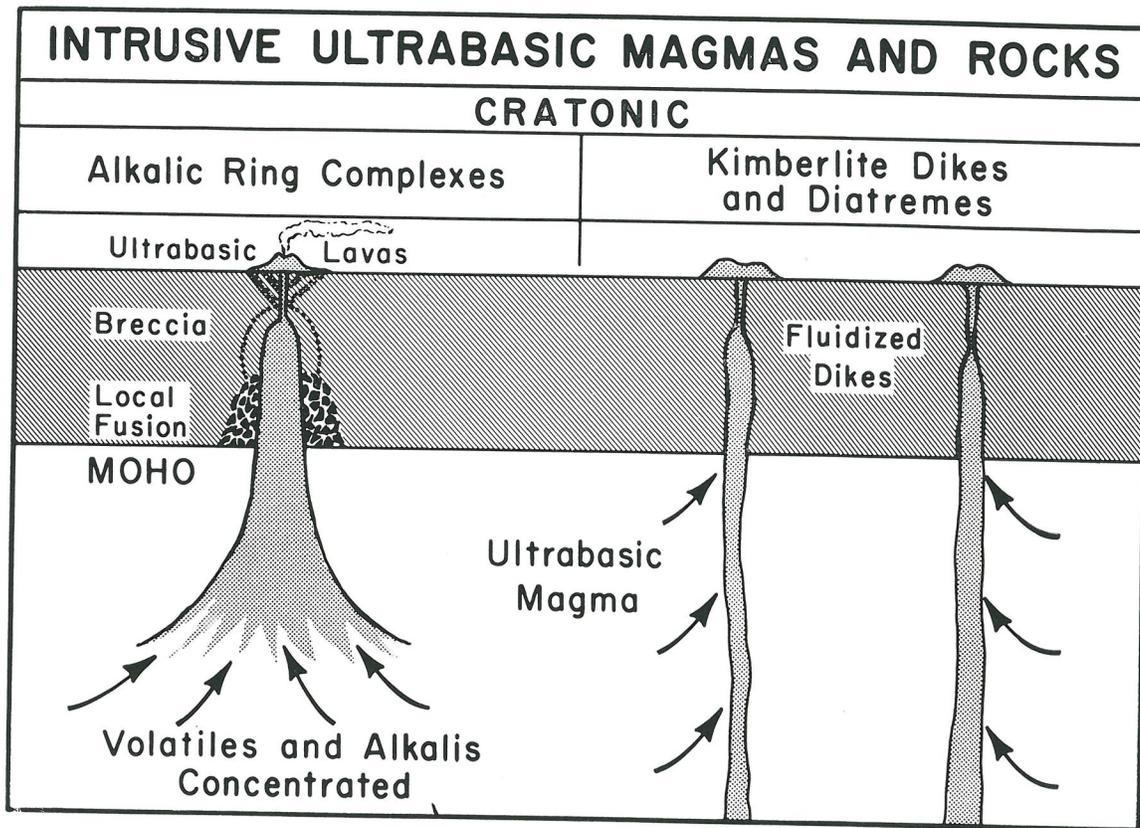


FIG. 2. Schematic representation of the processes involved in the origin and emplacement of intrusive ultrabasic magmas and rocks. See text for explanation. Stipple: ultrabasic material

1967). Crystal settling, flow differentiation (Simkin, 1967), and other processes of crystal concentration may yield ultramafic rocks in the general locations shown in the figure. These ultramafic rocks correspond to group 1 in Table 3 (see Wager and Brown, 1967).

In the middle diagram of Figure 1, representing the root-zone peridotites of orogenic regions (group 3b of Table 3), the ultrabasic crystal mush rises into the crust, under conditions permitting little physical separation of basaltic magma. Flow layering develops with characteristics distinct from that of ultramafic cumulates in stratiform intrusions (Thayer, 1960, 1967).

The right-hand diagram of Figure 1 represents the ophiolite complexes of group 3a in Table 3, where the crystal-mush breaks through the oceanic crust and produces a thick peridotite-gabbro-basalt complex on the ocean floor. Gravity slides occurring on regional scales may produce chaotic structures (Maxwell, 1968; 1969a; 1969b). Although gravity settling of crystals certainly occurs in ophiolite complexes, the processes involved for the formation of ultramafic rocks in the orogenic and nonorogenic environments are distinct. However, if the ultramafic cumulates and crystal concentrates depicted schematically in the left-hand diagram of Figure 1 become involved in orogenic processes, then the distinction may become obscured,

as indicated in Table 4.

Ultramafic and ultrabasic rocks are not always obviously associated with mafic rocks, and Figures 2 and 3 illustrate schematically the processes involved in the intrusion of ultrabasic magmas and rocks into the crust. The alkalic ultrabasic rocks of stable continental regions (group 2 of Table 3) are represented schematically in Figure 2. The main feature to note is the concentration of alkalis and volatiles along with the generation of ultrabasic magma. The magma rises in dyke, or pipe-like form into the crust and, eventually, ultrabasic lavas are erupted at the surface. All of the processes involved in the layered rocks of group 1 (the left-hand diagram of Fig. 1) operate also in the alkalic ring complexes, with gravity settling of crystals and convective circulation being dominant in the early stages of crystallization. The alkalis and volatiles become concentrated in residual magmas, and are largely responsible for many of the interesting petrological and mineralogical features of the association. Exsolution of volatile components causes brecciation, and explosive eruption of fragmental volcanic rocks. This effect is well illustrated by kimberlites: the deep-seated massive rock crystallizes from an ultrabasic magma, and higher level rocks are emplaced in diatremes as fluidized solid-gas systems, xenocrysts and xenoliths of wall rocks being mixed with the fragmented and

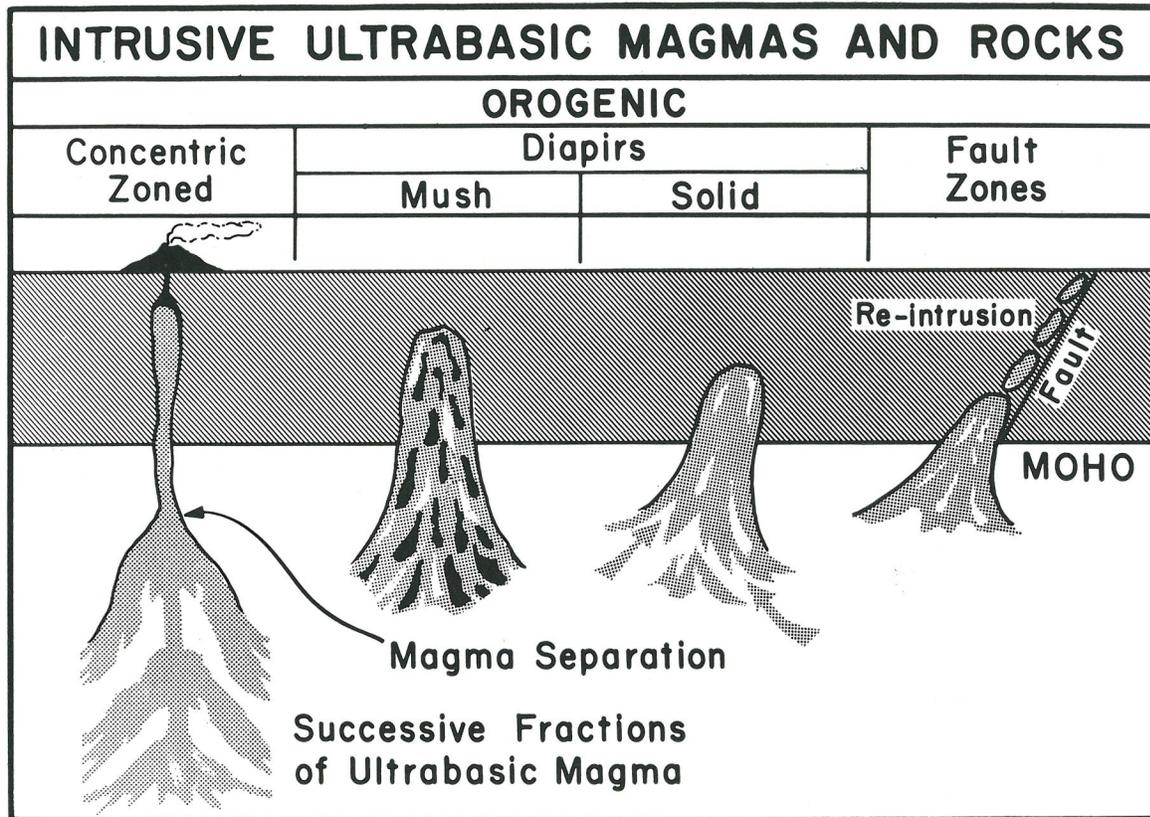


FIG. 3. Schematic representation of the processes involved in the origin and emplacement of intrusive ultrabasic magmas and rocks. See text for explanation. Stipple: ultrabasic magma or crystalline ultramafic material; black associated with stipple: interstitial basic liquid.

usually greatly altered kimberlite (Dawson, 1967b). The explosive effects, and the metasomatic effects of alkalis and volatiles in the residual magmas and exsolved gases and solutions, tend to obscure the early stages of petrogenesis. The petrogenesis of these rocks has been reviewed by Dawson (1967a, 1967b, 1968), Davidson (1967a), Watson (1967), Upton (1967), Gold (1967), von Eckermann (1967), and Kennedy and Nordlie (1968).

Processes involved in the intrusion of ultrabasic magmas and rocks in orogenic zones, group 3b of Table 3, are depicted in Figure 3. The left-hand diagram represents the remarkable concentrically zoned ultramafic rocks of the Urals and Alaska, which are believed by Taylor and Noble (1960; also Taylor, 1967) to result from the successive intrusion of liquid ultrabasic magmas derived from the mantle, with high contents of FeO, CaO, and H<sub>2</sub>O (Presnall, 1966). Crystal cumulates, and evidence for convection cells within the ultrabasic magma, confirm that processes operating are similar to those in group 1. Irvine (1967) and Taylor (1967) cited evidence that these plutonic magma chambers were connected to volcanoes. The second diagram, repeated from Figure 1, shows intrusion of a diapiric crystal mush of partly fused ultrabasic mantle, crystalline peridotite with interstitial basaltic liquid (Green, 1967).

Some bodies of ultramafic rock in this environment have been interpreted as diapiric intrusion of solid peridotite directly from the mantle as depicted in the third diagram (de Roever, 1957; Ragan, 1967; Lappin, 1967; Carswell, 1968a, 1968b; Rost, 1969; den Tex, 1969). The right-hand diagram is a representation of the reintrusion of solid peridotite or serpentinite in fault-bounded blocks (Jahns, 1967; Lapham, 1967; Ragan, 1967; Raleigh, 1967), which apparently can lead to extrusion of serpentinite breccias by plastic flow (Dickinson, 1966).

The processes involved in the formation of the nodules of group 5 in Table 3 can be related to the left-hand diagram of Figure 1, and to Figure 2. The nodules in kimberlites may represent fragments carried upwards from various levels in the mantle, or from metamorphic rocks of the deep crust. Nodules in basalts may represent fragments of the mantle, or fragments of crystal cumulates or flow concentrates from a magma chamber or the walls of a conduit or volcanic neck; these different sources for nodules can be located in Figure 1.

*Extrapolation to the upper mantle.* If we are able to unravel the petrogenesis of an ultramafic rock, or of an ultramafic-mafic rock association, and to conclude that the

rocks do represent mantle material, we still face the problem that the upper mantle is heterogeneous. According to the processes illustrated in Figures 1, 2, and 3, portions of the mantle have been depleted in the components that passed into a liquid phase during partial melting. The depletion will vary according to several factors; the original composition of the mantle, the depth (pressure) at which fusion occurred, the extent of partial melting, which is a function of temperature, and the effectiveness of magma separation from the mantle material. We may therefore expect that the upper mantle will include ultramafic rocks ranging in composition from original, primitive, undifferentiated mantle to residual peridotite or dunite from which the basaltic fraction has been partially or completely removed. It will also include layers and lenses representing basaltic or picritic magma derived by partial melting of the mantle, which failed to reach the crust and crystallized within the mantle as eclogite (Press, 1969). It may also include portions of the original peridotite somewhat enriched in eclogitic components, by percolation of basaltic magma and its local concentration in dispersed form, in favorable sites (Carswell, 1968a). Once we have traced an ultramafic rock to a mantle source, its composition and mineralogy have still to be related to the history of that portion of the mantle from which it was derived. This is not an easy problem.

We may now consider all five groups of ultramafic associations listed in Table 3 and consider which of these are most likely to provide information about the upper mantle. The ultramafic rocks of group 1, in stratiform and layered intrusions and extrusions (Fig. 1), are produced by the concentration of mafic minerals crystallized from basaltic magma. They can provide only indirect evidence of mantle chemistry. The kimberlites and other alkalic rocks of the cratonic ultramafic belts in group 2 do not provide direct representatives of the mantle because they have been strongly fractionated. The high concentrations of volatile components and alkalis obscure many details of their petrogenesis, and their chemical relationships to the upper mantle. However, the xenoliths carried upwards in kimberlite diatremes probably include mantle samples (group 5).

The ultramafic rocks of the orogenic zones, groups 3a and 3b, probably include representative mantle material, as depicted in Figures 1 and 3, and these rocks have been cited often as guides to mantle composition. Unfortunately, the effects of metamorphism and reintrusion (Fig. 3) tend to blur the petrogenesis of these associations. Ultramafic rocks in this group may have been subjected to several periods of metamorphism, deformation, plastic flow, serpentinization and tectonic block transportation. Therefore, peridotites of any origin that have become involved in the orogenic process may be difficult to distinguish from those of alpine-type that are contemporaneous with the orogenic cycle.

The ultramafic rocks of the oceanic regions (group 4 in Table 3), especially those of the mid-oceanic ridges, may provide good prospects for correlation with upper mantle

material (Hess, 1964; Melson *et al.*, 1967; Thayer, 1969b), but we need more information about their associated rocks and geological structures, and many more specimens.

The ultramafic nodules, group 5 of Table 3, may provide the most direct evidence of mantle mineralogy and chemistry. However, even if distinction can be established among nodules that are mantle-derived, those that are derived from the crust, and those that are crystal cumulates or concentrates, there remain other problems of interpretation. These include the extent of depletion of the mantle source in basaltic components, and the extent of reaction between the nodule and its magmatic host during uprise. If it appears that a nodule has residual character, then it may be important to know whether the nodule represents the residual mantle from which its host magma was derived, or whether it is a residual fragment plucked from the mantle at some higher level and quite unrelated to this magmatic episode. Distinction among these varied types of nodules should be made by textural and chemical means before they can be related to the mantle (White, 1966; den Tex, 1969; Jackson, 1969).

Estimates of the composition and mineralogy of the upper mantle may thus be provided by ultramafic rocks in groups 3, 4 and 5 in Table 3, but for none of these groups can it be assumed automatically that a specific rock has been derived from the upper mantle. The petrogenetic processes involved are varied and complex (Figs 1, 2, and 3), and in our discussion of estimates of mantle composition based on terrestrial ultramafic rocks, the field association and petrogenesis of the rocks should be thoroughly investigated for satisfactory extrapolation to the mantle. Therefore, we next consider in more detail the petrogenesis of these three groups of ultramafic rocks.

#### PETROGENESIS OF ULTRAMAFIC ROCKS OF OROGENIC BELTS (ALPINE-TYPE)

Although ultramafic rocks located in orogenic belts have often been considered together as alpine-type, they include rocks with a variety of origins and histories. If we are to extrapolate down to the upper mantle from the study of rocks now exposed at the surface, it is necessary first to decipher these histories. There is no general agreement about the petrogenesis of alpine-type rocks, and the purpose of this section is to demonstrate the diversity of opinions.

The petrogenesis is complex because it involves metamorphic processes; indeed, den Tex (1965) discussed these rocks in terms of their metamorphic lineages rather than igneous origins. The problems of interpretation introduced by metamorphism, deformation, and reintrusion have contributed to what Hess in 1955 (p. 391) termed a "magnificent argument" between the field geologists, who concluded that many alpine-type ultramafic bodies were emplaced as fluid ultrabasic magmas, and the experimental petrologists, who denied that such materials could be liquid magmas at any reasonable temperatures. I reviewed the history of this argument in the light of new experimen-

tal data and changing field interpretations since 1955, (Wyllie 1967b), and its resolution in favor of solid emplacement was discussed by Hess (1966, p. 5-6). However, Osborn (1969) has since doubted alpine peridotite complexes having a specific gravity of about 3.3 could be squeezed or otherwise moved upwards for tens of kilometers through geosynclinal rocks of much lower average density. Maxwell (1969b) proposed that the mechanism is diapiric uprise of plastic rocks driven by differential specific gravity above some level of compensation within the upper mantle; this level could be as deep as the low-velocity zone.

The controversy concerning ultrabasic magmas tended to ignore the gabbros and basic volcanic rocks almost invariably associated with the peridotites and serpentinites, or to relate them to a different stage of the orogenic and magmatic activity. Thayer (1960, 1967) discussed and deplored the conceptual divorce of the ultramafic and mafic rocks, and he listed six criteria that characterize the intrusive peridotite-gabbro complexes of alpine type. Reconsideration of the petrogenesis of ultramafic and mafic rocks together (Miyashiro, 1966) has contributed to the formulation or revival of several hypotheses.

In Table 3, the ultramafic rocks of orogenic belts are subdivided into the ophiolite suite and the root-zone peridotites, following Forestier (1968) and den Tex (1969). In 1967a and 1969a, I proposed a more detailed subdivision based on the metamorphic grade of the host rocks; this has some advantages, but also disadvantages discussed by Peters (1968a) and Forestier (1968). Vuagnat (1968) noted that most examples of ultramafic rocks considered in the development of the idea of an alpine-type were not in the Alpine mountain chain proper, that is, in the Mediterranean region. In particular, the ophiolite sequence so spectacularly developed in the Mediterranean Alps and extending through to the Himalayas contributed little to the concept of alpine-type ultramafic rocks, receiving little attention from petrologists other than Alpine geologists until the present decade. Thayer (1967) presented reasons for concluding that ophiolite complexes are essentially identical with normal intrusive peridotite-gabbro alpine complexes, but according to Vuagnat (1968), the ophiolites of the Alps depart in significant features from the plutonic complexes characterized by Thayer. Mafic rocks in the ophiolite sequence typically include spilites, and the association of spilites and ultramafic rocks with chert is known as Steinmann's trinity.

*The ophiolite suite (Group 3a, Table 3).* A representation of the processes involved in the petrogenesis of the ophiolite suite is given in the right-hand diagram of Figure 1. There are four main types of interpretation for the origin of the associated ultramafic and mafic rocks. Three involve the extrusion of very large volumes of material in submarine eruptions, and later development of chaotic structures through gravity slides. The fourth involves metamorphic differentiation and metasomatism.

1. Extrusion of essentially liquid basaltic magma, which undergoes differentiation on the sea floor.

2. Extrusion of a partially molten diapir of rising mantle with fractional separation of basic magma during and after uprise, and continued differentiation of the massive pile on the sea floor.

3. A two-stage process, with solid intrusion of peridotite and gabbro from the mantle during a pre-geosynclinal period being followed by intrusion and extrusion of basaltic magma at intervals through the whole geosynclinal cycle.

4. Basaltic lavas and sediments are metasomatically transformed into peridotites and serpentinites.

*Root-zone peridotites (Group 3b, Table 3).* The processes involved in the petrogenesis of alpine-type rocks enclosed within crystalline rocks are depicted in Figure 3. The five main types of interpretation for the origin of these rocks are:

1. Intrusion of ultrabasic magma.

2. Intrusion of a crystal mush of partially melted mantle.

3. Diapiric or tetronic intrusion of solid mantle peridotite, or of peridotite plus gabbro.

4. Reintrusion of peridotite or serpentinite from any source.

5. Metasomatism and metamorphism of basic lavas and dolomites.

*Review of current petrogenetic hypotheses (Table 4).* There are ultramafic rocks in orogenic belts that are not readily accommodated into the conventional classification of groups 3a and 3b of Table 3, which is based on the present environment of the ultramafic rocks.

Vuagnat (1968) noted that although two groups of ultramafic rocks are recognized in the Western Alps, corresponding to groups 3a and 3b in Table 3, this does not preclude the possibility that other rocks were parts of ultramafic complexes in the old basement, which became reworked by a new orogenesis. A major question raised by Challis (1965a, 1965b, 1969), O'Hara (1967c), and Osborn (1969) is the extent to which ultramafic rocks of cumulate origin, depicted in the left-hand diagram of Figure 1, have become involved in the orogenic process, and thus converted into rocks which merit the appellation "alpine-type". O'Hara (1967c) suggested that the only feature that rocks designated "alpine-type" have in common is an alpine setting that causes tectonic transportation and re-intrusion. Therefore, I have developed in Table 4 a summary review of current hypotheses based primarily on the inferred processes occurring in the mantle and during emplacement into the crust, and the nature of the material entering the crust. For successful extrapolation from exposed ultramafic rocks to the upper mantle, it is this early stage of petrogenesis that must be understood.

The organization of Table 4 is similar to earlier versions (Wyllie, 1967b, 1969a). Varying degrees of fusion of the mantle may produce basic liquids, or a range of ultrabasic

TABLE 4. CURRENT PETROGENETIC HYPOTHESES PROPOSED FOR ULTRAMAFIC ROCKS OF OROGENIC BELTS (ALPINE-TYPE)

| Hypothesis |  | Process in mantle             | Material from mantle   | Successive stages in crust   |   |   |
|------------|--|-------------------------------|--|--|---|---|
| Number     | Advocate                               |                               |  | 1  | 2   | 3   |
| I          | A                                      | Taylor 1967                   | Complete(?) or partial fusion  | Ultrapasic liquid magma, with or without suspended crystals          | Successive intrusions (VH)                                | Reintrusion of crystal mushes (H-M)                                   |
|            | B                                      | Onuki 1965                    |  |  |   |   |
| II         | A                                      | Green 1967                    | Partial fusion   | Basaltic liquid magma, with or without suspended crystals            | Gravity differentiation forming stratiform intrusions (H) | Reintrusion of cumulate mush (H-M)                                    |
|            | B                                      | Challis 1965, 1969            |  |  | Differentiation with high oxygen fugacity (H)             | Removal of basic magma above cumulates (H-M)                          |
|            | C                                      | Osborn 1969                   |  |  | Differentiation in magma chambers(?) (H)                  | Andesites erupted above peridotite cumulates (H-M)                    |
|            | D                                      | Brunn 1960                    |  |  |   | Submarine extrusion and differentiation (H-M)                         |
| III        | A                                      | Maxwell 1969                  | Partial fusion, diapiric uprise of mush. Crystallization. Diapiric uprise of solid | Peridotite mush with basic liquid, possibly gravity stratified       | Mush intrusion through oceanic crust (H)                  | Submarine extrusion. Gravity slides. (H)                              |
|            | B                                      | Green 1967<br>Thayer 1967     |  |  | Mush intrusion with flow layering (H)                     | Plastic flow of layered complex (M)                                   |
|            | C                                      | Reinhardt 1969                |  | Solid peridotite and gabbro; later basic magma                       | Solid intrusion of mantle rocks into crust (H-M)          | intrusion of basalt, assimilation, extrusion of spilites (H)          |
|            | D                                      | Carswell 1968<br>den Tex 1969 |  | Solid mixtures of dunite and eclogite                                |   | Upward migration by plastic flow or tectonic block transportation (M) |
|            | E                                      |                               |  | Solid peridotite   |   |   |
| IV         | A                                      | Dietz 1963                    | Downward movement of convection current  | Serpentinized layer 3 of oceanic crust, interpreted as "mantle rind" | Depression and heating of serpentinite (L-M)              | Tectonic emplacement in overlying sediments (L)                       |
|            | B                                      | Hess 1960                     |  | ?  | Uplift (L)  | Erosion exposing mantle (L)   |
| V          | Avias 1955<br>Perrin and Roubault 1955 |                               |  |  | Metasomatism of basic rocks, dolomites, or marls (L)      |   |
| VI         | O'Hara 1967<br>Vuagnat 1968            |                               |  |  | Metamorphism of pre-existing ultramafic rocks (L-M)       |   |

VH—very high temperatures above basalt liquidus. H—high temperature within crystallization interval of anhydrous basalt. M—medium temperatures between anhydrous basalt solidus and serpentine stability. L—low temperatures in serpentine stability field.

liquid compositions. The hypotheses involving ultrabasic liquid magmas form group I. Hypotheses involving the formation of ultramafic rocks in the crust by concentration of mafic minerals from a basic liquid form group II. Group III includes the hypotheses suggesting that ultramafic rocks, or peridotite-gabbro associations, constitute samples of a portion of the mantle, although the sample may have been partially melted and gravity stratified before intrusion in mush or crystalline state. Group IV includes hypotheses relating serpentinites and peridotites of the ocean floors to the orogenic cycle. In group V are included all hypotheses involving the metasomatic transformation of preexisting rocks to ultramafic rocks. Group VI refers to ultramafic rocks of any origin that have been remetamorphosed, and have become an integral part of a terrane of regional metamorphism. One or two authors are listed as advocates of each hypothesis summarized in the table.

The columns summarizing the process occurring in the mantle and the nature of the material rising from the mantle, for each hypothesis listed, are followed by a set of three columns representing successive stages within the crust. Stages 1 and 2 are specific for each hypothesis, but stage 3 need not be reached or completed by all rocks of alpine-type. Stage 3 involves reintrusion, serpentinitization, and possibly remetamorphism; all of the late-stage processes which make petrogenetic interpretation of these rocks so difficult.

In parentheses are given the ranges of temperatures of ultramafic rocks at the successive stages of formation and emplacement, upon leaving the mantle and entering the crust. Four temperature ranges are indicated: (VH) very high temperatures above the basalt liquidus for ultrabasic magmas, (H) high temperatures within the crystallization interval of anhydrous basalt, for basic magmas or crystal

(M-L)  
Plastic flow, serpentinitization, tectonic block transport, solid extrusion of serpentinite

mushes lubricated by basic magma, (M) medium temperatures between the anhydrous basalt solidus and the maximum stability temperature of serpentine, related to high grades of metamorphism, and (L) low temperatures within the serpentine stability range for low metamorphic grades.

The zoned ultramafic complexes of Alaska constitute a distinctive group of very high-temperature intrusions with pronounced metamorphic and metasomatic effects at their margins. Taylor's (1967) review concluded that they were formed by the successive intrusion of liquid ultrabasic magmas of different compositions, all with high contents of FeO, CaO, and H<sub>2</sub>O (see Presnall, 1966). This is hypothesis IA. Taylor considers the ultramafic belt in the Ural Mountains, U.S.S.R., to be of the same type. Onuki (1965, 1966) described ultramafic rocks from Japan which are petrographically similar to the Alaskan rocks, and he concluded that intrusion and differentiation of a parent ultrabasic liquid magma produced the "ultramafic rock series" formed from essentially liquid magmas, crystal mushes, and residual basic magmas. This is hypothesis IB.

Green (1967) recently reviewed a distinctive group of peridotites with high-temperature aureoles (Lizard, Cornwall; Mount Albert, Canada; Tinaquillo, Venezuela), and he considered their source to be either deep-seated cumulate or the upper mantle, and their mode of intrusion to be as a crystal mush followed by diapiric flow after complete crystallization. The alternatives correspond to hypothesis IIA, or to hypothesis IIIB (with gabbro removed). Thayer (1967) remained unconvinced that these intrusions were unusually hot when intruded. He also maintained (Thayer, 1967, 1969b) that the peridotite and gabbro at Lizard were comagmatic. O'Hara (1967c) has developed a *P-T* grid based on compositions of coexisting pyroxenes, and representatives of these rocks plotted on the grid give estimated conditions of equilibration at temperatures usually greater than 1100°C, and at pressures corresponding to depths ranging from near-surface to 50 or 60 kms; the conditions correspond closely to the crystallization interval of anhydrous basaltic liquid (Green and Ringwood, 1967a; Cohen, *et al.*, 1967). Challis (1965a, 1965b, 1969) reported pyroxene granulite facies metamorphism at the unfaulted contact of the Red Hills ultramafic intrusion of New Zealand, which has long been classified as a typical low-temperature alpine-type intrusion, and she interpreted the ultramafic rocks in terms of hypothesis IIB. Several other alpine peridotites have recently been interpreted as gravity-stratified crystal cumulates, following hypotheses IIA or IIB (O'Hara, 1967b; 1967c). Smith (1958) suggested that there is a continuous series of peridotite-gabbro associations between the stratiform types and the alpine types. Challis and Lauder (1966) traced similarities indicating a transition from stratiform lopoliths, through trough-like layered complexes and central volcanic complexes, to layered "alpinotype" ultramafic complexes.

Several recent proposals have connected volcanic activ-

ity with stratiform and other gravity-layered intrusions (Brown, 1956; Challis, 1965a; Irvine and Smith, 1967; Irvine, 1967; Taylor, 1967; Osborn, 1969). Comagmatic lava flows could thus be associated with alpine-type ultramafic rocks formed from ultrabasic magmas (hypothesis I), or from crystal concentrates derived from intrusive basic magmas or stratiform intrusions (hypotheses IIA and IIB).

Osborn (1969) reviewed the origin of the calc-alkaline andesites of the orogenic belts and concluded they formed by fractional crystallization of olivine basalt magma under conditions of high oxygen fugacity, produced by flow of water from the surrounding geosynclinal rocks into the magma, and migration away of hydrogen. According to hypothesis IIC, the fractional crystallization of the basic liquid under these conditions produces two complementary fractions; the andesite liquid which is erupted at the surface, and bodies of settled crystals which remain at depth. The crystal cumulates are alpine peridotites although Osborn does not argue that all alpine peridotites are complementary fractions of andesites.

Hypothesis IID is one interpretation of the ophiolite suite, the classical interpretation, with the whole assemblage being considered as a massive, differentiated submarine lava flow ranging in composition from basic to ultrabasic, together with some intrusive rocks (Brunn, 1960).

In hypothesis IIIA, Maxwell (1969a) interprets the ophiolite sequence as a complex formed by extrusion and differentiation of a diapir of partly fused mantle peridotite. Great submarine slides of flysch and broken masses of ophiolite occur, and subsequent regional metamorphism produces streamlined masses and blocks of serpentinite, which have often been interpreted as cold intrusions of solid peridotite or serpentinite.

Thayer (1960, 1967, 1969b) discussed evidence supporting hypothesis IIIB. Anhydrous crystal mushes from the upper oceanic mantle, previously gravity stratified into peridotite and gabbro and later developing flow layering, are emplaced into crustal rocks entraining masses of chromitites; associated quartz diorite and granophyre complete the calc-alkaline "alpine mafic magma stem". This association he regards as the characteristic intrusive peridotite-gabbro complex of alpine type. Thayer (1969a) stated that these intrusive complexes have many features in common with the ophiolitic complexes, and he believes that they are the same. Vuagnat (1968) apparently is not convinced of this.

Hypotheses IIIC, IIID, and IIIE involve the intrusion of solid mantle rocks. The hypothesis of solid intrusion for the emplacement of ultramafic rocks in alpine belts is now widely accepted (de Roever, 1957; Ragan, 1967; Jahns, 1967; Peters, 1969a), and the methods of structural petrology have helped to unravel the processes involved in their emplacement by correlation of deformation patterns in the ultramafic rocks with the regional orogenic history (Lappin, 1967; Lappin, 1967; den Tex, 1969). Recent experiments on deformation have provided valuable information on the mechanisms of plastic flow of peridotite minerals

(Raleigh, 1967), and the experimentally observed weakening and embrittlement of serpentinite at dehydration (Raleigh and Paterson, 1965) provides a satisfactory explanation for the field conclusion that many serpentinites were emplaced by block tectonic transport. However, as noted previously, Osborn (1969) has raised questions about the feasibility of the proposed uprising of dense ultramafic rocks through less dense geosynclinal rocks. Relative deformability may be a more significant factor in controlling movement than relative density.

Hypothesis IIIC is an interpretation of the ophiolite suite by Reinhardt (1969), based on his studies of the Oman Mountains geosyncline. Partial melting and differentiation in the upper mantle was followed by crystallization without intrusion, and solid peridotite and gabbro from the upper mantle were then emplaced during a pregeosynclinal period. He cited Peters' (1968b) analytical data indicating that mineral equilibria in some ophiolitic peridotites may have been attained at depths of the order of 30 kms. These plutonic rocks were then partially assimilated by a younger basaltic intrusive-extrusive suite at intervals throughout the geosynclinal cycle. This produced gradual transitions between the ultramafic and mafic rocks which tends to obscure the genetic hiatus between them. A similar interpretation was proposed by Vuagnat and Cogulu (1967) for the ophiolite complex at Kizil Dag in Turkey.

Carswell (1968a) examined in detail the garnet peridotites at Kalskaret in Norway and explained them in terms of hypothesis IIID. Statistical analysis of bulk rock compositions revealed a strong linear trend between two end members, one an eclogite, and the other a dunite. The garnet peridotites are in dis-equilibrium with the enclosing almandine-amphibolite gneiss (O'Hara and Mercy, 1963). Carswell interpreted the field and textural relationships to indicate tectonic emplacement of the garnet peridotites as relatively cold intrusions. He concluded that within the upper mantle a picritic partial melt fraction was not completely liberated from a dunitic residual fraction of the original mantle material, but became trapped and mixed with the dunite; there it crystallized as eclogite. The mixed rocks were then emplaced into the crust in the solid state. The compositions of the eclogite and dunite end members of the partial melting process are listed in Table 7, analyses 3 and 2, respectively.

According to hypothesis IIIE, solid emplacement of mantle peridotite yields ultramafic rocks of alpine type. Den Tex (1969) presented petrofabric and metamorphic data for a root-zone peridotite of the Swiss Pennine Alps, and for the ophiolitic peridotite of the French Pyrenees. Both rocks had experienced several metamorphic events not represented in the fabrics of their host rocks. This indicates that they were emplaced in the solid state. Den Tex concluded that solid derivation from upper mantle material, not depleted in basaltic components, is the most likely origin for the rocks.

According to Hess' (1962) concept of sea floor spreading, layer 3 of the oceanic crust is composed of mantle peridotite

which became serpentinized beneath the midoceanic ridges, and was then transported laterally above a convection cell within the solid mantle. Dietz (1963) proposed that alpine-type peridotites would be produced by the tectonic incorporation of this mantle rind within sediments of the continental rise which constitute a eugeosyncline. This is hypothesis IVA. Similarly, in hypothesis IVB, Hess (1960) suggested that the serpentinites of Puerto Rico represented uplifted oceanic crust of mantle derivation.

It has been proposed that ultramafic rocks in the ophiolite sequence were produced by metasomatism of basic rocks (Avias, 1955) or of dolomites or marls (Perrin and Roubault, 1955). Sørensen (1967) reviewed these and similar proposals, and concluded that it was doubtful if the large masses of alpine-type ultramafic rocks were formed by metasomatic processes. However, he presented evidence that peridotite and garnet peridotite, when enclosed in amphibolite, are the products of metamorphic differentiation in zones of stress concentration.

O'Hara's (1967c) pyroxene grid indicates that alpine-type ultramafic rocks have equilibrated under a wide range of conditions, and some high temperature ultramafic rocks have been reentruded upwards through several kilometers. Hypothesis VI implies that a peridotite of any origin becomes an alpine type if it is involved in regional metamorphism with the associated processes of deformation being represented by stage 3. Vuagnat (1968) drew attention to this in the context of the Western Alps. Carpenter and Phyfer (1969) have proposed that the dunite-peridotite bodies of the Appalachian orogenic belts were emplaced into the geosynclinal sequence as solid serpentinite which was subsequently dehydrated during regional metamorphism. The origin of the serpentinite was not discussed but, according to this interpretation, the peridotites are metamorphic rocks.

This review of Table 4 has demonstrated the diversity of current opinions about the petrogenesis of ultramafic rocks in orogenic belts. A report by a large research group studying the ultrabasic rocks of Japan (Research Group, 1967) tends to support my contention that this diversity of views results from a diversity of origins, and not merely from a diversity of observers and interpretations. Major ultrabasic masses have been found in ten geologically distinct areas, and it is stated that there are considerable differences among the rocks in the first five zones so far studied. Certainly, petrologists must exercise careful judgement before equating any specific orogenic peridotite with the upper mantle.

#### PETROGENESIS OF ULTRAMAFIC ROCKS OF THE OCEANS

Hess (1960, 1962, 1964) proposed that layer 3 of the oceanic crust was composed of serpentinized mantle peridotite migrating laterally away from midoceanic ridges. During the past few years, ultramafic rocks have been dredged from many places in the Atlantic and Indian

oceans, especially near fracture zones, although very few have been found in the Pacific. Certainly serpentinites are widely distributed on the ocean floors, but it is not established that layer 3 of the oceanic crust is ultramafic; the debate was discussed by Vine and Hess (1970). St Peter and St. Paul Rocks near the equator on the mid-Atlantic ridge are composed of fresh peridotites, and these have long been postulated as a faulted exposure of the upper mantle (Tilley, 1947). Melson *et al.* (1967) reported the occurrence of three types of mylonitic peridotites containing spinel or hornblende, or both, and concluded that these were derived from the upper mantle at depths of 30 kms or more. The presence of hydrous and chlorine-rich phases in recrystallized assemblages suggested that these rocks contained an abundant interstitial fluid phase during intrusion.

The composition and petrogenesis of serpentinites from the mid-Atlantic Ridge were reviewed by Miyashiro *et al.* (1969), who discussed alternative structural interpretations for the occurrence of serpentinite. They concluded that the serpentinites were derived from heterogeneous upper mantle peridotite. Several recent interpretations of ultramafic rocks of the ophiolite suite (see Table 4) involve the supposition that these represent ocean floor or suboceanic mantle in eugeosynclines, which has been incorporated into the orogenic belt by thrusting or gravity sliding (Gass, 1968; Davies, 1968; Peters, 1969; Reinhardt, 1969). Thayer (1969b) concluded that the structure of the midoceanic ridges is probably very similar to that of the chromitite-bearing alpine peridotite-gabbro complexes of continental environments (including ophiolites of some other authors), which implies that gabbro interlayered with peridotite is an essential constituent of the upper mantle. Thus, there is increasing evidence that many ultramafic rocks of the oceanic regions are mantle-derived, but more data are required before extrapolation to the mantle can be made with confidence. Hydrothermal action is known to have caused marked changes in chemical composition and mineralogy in some locations (Cann and Vine, 1966).

According to the sea-floor spreading hypothesis, ultramafic rocks of orogenic belts, island arcs and ocean trenches, the midoceanic ridges, and the ocean floor between these features all have common petrogenetic links with the upper mantle.

#### PETROGENESIS OF ULTRAMAFIC NODULES

The various sources from which ultramafic nodules in basalts and kimberlites could be derived have already been reviewed, and the various types of mantle material some of them could represent have also been discussed. There is good evidence that in a single locality the nodules may be of several types with distinct origins (White, 1966; Aoki and Kushiro, 1968; Jackson, 1969).

Forbes and Kuno (1967) presented a world-wide review of peridotite inclusions and their basaltic host-rocks, and they noted that the spinel-lherzolite most frequently reported appeared to be related to the genesis of the basaltic mantle within the mantle. According to Hurley (1967),

$Sr^{87}/Sr^{86}$  ratios confirm that the inclusions are genetically related to the host basalts, and indicate that there is no contemporary relationship with alpine-type intrusions. Their low K/Rb ratios (Murthy and Stueber, 1967) and O'Hara's (1967c) pyroxene grid indicate that they equilibrated in the uppermost mantle, within the crystallization interval of anhydrous basaltic liquids. Goles (1967) and Murthy and Stueber (1967) drew attention to their residual character. More recent geochemical studies of nodules in basaltic rocks lead to diverse interpretations. Nagasawa *et al.* (1969) concluded from rare earth abundances in peridotite nodules that the nodules would have no genetic relationship to their host basalts, but would be fragments of mantle peridotite which supplied basaltic magma. In contrast, the contents of uranium, thorium and potassium in nodules, reported by Wakita *et al.* (1967), suggested that these nodules were not representative of the suboceanic mantle. Kleeman *et al.* (1969) determined the uranium distribution in peridotite inclusions from basalts using fission tracks, and they concluded that these inclusions could not have formed as cumulates, nor were they residual mantle material, but accidental fragments of inhomogeneous mantle. MacGregor (1968) considered mineral facies for mafic and ultramafic rocks (see Figs 9 and 11) as a guide for interpretation of the mineralogy of inclusions in terms of the minimum depth of origin of the host basaltic magmas. The petrogenesis of peridotite nodules has also been reviewed by O'Hara (1968), den Tex (1969), and Jackson (1969).

Interpretation of the peridotite nodules in kimberlites is subject to the same problems as those for basaltic hosts. Both eclogite and garnet peridotite nodules in kimberlites appear to be adequate sources of basaltic magmas. The origin of garnet peridotite nodules has been reviewed by Davidson (1967b) and Dawson (1968). O'Hara's (1967c) pyroxene grid indicates that these equilibrated in the mantle at depths of 100 to 140 kms on the geotherm beneath stable shelf areas. A wide variation in mineralogy is established for garnet peridotite nodules in kimberlites, and if these rocks are mantle fragments, Davidson's (1967b) comment demands attention: "If these inclusions represent the upper mantle, then the latter is clearly of so varied a composition that petrogenetic studies based on a single rock-type such as garnet-peridotite must be of questionable value."

Harris *et al.* (1967) proposed that nodules in basalts and kimberlites represent mantle fragments from a pyroxene-peridotite layer and deeper garnet-peridotite (see Fig. 9). The variable composition of pyroxene-peridotite nodules corresponds to various stages of depletion of basaltic components, the original mantle pyroxene-peridotite probably being chemically equivalent to the deeper garnet-peridotite. They pointed out that the few garnet-peridotite nodules so far analyzed may not be representative.

Peridotite nodules in kimberlites and basalts may be accompanied by eclogitic xenoliths, which range in composition from basalt to picrite basalt. The origin of eclogite nodules in kimberlite has been reviewed recently by Ku-

shiro and Aoki (1968) and by Dawson (1968). Mason (1968a) reviewed the petrogenesis of the eclogitic xenoliths in the breccia at Kakanui, New Zealand. In addition to peridotite and eclogite xenoliths, the breccia encloses large xenocrysts of pyrope, kaersutitic hornblende, and tschermakitic augite (Mason, 1966b; Dickey, 1968). Mason noted the rarity of eclogitic xenoliths in basaltic rocks, and he compared the Kakanui eclogites with the few analogous occurrences he could find in the literature. He concluded that eclogitic xenoliths in basaltic rocks probably crystallized at depths of about 60 kms, which raises the question of why eclogitic xenoliths of this kind are so rare when peridotite occurrences are so common. He suggested eclogitic material might exist below peridotite in the mantle which is not inconsistent with Press' (1969) conclusion that eclogite is a dominant constituent of the suboceanic mantle between 80 and 150 kms. Alternatively, eclogite within the mantle may be readily remelted and incorporated in its host during uprise to the surface. Kushiro and Aoki (1968) concluded the chemistry of eclogite nodules and their minerals in the Roberts-Victor kimberlite, South Africa, are consistent with their formation by crystallization of picritic magma at high pressures within the mantle, the eclogite forming pockets within the mantle peridotite.

Jackson (1969) emphasized the need, for interpretation of nodule suites, to determine both depth of origin and process of formation; processes may be elucidated by recognition of cumulus, tectonite, or metasomatic textures. He proposed that a statistical approach was required for each suite studied in order to determine the proportions of inclusions formed at different depths by each recognizable process. His concluding paragraph reviews the problem perfectly: "Trying to reconstruct the depth of origin and process of formation of xenoliths in any one individual tuff or flow is akin to trying to map the geology of an inaccessible highland area by looking at the boulders in the bed of an emergent stream. Multiple origins and differences in distances of transport of fragments, even those deposited side by side, might as well be assumed in both cases, and generalizations ought to be made with extreme care."

#### ESTIMATES OF MANTLE COMPOSITION BASED ON TERRESTRIAL ROCKS

Review of the petrogenesis of the ultramafic rocks in groups 3, 4, and 5 of Table 3 confirms that some of these rocks represent mantle material, but it also posts clear warnings that extrapolation to the upper mantle should be made with caution. Ringwood is largely responsible for directing the attention of petrologists to model compositions for the upper mantle based on petrological considerations (see Table 8), but attempts to correlate the composition of the upper mantle with specific mantle-derived ultramafic rocks appear to have begun with Hess' (1964) review of the chemistry of oceanic serpentinites. Since 1967, many accounts of the chemistry of ultramafic rocks in a variety of environments have included discussions of mantle chemistry.

TABLE 5. AVERAGE COMPOSITIONS OF ULTRAMAFIC NODULES

| Wt. %                          | 1      | 2     | 3     | 4      | 5      | 6      | 7     |
|--------------------------------|--------|-------|-------|--------|--------|--------|-------|
| SiO <sub>2</sub>               | 44.35  | 44.18 | 41.10 | 45.52  | 44.44  | 45.62  | 44.14 |
| MgO                            | 37.00  | 40.95 | 46.33 | 41.74  | 39.87  | 41.41  | 40.71 |
| FeO                            | 9.70   | 7.34  | 9.31  | 7.32   | 7.76   | 5.89   | 7.16  |
| Fe <sub>2</sub> O <sub>3</sub> | 0.28   | 1.16  | 1.24  | —      | —      | 1.38   | 1.93  |
| Al <sub>2</sub> O <sub>3</sub> | 4.58   | 2.81  | 0.56  | 2.42   | 3.06   | 2.55   | 2.56  |
| CaO                            | 3.01   | 2.49  | 0.17  | 2.04   | 3.42   | 1.90   | 2.24  |
| Na <sub>2</sub> O              | 0.34   | 0.22  | 0.23  | 0.18   | 0.52   | 0.19   | 0.28  |
| K <sub>2</sub> O               | 0.02   | 0.04  | 0.03  | 0.17   | 0.17   | 0.12   | 0.07  |
| Cr <sub>2</sub> O <sub>3</sub> | 0.27   | 0.3   | 0.35  | 0.33   | 0.33   | 0.34   | 0.32  |
| NiO                            | 0.21   | 0.27  | 0.44  | 0.23   | 0.46   | 0.25   | 0.26  |
| CoO                            | 0.017  | —     | —     | —      | —      | —      | —     |
| MnO                            | 0.14   | 0.14  | 0.15  | 0.13   | 0.08   | 0.13   | 0.13  |
| P <sub>2</sub> O <sub>5</sub>  | —      | —     | —     | —      | —      | 0.03   | 0.03  |
| TiO <sub>2</sub>               | 0.20   | 0.09  | 0.08  | 0.18   | 0.34   | 0.20   | 0.15  |
| Total                          | 100.12 | 99.99 | 99.99 | 100.26 | 100.45 | 100.01 | 99.98 |

1. Carter (1966). Analysis based on modal abundances of minerals for 150 nodules from five basaltic hosts.
2. Harris *et al.* (1967). Mean of 5 high calcium, high aluminum olivine nodules.
3. Harris *et al.* (1967). Average of 3 olivine nodules with CaO and Al<sub>2</sub>O<sub>3</sub> contents less than 1%; residual nodules.
4. Carswell (1968b). Mean of 5 determinations (except for 3 oxides) from South African kimberlite pipes.
5. Carswell (1968b). Mean of 4 determinations from Yakutia kimberlite pipes.
6. Ito and Kennedy (1967). Average of 4 lherzolite nodules in kimberlite.
7. Ito and Kennedy (1967). Average of 8 peridotite nodules in basalt.

*Peridotite nodules.* Table 5 lists seven compositions specified as upper mantle material on the basis of study of peridotite nodules in kimberlites and basalts. Carter (1966) examined 150 peridotite inclusions from five basaltic host-rocks and he plotted modal abundances and the chemistry of coexisting minerals versus the fayalite content of the olivine. He found continuity except for a small gap between 12-14 percent fayalite, and he argued that this gap represented the composition of the primitive upper mantle. From his plots, he estimated the mineralogy of the upper mantle to be as shown in Table 9, and he computed the bulk composition from the mode. The value listed under column 1 in Table 5 is the average value, not taking into account the estimated modal errors listed in Table 9.

Harris *et al.* (1967) derived compositions 2 and 3 in Table 5 from the average chemistry of 27 peridotite and three garnet-peridotite nodules in basalts and kimberlites. The average of five selected analyses of peridotite nodules with CaO and Al<sub>2</sub>O<sub>3</sub> contents between 2 and 3 percent is given in column 2, and this they considered to represent mantle material that has undergone only limited or no depletion in fusible components. Column 3 is the average of three peridotite nodules with CaO and Al<sub>2</sub>O<sub>3</sub> each less than 1 percent. This is considered by Harris *et al.* to represent samples of the infusible residue of the mantle after extraction of basaltic magma, or of cumulate olivine from magma rising through the mantle.

Columns 4 and 5 are the mean compositions of garnet-peridotite nodules from kimberlite pipes in South Africa and Yakutia, respectively, which were compared by Carswell (1968b) with garnet peridotites in the basement gneiss in Norway (see Table 7). Carswell concluded that these represented upper mantle material.

Columns 6 and 7 are averages of peridotite nodules from

TABLE 6. AVERAGE COMPOSITIONS OF OCEANIC ULTRAMAFIC ROCKS

| Wt.<br>%                       | Hess, 1964        |                   | Melson <i>et al.</i><br>1967 |
|--------------------------------|-------------------|-------------------|------------------------------|
|                                | 1                 | 2                 | 3                            |
| SiO <sub>2</sub>               | 39.82             | 43.56             | 43.55                        |
| MgO                            | 48.60             | 41.53             | 38.45                        |
| FeO                            | 7.86 <sup>a</sup> | 7.77 <sup>a</sup> | 6.64                         |
| Fe <sub>2</sub> O <sub>3</sub> | 1.00 <sup>a</sup> | 1.00 <sup>a</sup> | 1.45                         |
| Al <sub>2</sub> O <sub>3</sub> | 0.87              | 2.36              | 3.69                         |
| CaO                            | 0.37              | 2.51              | 2.61                         |
| Na <sub>2</sub> O              | 0.37              | 0.32              | 0.33                         |
| K <sub>2</sub> O               | <sup>b</sup>      | <sup>b</sup>      | 0.10                         |
| Cr <sub>2</sub> O <sub>3</sub> | 0.46              | 0.40              | 0.51                         |
| NiO                            | 0.46              | 0.34              | 0.27                         |
| MnO                            | 0.10              | 0.10              | 0.14                         |
| P <sub>2</sub> O <sub>5</sub>  | 0.08              | 0.07              | 0.13                         |
| H <sub>2</sub> O+              | —                 | —                 | 1.59                         |
| TiO <sub>2</sub>               | 0.01              | 0.04              | 0.28                         |
| H <sub>2</sub> O               | —                 | —                 | 0.12                         |
| Cl                             | —                 | —                 | 0.14                         |
| CO <sub>2</sub>                | —                 | —                 | 0.01                         |
| Total                          | 100.0             | 100.00            | 100.01                       |

1. Hess and Otolara (1964). Average (D+E)-type serpentinite, recalculated water-free.

2. Hess and Otolara (1964). Average C-type serpentinite, recalculated water-free.

3. Melson *et al.* (1967). Weighted average mylonite.

<sup>a</sup> Ferrous-ferric ratio adjusted so that Fe<sub>2</sub>O<sub>3</sub> is 1%.

<sup>b</sup> Means less than 0.005.

kimberlite and basalt, respectively, computed by Ito and Kennedy (1967) and presented as reasonable compositions for the upper mantle.

*Oceanic serpentinites and peridotites.* The mineralogy and chemistry of the serpentinite cores from the drill hole near Mayaguez, Puerto Rico, were described by Mattson (1964) and Hess and Otolara (1964). Rock types D and E are serpentinitized dunite, and type C is serpentinitized diopside harzburgite. Hess (1964) proposed that these rocks represent serpentinitized upper mantle. The average of types D plus E, presented in column 1 of Table 6, he suggested might be the residue after extraction of basalt from original material with the average composition of type C, presented in column 2. Hess noted that alkalis, thorium, uranium, and titanium are rather too low to satisfy the requirements of mantle peridotite, and he speculated about the role of phlogopite as a carrier of these elements. Miyashiro *et al.* (1969) reviewed the chemistry of 23 serpentinitized peridotites from midoceanic ridges (including 16 new analyses) and noted chemical similarities with presumed mantle peridotites occupying a range in composition; but they did not relate these compositions specifically to model mantle compositions.

Melson *et al.* (1967) analyzed three spinel peridotite mylonites and a brown hornblende mylonite from St. Peter and St. Paul Rocks, which they concluded had per-

haps crystallized at different levels within the mantle and, therefore, had bearing on the composition, mineralogy, and heterogeneity of the upper mantle. Individually, the rocks analyzed were not considered appropriate for the average mantle composition. Analysis 3 in Table 6 was presented as an estimated average composition of St. Paul's Rocks which is in accord with suggested average mantle compositions.

*Ultramafic rocks of orogenic belts, and average ultramafic rocks.* Green (1967) described a small group of peridotites with characteristics justifying their recognition as a distinctive type of orogenic ultramafic intrusion. He proposed in 1964 that one of these, the Lizard intrusion of Cornwall, represented a portion of mantle peridotite mobilized by partial melting (hypothesis IIIB in Table 4). The average composition of the Lizard peridotite (analysis 1 of Table 7) was presented as representing undepleted upper mantle.

The origin of the peridotites of Kalskaret, Norway, according to Carswell (1968a) has already been outlined in connection with hypothesis IIID of Table 4. Partial melting of upper mantle produced a picritic liquid which crystallized at depth as an eclogite, listed as analysis 3 in Table 7, in close association with the residual dunite, listed as analysis 2 in Table 7. Carswell (1968b) also studied the garnet peridotites enclosed in gneiss at Ugelvik, Norway, and noted that these did not show the varied chemistry of those at Kalskaret. He suggested that the Ugelvik garnet peridotites represent primary mantle material, and their mean composition is given by analysis 4 in Table 7. Carswell (1968b) also reviewed the other occurrences of garnet peridotites in the basement rocks of Czechoslovakia and Switzerland and in kimberlites (see O'Hara and Mercy, 1963). Mean compositions for the orogenic garnet peridotites are listed as analyses 5 and 6 in Table 7, and the nodule means are analyses 4 and 5 in Table 5. Because all of these garnet peridotites are similar, chemically and mineralogically, occupying restricted chemical ranges, he concluded that they possibly represented primary upper mantle material; the mean of 20 garnet peridotites is analysis 7 in Table 7. He noted that the alkali contents of the garnet peridotites were rather low if these are the source of basaltic magmas.

Analysis 8 in Table 7 is White's (1967) estimate of undepleted upper mantle peridotite, based on frequency histograms for the chemistry of 84 peridotites, 53 serpentinites, and 31 peridotite nodules. White argued that if representatives of each fractionation stage of differentiation processes operating on the original mantle are randomly sampled, then the more differentiated a stage, the lower its abundance. Therefore, the mode, or most abundant class, of exposed ultrabasic rocks should represent original mantle peridotite. Partly fused material from the mantle should have a bimodal distribution with peaks for basaltic liquid and for residual dunite. White plotted separately the analyses for rocks with between 1 and 5 percent CaO plus Al<sub>2</sub>O<sub>3</sub> with a view to eliminating the dunitic and more

TABLE 7. AVERAGE PERIDOTITES OF OROGENIC BELTS, AND AVERAGE ULTRAMAFIC ROCKS

| Wt.<br>%                       | Green<br>1964 | Carswell 1968a and 1968b |       |        |        |        |        | White<br>1967 | Harris <i>et al.</i><br>1967 |
|--------------------------------|---------------|--------------------------|-------|--------|--------|--------|--------|---------------|------------------------------|
|                                | 1             | 2                        | 3     | 4      | 5      | 6      | 7      | 8             | 9                            |
| SiO <sub>2</sub>               | 44.77         | 40.66                    | 46.53 | 44.65  | 44.07  | 44.72  | 44.66  | 44.5          | 44.2                         |
| MgO                            | 39.22         | 46.48                    | 16.08 | 41.66  | 41.07  | 39.88  | 40.97  | 41.7          | 41.3                         |
| FeO                            | 8.21          | 3.00                     | 8.65  | 6.81   | 8.97   | 8.30   | 7.92   | 7.3           | 7.3                          |
| Fe <sub>2</sub> O <sub>3</sub> | —             | 3.40                     | 3.30  | —      | —      | —      | —      | 1.5           | 1.1                          |
| Al <sub>2</sub> O <sub>3</sub> | 4.16          | 0.48                     | 11.58 | 3.50   | 2.50   | 3.09   | 2.80   | 2.55          | 2.7                          |
| CaO                            | 2.42          | 0.09                     | 11.00 | 2.02   | 2.60   | 3.16   | 2.59   | 2.25          | 2.4                          |
| Na <sub>2</sub> O              | 0.22          | 0.04                     | 1.01  | 0.23   | 0.43   | 0.24   | 0.30   | 0.25          | 0.25                         |
| K <sub>2</sub> O               | 0.05          | 0.03                     | 0.02  | 0.04   | 0.04   | 0.00   | 0.12   | 0.015         | 0.015                        |
| Cr <sub>2</sub> O <sub>3</sub> | 0.40          | 0.28                     | 0.12  | 0.59   | 0.54   | 0.40   | 0.45   | —             | 0.30                         |
| NiO                            | 0.24          | 0.36                     | 0.06  | 0.29   | —      | 0.28   | 0.31   | —             | 0.20                         |
| MnO                            | 0.11          | 0.09                     | 0.22  | 0.14   | 0.12   | 0.13   | 0.12   | 0.14          | 0.15                         |
| H <sub>2</sub> O+              | —             | 5.51                     | 0.36  | —      | —      | —      | —      | —             | —                            |
| TiO <sub>2</sub>               | 0.19          | 0.01                     | 0.64  | 0.08   | 0.17   | 0.14   | 0.19   | 0.15          | 0.1                          |
| Total                          | 99.99         | 100.43                   | 99.57 | 100.01 | 100.51 | 100.34 | 100.43 | 100.36        | 100.02                       |

1. Green (1964). Average composition for the Lizard peridotite.
2. Carswell (1968a). Serpentinized dunite, T268, from Kalskaret, Norway.
3. Carswell (1968a). Eclogite, T153, from Kalskaret, Norway.
4. Carswell (1968b). Mean of 3 garnet peridotites from Ugelvik, Norway.
5. Carswell (1968b). Mean of 6 garnet peridotites from the Czech Massif.
6. Carswell (1968b). Mean of 2 garnet peridotites from Bellinzona, Switzerland.
7. Carswell (1968b). Mean of 20 garnet peridotites—4, 5, and 6 above, and 4 and 5 from Table 5.
8. White (1967). Upper mantle composition estimated from frequency histograms of 168 ultramafic rocks. Total iron is divided arbitrarily between FeO and Fe<sub>2</sub>O<sub>3</sub>. Na<sub>2</sub>O and K<sub>2</sub>O are from Stueber and Murthy (1966) and Stueber and Goles (1967).
9. Harris *et al.* (1967). Estimated upper mantle based on analysis 8, and on analysis 2 for nodules in Table 5.

basaltic analyses, so that the mode for unfractionated mantle peridotite would not be masked.

Harris *et al.* (1967) compared the dominant ultramafic rock derived by White (1967) with the average compositions of peridotite nodules (analyses 2 and 3 in Table 5), and presented a preferred composition for original or undepleted mantle based on these analyses and other considerations. This is analysis 9 in Table 7.

Element abundances in suites of ultramafic rocks have been the subject of several recent studies, using activation analysis and spectrochemical analysis (Stueber and Goles, 1967; Stueber and Murthy, 1966; Stueber *et al.*, 1968; Fisher *et al.*, 1969). Fisher *et al.* analyzed 58 ultramafic rocks from diverse geographic and geological settings, using most of the rocks studied by Stueber and his colleagues. They compared their results with several of the estimates of upper mantle composition reviewed above and concluded that, although their average values for peridotites agreed well with all of these estimates, the wide spread observed in their determinations cast doubt on the validity of averages based on only a few samples. Average values for peridotites, with their ranges, include: SiO<sub>2</sub>=41±4; TiO<sub>2</sub>=0-2; Al=1.6±1.6; Fe=7.2±4.0; Mn=0.117±0.04; Cr=0.3±0.3; Ni=0.16±0.16. Among the various rock types included in the suite, they could find no chemically coherent type that might represent the upper mantle. They concluded, therefore, that either the mantle is heterogeneous, or none of the rock types included in the study is representative of primitive mantle material.

*Hypothetical peridotites.* Another approach involves the formulation of a mantle composition based on petrological theory rather than on specific rocks. The general concept of a peridotitic mantle as a source for basaltic magmas by partial melting dates back at least as far as Bowen's treatise in 1928. Ringwood (1966a) outlined the development of ideas along these lines, leading to his formalization of the concept in 1962. He postulated the existence of a primitive, parental mantle material, which was "defined by the property that on fractional melting it would yield a typical basaltic magma and leave behind a residual refractory dunite-peridotite of alpine type." (Ringwood, 1966a, p. 303). Ringwood (1962) postulated a primitive mantle composition of four parts dunite and one part basalt and gave it the name "pyrolite." He defended the adoption of a new name in 1966a (p. 304) because "The composition and properties of pyrolite do *not* match those of natural peridotites, particularly in trace element chemistry." I agree with Ringwood (1966a) that there are valid objections to inventing new names for hypothetical materials; I disagree that a new name is required for this hypothetical material. Rocks are named on the basis of their mineralogy, not on their trace element content, and the "pyrolite" introduced above satisfies any definition of a peridotite. Most petrologists concerned with upper mantle constitution use the existing names for varieties of peridotite.

Four estimates of mantle composition based on this hypothetical peridotite are listed as analyses 3, 4, 5, and 6 in Table 8. In 1963, Green and Ringwood presented analyses

TABLE 8. HYPOTHETICAL MANTLE PERIDOTITES

| Wt.<br>%                       | Ringwood and Green, pyrolites |        |        |        |        |        | Nicholls, 1967 |        |        |        |
|--------------------------------|-------------------------------|--------|--------|--------|--------|--------|----------------|--------|--------|--------|
|                                | 4:1                           |        | I      | III    | II     |        |                |        |        |        |
|                                | 1966                          | 1963   | 1963   | 1963   | 1966   | 1967   |                |        |        |        |
|                                | 1                             | 2      | 3      | 4      | 5      | 6      | 7              | 8      | 9      | 10     |
| SiO <sub>2</sub>               | 48.72                         | 41.32  | 42.71  | 43.06  | 45.16  | 43.95  | 44.76          | 45.2   | 45.1   | 40.3   |
| MgO                            | 8.98                          | 49.81  | 41.41  | 39.32  | 37.47  | 39.00  | 10.23          | 44.2   | 36.7   | 32.7   |
| FeO                            | 9.88                          | 5.91   | 6.51   | 6.66   | 8.04   | 7.50   | 10.43          | 7.2    | 7.9    | 7.1    |
| Fe <sub>2</sub> O <sub>3</sub> | 1.70                          | 1.21   | 1.57   | 1.66   | 0.46   | 0.75   | 3.73           | 1.5    | 2.0    | 1.8    |
| Al <sub>2</sub> O <sub>3</sub> | 13.40                         | 0.54   | 3.30   | 3.99   | 3.54   | 3.88   | 15.88          | 0.8    | 4.1    | 3.7    |
| CaO                            | 11.30                         | Trace  | 2.11   | 2.65   | 3.08   | 2.60   | 10.47          | —      | 2.3    | 2.1    |
| Na <sub>2</sub> O              | 2.23                          | Trace  | 0.49   | 0.61   | 0.57   | 0.60   | 2.22           | 0.1(5) | 0.6    | 0.5    |
| K <sub>2</sub> O               | 0.58                          | Trace  | 0.18   | 0.22   | 0.13   | 0.22   | 0.08           | —      | 0.0(2) | 0.0(2) |
| Cr <sub>2</sub> O <sub>3</sub> | —                             | 0.56   | 0.45   | 0.42   | 0.43   | 0.41   | —              | 0.4    | 0.3    | 0.3    |
| NiO                            | —                             | 0.52   | 0.42   | 0.39   | 0.20   | 0.39   | —              | 0.3    | 0.2    | 0.2    |
| CoO                            | —                             | 0.02   | 0.02   | 0.02   | 0.01   | —      | —              | —      | —      | —      |
| MnO                            | 0.18                          | 0.11   | 0.13   | 0.13   | 0.14   | 0.13   | 0.12           | 0.2    | 0.2    | 0.1    |
| P <sub>2</sub> O <sub>5</sub>  | 0.24                          | Trace  | 0.06   | 0.08   | 0.06   | —      | 0.09           | 0.0(5) | 0.1    | 0.1    |
| H <sub>2</sub> O+              | —                             | Nil    | 0.17   | 0.21   | —      | —      | —              | —      | —      | 9.7    |
| TiO <sub>2</sub>               | 2.77                          | Trace  | 0.47   | 0.58   | 0.71   | 0.57   | 1.99           | 0.0(2) | 0.5    | 0.4    |
| CO <sub>2</sub>                | —                             | —      | —      | —      | —      | —      | —              | —      | —      | 0.8    |
| Cl                             | —                             | —      | —      | —      | —      | —      | —              | —      | —      | 0.2    |
| Total                          | 99.98                         | 100.00 | 100.00 | 100.00 | 100.00 | 100.00 | 100.00         | 100.00 | 100.00 | 100.00 |

1. Ringwood (1966). Primitive Hawaiian olivine tholeiite.
2. Green and Ringwood (1963). Average of 4 anhydrous dunites containing less than 22% H<sub>2</sub>O.
3. Green and Ringwood (1963). Pyrolite with 4:1 of, respectively, dunite 2 and the mean of Nockold's (1954) average normal tholeiite and normal alkali basalt.
4. Green and Ringwood (1963). Pyrolite with 3:1 of dunite to mean basalt as in 3. Designated pyrolite I by Green and Ringwood (1967a).
5. Ringwood (1966). Pyrolite with 3:1 of, respectively, a "synthetic" alpine-type peridotite and the Hawaiian basalt 1. The "synthetic" peridotite is assumed to consist of 80% olivine, 19.3% orthopyroxene, and 0.7% chrome spinel. Individual compositions for these minerals are taken from values determined by Ross, Foster and Myers (1954) for minerals separated from alpine-type peridotites. Designated pyrolite III by Green and Ringwood (1967a).
6. Green and Ringwood (1967b). Synthetic peridotite used as basis for experimental phase studies; designated pyrolite II.
7. Nicholls (1967). Calculated preferred basaltic fraction from the mantle, based on high-aluminum tholeiitic glass and a spinellid mineral (Nicholls analysis 4, table 5).
8. Nicholls (1967). Estimated residual fraction of mantle, based largely on oceanic serpentinites (Nicholls analysis 2, table 9).
9. Nicholls (1967). Estimated mantle material from which volatile components have been removed, but not the basaltic components (Nicholls, analysis 1, table 9).
10. Nicholls (1967). Composition of volatile-rich parts of the upper mantle, such as may occur beneath the midoceanic ridges (Nicholls, analysis 3, table 9).

3 and 4 based on ratios of 4:1 and 3:1, respectively, of dunite 2 and of Nockold's (1954) average normal tholeiite and normal alkali basalt. In 1966a, Ringwood modified the 3:1 model as in analysis 5, by combining a "synthetic" alpine-type peridotite with a primitive Hawaiian olivine tholeiite (analysis 1, Table 8). For experimental studies of peridotite mineral facies, Green and Ringwood (1967b) presented another composition listed in column 6 of Table 8. The three 3:1 models were then designated "pyrolites" I, II, and III as shown in Table 8. Ringwood has emphasized flexibility. Although he considers the 3:1 ratio as the most likely, he has stated that the ratio of peridotite: basalt may vary within the limits 1:1 to 4:1; the model compositions can then be revised as additional restraints are uncovered. By Ringwood's definition, "pyrolite" forms the upper mantle, whatever the final solution to the chemical problem may be.

Nicholls (1967) approached the problem of upper mantle composition by assuming that it was composed of a volatile fraction, consisting of low molecular weight compounds that have been liberated from volcanoes to the atmosphere and hydrosphere, a basaltic lava fraction derived by partial melting of the mantle, and a residual fraction remaining after depletion in volatiles and basalt. Nicholls examined the possible compositions of each of these fractions to define their limits, and then evaluated the proportions in which the three fractions might be combined in undepleted mantle material, thus placing limits on the composition of the upper mantle beneath the oceans. He concluded the preferred composition for the volatile fraction of the mantle is 91 percent H<sub>2</sub>O, 7 percent CO<sub>2</sub>, 2 percent Cl, and traces of other components. Analysis 7 in Table 8 gives his estimated composition for the basaltic fraction of the mantle, which is based on 93 percent high-aluminum

tholeiitic glass and 7 percent spinellid mineral of composition 50 percent  $\text{Fe}_2\text{TiO}_4$  and 50 percent  $\text{MgFe}_2\text{O}_4$ . The spinellid mineral was introduced into the calculations to account for the concentrations of  $\text{TiO}_2$  in various basaltic rocks. For the residual fraction in the mantle, Nicholls presented analysis 8 in Table 8, an estimate based on deep sea serpentinites (compare Table 6). Analysis 10 in Table 8 is an estimate of undepleted upper mantle material, still rich in volatile components, which Nicholls considers may exist beneath the midoceanic ridges; notice the high water content. Analysis 9 is his estimate of mantle material from which the volatile components have been removed, but which still retains the basaltic fraction. This is more akin to the other hypothetical peridotites in Table 8.

*Trace elements.* It is difficult to determine the abundances of trace elements in ultramafic rocks, and even more difficult to extrapolate to the upper mantle from the rocks. There is a sampling problem that may be severe. For example, the Th and U contents of the Twin Sisters dunite (Goles, 1967) show marked variation from one specimen to another, and the concentrations of Rb, Sr, and K in ultramafic rocks are very sensitive to the distribution of small proportions of amphibole (Hart and Aldrich, 1967).

There is much evidence to suggest that the upper mantle is heterogeneous with respect to trace elements (Lovering and Tatsumoto, 1968). Berg (1968) has demonstrated that eclogite nodules from the Roberts-Victor kimberlite pipe have been contaminated by K, Rb, and Cs, through the formation of secondary phlogopite and a later zeolite. He concluded that analytical data obtained on whole rocks for K and geochemically related elements, and probably also Sr and the rare earth group, should be treated with some circumspection. Condie *et al.* (1969) studied the U distribution in clinopyroxenes separated from eclogite nodules from various sources, and found that most of the U occurs as contaminants along cracks or in micrinclusions. They concluded that the abundances of K and U, and the K/U ratio, of a substantial fraction of the earth's inventory remains unknown. Kleeman *et al.* (1969), on the other hand, found that in peridotite nodules from Victorian basalts, the U was concentrated homogeneously in clinopyroxenes, and it did not occur along grain boundaries. The presence and distribution of phlogopite or amphibole in the mantle, or in ultramafic rocks, exerts a marked control on the abundance and distribution of many elements.

The concentrations of U, Th, K, Rb, and Sr in the mantle have been estimated by geochemical methods without reference to specific mantle-derived ultramafic rocks (Hurley, 1968a, 1968b; Shaw, 1968). Fractionation of trace elements in the mantle has been discussed by Harris (1967) and Gast (1968). There have been many proposals that the uppermost mantle has trace element concentrations different from the whole mantle, arising from differentiation of the mantle at an early stage of earth history, and from the continued depletion of the upper mantle in basaltic components.

*Volatile components.* The composition of the volatile fraction of the mantle, and its relationship to the atmosphere and hydrosphere, was reviewed by Nicholls (1967). He concluded that in certain environments, such as beneath the midoceanic ridges, the mantle peridotite might contain as much as 9.7 percent  $\text{H}_2\text{O}$  and 0.8 percent  $\text{CO}_2$  (analysis 10 in Table 8). Melson *et al.* (1967) concluded that the ultramafic rocks of St. Peter and St. Paul rocks, derived from the upper mantle, contained an abundant interstitial fluid phase during intrusion. Elsewhere, however, it is generally held that the  $\text{H}_2\text{O}$  content of the mantle is very low, probably no more than 0.1 percent. The presence of  $\text{CO}_2$  is inferred from the existence of  $\text{CO}_2$ -filled inclusions in olivines derived from upper mantle sources (Roedder, 1965).

Petrographic study of nodules and ultramafic rocks believed to be derived from the mantle provides evidence that phlogopite and hornblende are primary constituents of the mantle, at least locally. The existence of phlogopite in the mantle, and its role as a host for potassium, has received much speculative discussion; it is also a source of  $\text{H}_2\text{O}$ . Dawson and Powell (1969) recently described unambiguous evidence that phlogopite is a primary phase in peridotite nodules from Lashaine volcano, in Tanzania. The presence of amphibole in the mantle is indicated by the hornblende peridotites of St. Paul's Rocks (Melson *et al.*, 1967), and by the hornblende-eclogite nodules reviewed by Mason (1968a, 1968b). Wilshire and Trask (1969) concluded that hornblende and phlogopite in peridotite nodules from the Dish Hill craters in California are mantle constituents of the host peridotites, which were accidentally incorporated in the basanite agglomerate.

The proportion of volatile components in the mantle is very small, and probably variable both laterally and vertically. However, even a small proportion of volatiles may have a very pronounced effect, as shown by Figure 12. The most complete discussion available on the behavior of water under mantle conditions is that by Burnham (1967). The regime of volatiles in the earth is of fundamental significance for many problems.

#### CHEMICAL VARIATION DIAGRAMS

The estimated mantle compositions listed in Tables 1, 5, 6, 7, and 8 are compared in chemical variation diagrams in Figures 4 through 8. Distinction is made in each diagram among extraterrestrial, terrestrial, and hypothetical estimates. Terrestrial estimates are identified as average rocks from oceanic or continental environments, or as nodules. The points distinguished as continental include some averages of all types of ultramafic rocks, but these are strongly weighted by continental analyses. The terrestrial estimates for residual mantle material are labelled separately from the terrestrial estimates for undepleted mantle. Average basalts plotted in some figures include those listed in Table 8, and some from Manson's (1967) compilation.

Figure 4 shows the variation of total alkalis with respect to CaO, and Figure 6 introduces also the variation in  $\text{Al}_2\text{O}_3$ . These are the components of particular interest with respect to the generation of magmas from mantle material.

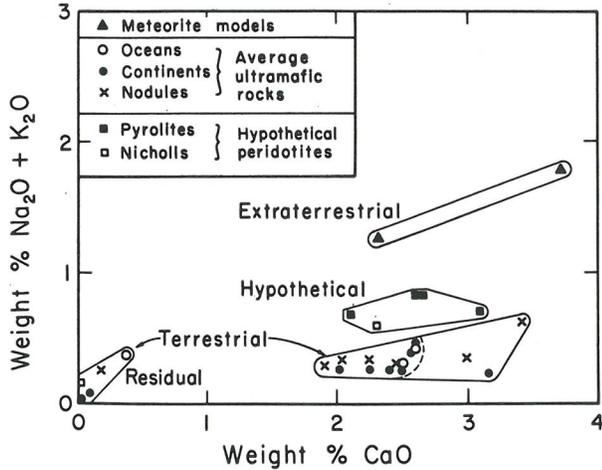


FIG. 4. Chemical variation of estimated mantle compositions. See Fig. 5.

Figure 5 is an expanded version of Figure 4 showing the relationship of basalts to the mantle estimates. Figure 7 shows the mantle estimates plotted against two standard differentiation indices, and Figure 8 is an expanded version of this showing the relationship of basalts to the mantle peridotites.

In all three chemical variation diagrams, the extraterrestrial, terrestrial, and hypothetical estimates occupy distinct areas, with no overlap. The terrestrial estimates of undepleted mantle material are also distinct from estimates of residual mantle, but of course there does exist a whole range of rock compositions between the two, representing various stages of depletion in basaltic components. Figure 4 shows that according to the meteorite models, the mantle as a whole contains significantly more alkalis than the upper mantle according to estimates based on terrestrial rocks. Also, the observed ultramafic rocks contain distinctly less alkalis than the hypothetical models. The estimates of undepleted mantle based on ultramafic rocks from a variety of environments are grouped together closely; the

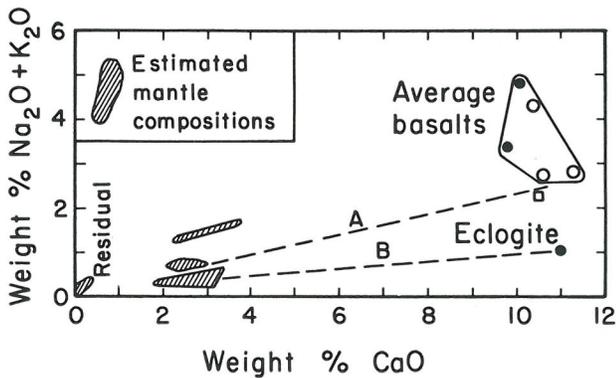


FIG. 5. Expanded version of Fig. 4, comparing mantle estimates with average basalts and a picritic eclogite.

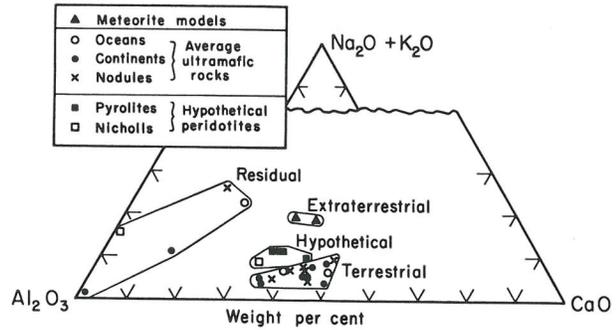


FIG. 6. Chemical variation of estimated mantle compositions. Compare Fig. 4.

nodule estimates of Carter (1966) and garnet peridotites from Yakutia and Switzerland (Carswell, 1968b) depart from the main group. Figure 6 shows that  $CaO/Al_2O_3$  ratios for the mantle estimates occupy a limited range. Relative to lime plus alkalis,  $Al_2O_3$  is stored up in the residual types.

Figures 5 and 8 compare the mantle estimates with average basaltic compositions. The point labelled eclogite is analysis 3 in Table 7, interpreted by Carswell (1968a) as a product of partial melting in the mantle, a picritic liquid, that crystallized at depth. In Figure 5, the line A is drawn through the origin, which corresponds closely to residual dunite, in a position separating the terrestrial mantle estimates from the hypothetical peridotites. The position of the basaltic averages above this line confirms that they can be derived by subtraction from the hypothetical peridotites, leaving residual peridotite, as is required by the models adopted for the mantle and hypothetical mantle materials. The position of the terrestrial estimates below line A suggests that the average basalts cannot be derived by direct subtraction from known ultramafic rocks if the residual product is to be equivalent to those peridotites normally regarded as residual types.

The line B connects Carswell's (1968a) residual dunite near the origin (analysis 2, Table 7) with the coexisting eclogite, and this line passes right through the independent mantle estimates based on ultramafic rocks. This indicates that the eclogite of picritic composition could represent a liquid fraction developed at depth by partial melting of the observed ultramafic rocks, leaving the appropriate residual peridotite or dunite.

The hypothetical models assume that the average basalts are primary magmas derived from the mantle by partial melting, uprise, and eruption. If this is so, then it appears unlikely that the terrestrial ultramafic rocks provide reasonable estimates for the upper mantle; their alkali contents are too low. However, in two stimulating reviews on basalt genesis, O'Hara (1965, 1968) pointed out that the compositions of the basalts normally regarded as primary magmas could not have been in equilibrium with mantle material at the depth where partial melting occurred. The concept of primary magmas, which has held a prominent

position in petrogenesis for many years, is only one limiting case for magma generation. The other limiting case involves magma generation at depth, followed by continuous fractionation until the magma is erupted at the surface. It is now clear from a number of experimental studies that the liquid produced by partial melting of mantle peridotite is picritic rather than basaltic in composition (Ito and Kennedy, 1967, 1968; Green and Ringwood, 1967a; O'Hara, 1965, 1968). This magma changes composition during uprise through the mantle by precipitation of one or more of the minerals stable in the mantle peridotite through which it passes; the fractionation process may terminate with magma separation. The basaltic magma erupted at the surface is thus the end product of a complex fusion and crystallization process, and not a simple subtraction from the mantle material. Figure 8 confirms that the average basalts are more fractionated than the picritic eclogite, and fractionation of the picritic magma during uprise could therefore yield the common basaltic magma types, given appropriate conditions.

The chemical variation diagrams in Figures 5 and 8 are consistent with O'Hara's (1956, 1968) model and, if O'Hara is correct, the estimated mantle compositions based on ultramafic rocks may provide reasonable indications of the range of peridotite compositions occurring within the upper mantle. Then, the basic assumptions of the hypothetical models (Ringwood, 1962; Nicholls, 1967) would be incorrect, because high-pressure picritic magma and not

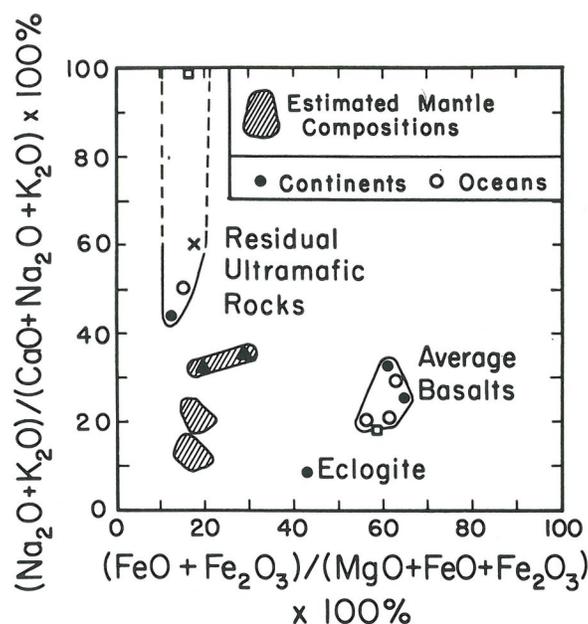


FIG. 8. Expanded version of Fig. 7, comparing mantle estimates with average basalts and a picritic eclogite. Low alkali contents in the residual ultramafic rocks make definition of the range of residual rock compositions difficult to establish.

low-pressure basaltic magma is the material to be combined with residual peridotite to reconstitute the mantle.

MINERAL FACIES FOR UPPER MANTLE MATERIALS

Evidence from extraterrestrial sources, from petrology, and from geophysics indicates that the upper mantle is composed of material with compositions corresponding to dunite, peridotite, and gabbro, and possibly with mixtures of these. Experimental studies on these compositions at appropriate pressures and temperatures delineate the mineral facies occurring within the upper mantle. There is reason to believe that water exists in the mantle, and the effect of water on the mineral facies must therefore be taken into account.

*Experimental results for peridotite-water.* There has been much speculation about the phase diagram for peridotite. Among those who presented schematic diagrams for peridotite mineral facies are Ringwood and Green, in a series of models reviewed by O'Hara (1968, p. 105-118), O'Hara (1967a, 1967c), and MacGregor (1967, 1968). Figure 9 shows the general pattern of the phase changes, with a peridotite consisting of olivine, clinopyroxene, and orthopyroxene, together with an aluminous phase; this is plagioclase feldspar in the low-pressure plagioclase-peridotite facies. With increasing pressure the plagioclase reacts with olivine to yield aluminous pyroxenes and spinel, producing a spinel-peridotite. At higher pressures, garnet-peridotite is produced by reaction of the spinel with orthopyroxene yielding pyrope-rich garnet and olivine. Several of the

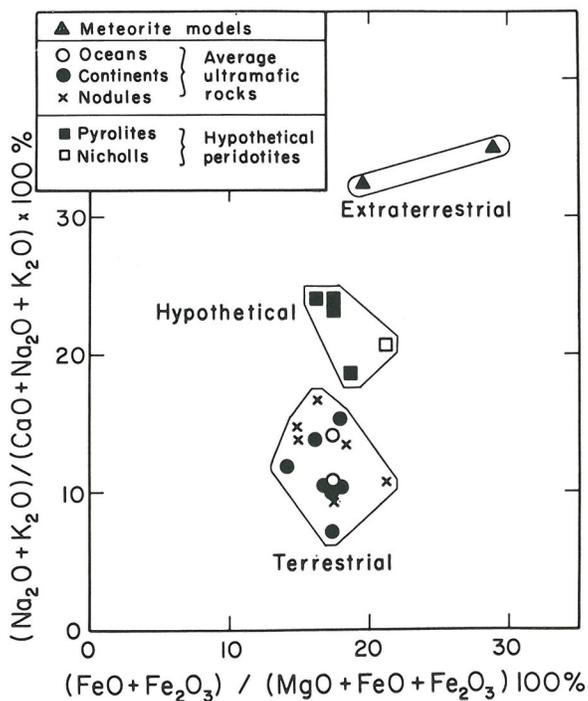


FIG. 7. Chemical variation of estimated mantle compositions in terms of indices of fractionation. Compare Fig. 8.

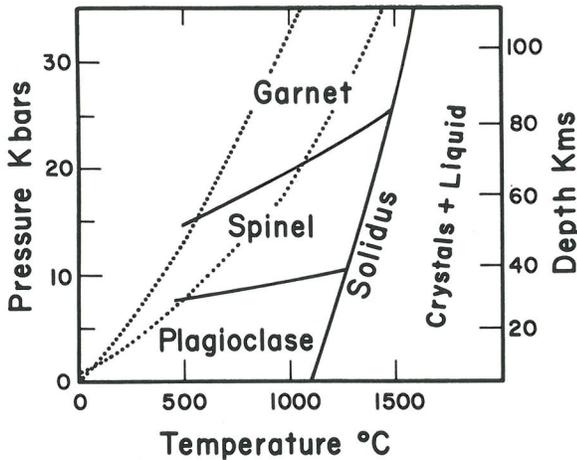


FIG. 9. Generalized phase diagram for an estimated mantle peridotite; the lines indicate the solidus of the peridotite, and the positions of divariant reaction intervals separating facies for spinel-peridotite from plagioclase-peridotite and from garnet-peridotite. Dotted lines represent geotherms for oceanic and shield regions.

simple reactions related to peridotite in the system  $\text{CaO-MgO-Al}_2\text{O}_3\text{-SiO}_2$  have been studied experimentally and these were reviewed by O'Hara (1968), MacGregor (1968) and by Ito and Kennedy (1968, Figs. 6 and 7). The few experimental studies completed on peridotites or equivalent synthetic mixtures confirm the general pattern of phase relationships illustrated in Figure 9, although the positions of the boundaries may change considerably with variation in the concentrations of  $\text{CaO}$ , and minor components such as  $\text{Al}_2\text{O}_3$ ,  $\text{Fe}_2\text{O}_3$ , and  $\text{Cr}_2\text{O}_3$ . MacGregor (1970) concluded that variations in the trivalent oxide ratio may move some reaction boundaries through 10 kbars pressure.

Green and Ringwood (1967b) presented stability fields for three synthetic peridotites with compositions equivalent to "pyrolites" I, II, and III, listed in Table 8. Experiments were conducted at  $1100^\circ\text{C}$  and above. The peridotite compositions I and II conformed to Figure 9, but peridotite III, with a lower  $\text{MgO/SiO}_2$  ratio, included an additional phase field just below the solidus; this is pyroxene-peridotite without spinel. With increasing temperature within the spinel-peridotite field, the spinel dissolves in the aluminous pyroxenes, and solution is complete at about  $1250^\circ\text{C}$ . The generation of garnet from the spinel-free peridotite III occurred at a higher pressure; garnet did not appear at the solidus until 30 kbars, instead of 25 kbars as depicted in Figure 9 for peridotites I and II. The boundaries between the subsolidus facies are plotted as univariant lines, but in fact they are reaction intervals.

Ito and Kennedy (1967) studied the melting relationships of a garnet peridotite nodule from a kimberlite pipe and noted just below the solidus the existence of a reaction interval between the facies for spinel-peridotite and garnet-peridotite, with a width of about 5 kbars. The mineral facies for a picrite just below the solidus were presented by Ito

and Kennedy in 1968. Kushiro *et al.* (1968a) located the solidus for a natural spinel-peridotite nodule from basaltic tuff and recorded a subsolidus change from spinel- to garnet-peridotite, but they did not attempt to plot subsolidus phase boundaries.

Addition of water to peridotite causes depression of the melting temperature and introduces three hydrous minerals, serpentine, amphibole, and phlogopite. Kushiro *et al.* (1968a) presented the solidus shown as a dashed line in Figure 10 for the beginning of melting of a spinel-peridotite nodule in the presence of excess water. This is the only experimental data for a rock, but the results of Kushiro *et al.* (1968b) on the join enstatite-water, and of Kushiro (1969a) on the system forsterite-diopside-silica-water are relevant to melting in the peridotite-water system (see Kushiro, 1969b).

The curve for serpentine gives the maximum temperature for the stability of serpentine in the presence of water. This is based on the data of Kitahara *et al.* (1966) at pressures above 6 kbars, and on data of Scarfe and Wyllie (1967; manuscript in preparation) at pressures of 3 kbars and below.

The curve for hornblende is based on results in the system gabbro-water by Yoder and Tilley (1962) at pressures below 10 kbars, and by Lambert and Wyllie (1968, 1970, manuscript in preparation) between 10 and 25 kbars. Lambert and Wyllie discovered that at pressures greater than about 17 kbars, hornblende breaks down in a pressure-sensitive reaction to yield pyroxene and garnet. This assemblage is more dense than the hornblende, so that  $dP/dT$  for the reaction becomes negative, in contrast with the normal pattern for dehydration reactions. There have been several predictions that amphibole break-down slopes

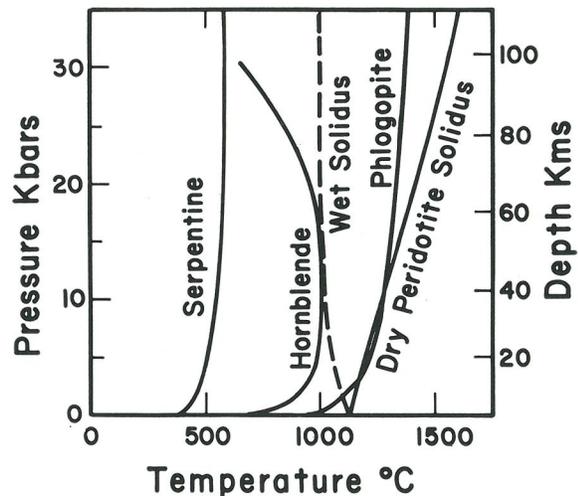


FIG. 10. Experimentally determined reactions involving peridotite and water, based on various experimental studies cited in the text. Curves are plotted for the solidus of peridotite dry, and in the presence of excess water (dashed); other curves give the upper temperature stability limits for serpentine, hornblende, and phlogopite.

should become negative at high pressures, but the measured change is more marked than anticipated. Gilbert (1969) presented a reconnaissance study of amphibole stability at high pressures in the presence of excess water, confirming the results of Lambert and Wyllie in the system gabbro-water. He determined that the position of the breakdown curve varied with composition, and amphiboles from ultrabasic rocks appear to be stable to somewhat higher pressures than those from basic or intermediate rocks (personal communication, 1969).

The curve drawn for phlogopite has not been accurately determined experimentally. High pressure studies on the stability of phlogopite have been reported by Markov *et al.* (1966), Kushiro *et al.* (1967), and Yoder and Kushiro (1969). Yoder and Kushiro located the melting curve for phlogopite plus forsterite in the presence of excess water with a very slight increase in temperature to 1200°C at 30 kbars. The maximum temperature stability curve for phlogopite itself, however, rises to about 1300°C at 10 kbars, with the phlogopite melting incongruently to a vapor-deficient assemblage forsterite plus liquid.

*Experimental results for gabbro-water.* Subsolidus mineral facies involve the gabbro-eclogite phase transition, and available experimental data indicate that this is strongly dependent on composition. The transition involves a series of reactions, producing four mineral facies: an olivine gabbro at low pressures in the pyroxene-hornfels facies passes with increasing pressure through the pyroxene-granulite facies by the reaction of olivine with plagioclase, then into the garnet-granulite facies by the formation of garnet, and then into the eclogite facies by the elimination of plagioclase. Mineral facies in gabbros have been reviewed by Ringwood and Green (1966), MacGregor (1968), and Ito and Kennedy (1968).

The work of Yoder and Tilley (1962) was followed by a more detailed study by Green and Ringwood (1967a), and the results were reviewed by Ringwood and Green (1966), along with results for additional basaltic compositions determined by their colleague T. H. Green. Most experiments were completed at a temperature of 1100°C, with a few runs down to 1000°C. There is still no experimental data for lower temperatures. Cohen *et al.* (1967) located the transition just below the solidus for a tholeiitic basalt, and similar transitions were determined by Ito and Kennedy (1968) for a picrite and an olivine-rich tholeiite, with reconnaissance results for alkali basalts.

Ringwood and Green (1966) reported that the density changed in continuous fashion through the gabbro-eclogite transition interval, but Ito and Kennedy (1970) have demonstrated that there are two sharp density changes within the transition interval at temperatures near the solidus. They also demonstrated that by seeding their starting materials with garnet, the boundary for the appearance of garnet was lowered by 5 kbars compared to that located for the same material, in unseeded runs, by Cohen *et al.* (1967). This indicates some of the problems involved

in subsolidus experimental work with gabbroic compositions, even at temperatures close to the solidus.

There are now several experimental studies dealing with the phase relationships through the melting interval of various basaltic rocks. These include those of Yoder and Tilley (1962), Green and Ringwood (1967a), Cohen *et al.* (1967), Ito and Kennedy (1968), and Green T. H. and Ringwood (1968). The results agree in outline, but differ in detail. Figure 11 shows a generalized melting interval for a tholeiitic basalt, with the position of the gabbro-eclogite transition interval indicated below the solidus.

Addition of water to gabbro causes depression of the melting temperature, and introduces one hydrous mineral, an amphibole. Yoder and Tilley (1962) determined the phase relationships of a number of basalts between 1 bar and 10 kbars, in the presence of excess water. Lambert and Wyllie (1968, 1970 manuscript in preparation) determined the phase relationships for tholeiitic material between 10 and 25 kbars, in the presence of excess water; they did not reach the liquidus temperature. The shaded band in Figure 11 for the melting interval of gabbro in the presence of excess water is based on these results. Notice that the solidus changes slope at 15 kbars where plagioclase is replaced by jadeitic pyroxene (Boettcher and Wyllie, 1967; 1968; 1969). Note also that hornblende is stable subsolidus only to pressures near 25 kbars. The gabbro-eclogite transition interval is masked by the hornblende. These results are for excess water, and with oxygen fugacity being determined by the system. Tuthill (1969) reported the results of experiments on three basalt types at 5 kbars in the presence of excess water, with the oxidation state being controlled by buffers. The solidus temperature was independent of the basalt composition and the buffers used, but the stability fields of iron-bearing minerals did vary. Holloway and Burnham (1969) reported results obtained for a tholeiite at 2, 5 and 8 kbars pressure, under the Ni-NiO buffer, in the presence of water and carbon dioxide. Reducing the water content of the volatile components to 58.5 mole percent produced an increase in solidus temperature by 155°C at 8 kbars.

*Mineral facies for peridotite-gabbro-water.* The mineral facies for peridotite and gabbro, dry, are illustrated in Figure 9 and, in part, in Figure 11, and they have been reviewed recently by MacGregor (1968), O'Hara (1968), and Ito and Kennedy (1968). We are interested not in the effect of excess water on these mineral facies, as depicted in Figures 10 and 11, but on the effect of traces of water.

Given the results of Figure 11, we can deduce the general pattern of phase relationships for a gabbro in the presence of a small quantity of water, and Figure 12 illustrates such an isopleth. A few tenths of a percent of water would depress the liquidus temperature only slightly compared to the dry liquidus. Within the stability field of hornblende, the small amount of water would be fixed in the hornblende and no melting would occur. Figure 12 shows two vapor-deficient fields for hornblende-gabbro and hornblende-

precipitated peridotite assemblages. Peridotite in the upper mantle may thus be of at least three types: the original peridotite, magma-depleted residual peridotite, and peridotite precipitated from basic magma. The relationships among these three types, although still debatable, are becoming better understood through phase equilibrium studies on ultrabasic and basic compositions. Reviews of such experiments have been provided by O'Hara (1965, 1968), Green and Ringwood (1967a), and Ito and Kennedy (1968).

Extrapolation of the available data for gabbroic compositions to temperatures below 1000°C, which is not very satisfactory at this stage, indicates that gabbroic material would exist as eclogite in the mantle, except in regions of unusually high heat flow where garnet granulite could become stable (see Fig. 11). Gabbro (pyroxene granulite facies) could crystallize from magma at high levels in the suboceanic mantle, but as the temperature falls to normal, this should be converted to garnet granulite and eclogite. Press (1969) incorporated this process into his model for the suboceanic upper mantle. The occurrence and distribution of eclogite and garnet granulite in the mantle have been reviewed by Ringwood and Green (1966) and Ito and Kennedy (1970). The mineralogy of these materials is rather sensitive to composition. Eclogite consists essentially of garnet plus omphacite, with possible additional minerals orthopyroxene, quartz, or olivine. Rocks in the garnet granulite facies have the same mineralogy, with plagioclase present as well.

Consider now a mantle composed of peridotite with 0.1 weight percent of water. This would be stored in hornblende in the upper mantle, giving a vapor-deficient assemblage. If carbon dioxide is present, an intergranular fluid would exist, but it would probably contain very little of the water. Superposing Figures 9 and 10 then shows that traces of hornblende would be present in spinel-peridotite. Hornblende-bearing garnet-peridotite could exist down to depths of about 80 kms beneath the shield regions, but the thickness of this layer is very sensitive to the geotherm. The higher geotherm in oceanic environments, for example, makes hornblende unstable within the spinel-peridotite zone, and no hornblende is likely in the garnet-peridotite layer beneath the oceans. At depths greater than the hornblende stability limit, any water present would be distributed between an intergranular pore fluid and phlogopite. Phlogopite remains stable to depths of 100 kms or more in the suboceanic environment, and to much greater depths beneath the continents. With no intergranular fluid phase, phlogopite-peridotite would begin to melt below the phlogopite curve in Figure 12, when water was released from the phlogopite through reaction with olivine or pyroxene or both, in the peridotite. Kushiro *et al.* (1968) discussed the partial melting of peridotite in the presence of traces of water. Despite the dearth of reliable experimental data, it appears that mantle peridotite with a trace of water will undergo incipient melting at a depth of 80 to 100 kms in the suboceanic environment, but not until very much greater depths beneath the shield regions. The amount of liquid

developed would be extremely small, and strongly dependent upon the amount of water present.

Figure 12 is our guide to the mineralogy of an upper mantle layer composed of eclogite (Press, 1969) in the presence of 0.1 weight percent of water. Eclogite with a trace of hornblende and no intergranular fluid phase (unless carbon dioxide is present) exists down to depths of 60 to 80 kms, depending upon the geotherm. In regions with high temperature gradients, plagioclase would be present as well, in a hornblende-garnet granulite. The vapor-deficient hornblende-eclogite is followed by eclogite with any water present being contained within traces of interstitial magma of intermediate composition. The composition of this interstitial magma changes little with depth, and the amount is controlled almost entirely by the water content. No significant amount of liquid forms unless the geotherm reaches the solidus for dry eclogite (heavy shading in Fig. 12). Extrapolation of Figure 12 indicates that eclogite with interstitial magma would extend far below 110 km, as long as water is present.

*Mineralogy of the upper mantle and low-velocity zone.* The upper mantle contains peridotite of three types, undepleted (original), residual, and precipitated, with layers and pods of eclogite. Residual peridotite may be dominant beneath the continents. According to Press (1969), there is at least 50 percent eclogite in the suboceanic mantle between depths of 80 and 150 km. There is evidence that traces of water are present. Superimposed on this heterogeneous mantle is a concentric succession of mineral facies. The depths to transitions between facies are sensitive to temperature variations within the earth. Local variations in temperature cause migrations of phase transitions with tectonic effects at the surface.

The uppermost mantle is composed of hornblende-spinel-peridotite and hornblende-eclogite, with most of the water present being stored in crystalline phases. Hornblende-plagioclase-peridotite may occur in oceanic regions with high temperature gradients. The presence of hornblende-garnet-granulite is inferred for regions where the uppermost mantle has abnormally low seismic velocities. A number of changes may occur within the depth interval 60 to 100 kms, depending upon the temperature distribution (see Figs. 9, 10, and 12), but at depths greater than 100 kms, the upper mantle is composed essentially of garnet-peridotite and eclogite, with phlogopite as an accessory mineral present in low average concentrations; residual and precipitated peridotite should also be present, but presumably these decrease in proportion with increasing depth. Residual and precipitated peridotite may be more abundant beneath stable continental shields than beneath the oceans. These materials constitute the upper mantle down to the beginning of the transition zone, at about 350 kms.

There have been many attempts to explain the existence of the low-velocity zone in the upper mantle in terms of departure from the critical temperature gradient for which the seismic velocity remains constant, and in terms of

chemical or mineralogical zoning. Ringwood and his associates have been the most inventive in developing mineralogical models to match the geophysical measurements. Green and Ringwood (1967b) reviewed some of the earlier explanations and, using an upper mantle model based on their experimental study with synthetic peridotite, they demonstrated that the mineralogical zoning produced by changes in peridotite facies with depth was quite capable of explaining a low-velocity zone. They showed further that mineralogical variation along a geotherm may even yield two low-velocity channels (for  $V_s$ ) at depths of about 65 kms and between 100 and 150 kms.

These explanations have not proved adequate to account for the measured seismic velocity profiles and, in recent years, geophysicists have been stating that melting is required to explain their observations. Knopoff (1969), for example, reported that  $S$ -wave velocities are so low in some channels that it is unlikely that they can be produced without involving some partial melting (see also Anderson, 1970). In this and other statements by geophysicists, the assumption has been implicit that partial melting occurred because the geotherm rose above the dry solidus of peridotite or eclogite. Figures 9 and 11 indicate that this explanation requires rather higher temperatures in the uppermost mantle than appear to be reasonable. It also requires that the geotherm be rather sensitively controlled and constrained close to the solidus, lest too much liquid is produced. This explanation, with the problems introduced, has been reviewed recently by Birch (1969).

Almost simultaneously, Kushiro *et al.* (1968) and Lambert and Wyllie (1968) published experimental data which produced reasonable explanations for the low-velocity zone in terms of incipient melting. Kushiro *et al.* (1968) described the effect of water on the melting of peridotite, and concluded that partial melting of mantle peridotite could play an important part in the formation of the low-velocity zone, provided that water exists. As illustrated in Figure 10, the melting could begin at depths varying between about 70 and 100 kms, depending upon the temperature distribution and other factors.

Lambert and Wyllie (1968) presented their phase diagram for the system gabbro-water, similar to that given in Figure 11, with the location of the boundary for the pressure-sensitive breakdown of hornblende. They concluded that the low-velocity zone in the upper mantle corresponded to the level where hornblende became unstable, because at greater depths traces of water in eclogite would cause formation of interstitial magma. This scheme is illustrated in Figure 12, which indicates that the depth where the zone of incipient melting begins is sensitive to the temperature gradient, and that variation in temperature at depth would cause the boundary between vapor-deficient hornblende-eclogite and a partially melted eclogite to migrate up or down. The amount of liquid in the lightly shaded area is more dependent on water content than on temperature. The hornblende-eclogite would act as a trap for water rising from deeper levels in the mantle, because

the water would become fixed in hornblende, until the rock became saturated with water.

Ringwood (1969b) independently deduced that incipient melting in the presence of water was a likely explanation for the low-velocity zone. He noted that "amphiboles are not generally stable above 1200°C (1000°C in some systems) at pressures greater than 20 kbar." Therefore, most of the water present in the mantle will be present as a free phase at depths greater than about 60 kms, and this would lower the temperature of beginning of melting of mantle peridotite by about 200°C. This leads to conclusions very similar to those of Kushiro *et al.* (see Figs. 9 and 10).

Press' (1969) evidence that an eclogite layer exists in the suboceanic mantle between 80 and 150 kms depth provides a ready explanation for the low-velocity layer in terms of hornblende instability, as depicted in Figure 12. Incipient melting of peridotite with pods and layers of eclogite, in the presence of traces of water, may be adequate explanation for the zone beneath parts of the continents, at least, but if the shield regions overly barren, residual mantle material, then recourse to mineralogical and chemical zoning may be required. Green and Ringwood (1967b) showed that temperature-pressure effects superimposed upon mineralogical and chemical zoning can produce complex and subtle changes in seismic velocity distributions.

If the low-velocity zone is caused by incipient melting in the presence of water, the question arises as to why the zone terminates downwards. Presumably the interstitial liquid is not present beneath the low-velocity zone. There would be no liquid if the water content decreased with depth becoming vanishingly small below this zone; or if water was stored in high-pressure hydrous crystalline phases, as proposed by Sclar *et al.* (1967) and discussed by Lambert and Wyllie (1968, 1970, manuscript in preparation). In the suboceanic mantle, the bottom of the zone could correspond to the change in composition from Press' eclogite layer to underlying peridotite; if eclogite and peridotite both contained the same small percentage of water there would be very much less liquid present in the peridotite than in the eclogite. The melting interval of hydrous eclogite or hydrous peridotite begins to trend almost parallel with the dry melting interval at higher pressures (see Figs. 11, 10; and in Kushiro *et al.*, 1968, Fig. 1). It is conceivable that as the temperature gradients decrease with depth, the geotherm may pass through the solidus into the subsolidus region at the depth corresponding to the base of the low-velocity zone.

Where peridotite is present in the upper mantle between 60 and 100 kms, or deeper, the instability of hornblende and the small proportion of phlogopite permitted by potassium abundances suggest that there may be an interstitial, hydrous pore fluid (Fig. 10). This fluid may contain a high proportion of elements not readily accommodated in the peridotite minerals; these would include potassium, uranium, and thorium, the elements so important in the thermal budget of the earth. The quantity, composition, distribution and migration of such a pore fluid thus be-

comes a matter of considerable importance with respect to major geological processes. Where eclogite occurs in the upper mantle below 50 kms, the same elements may be concentrated in interstitial silicate magma. The eclogitic low-velocity zone may be a vast reservoir of andesitic magma, present in quantities too small to permit migration except under the most favorable tectonic conditions such as prevail, perhaps, in the island arcs and active mountain chains. If the temperature rises sufficiently that the geotherm enters the dark-shaded band in Figure 12, then the interstitial andesitic magma becomes incorporated in basic magma. Similarly, if basic magma rising from greater depths passes through an eclogitic low-velocity zone, it may incorporate interstitial andesitic magma from the layer, and this could possibly modify trace element abundances and ratios in the rising basic magma.

#### DISCUSSION

As we learn more about the upper mantle the more complex and heterogeneous it appears to be. There is reasonable agreement about the major element concentrations in the upper mantle, but we have very little reliable data about the concentrations of trace elements, including the radioactive elements K, Th, and U which produce the heat that drives and controls the geological processes near the earth's surface. It appears that many elements present in trace amounts may exist in interstitial films of fluid or silicate magma, or in phlogopite or amphibole in the uppermost mantle.

The continued study of ultramafic and mafic rocks in ap-

propriate associations undoubtedly will continue to provide data for more refined speculations about the upper mantle. A major point emphasized in this article is that for successful extrapolation from rock to the mantle, the petrogenesis must be understood; and the petrogenesis is complex. The petrologist needs the information that can only be obtained by the use of modern chemical techniques, and the geochemist can apply his data with maximum effect only if his samples can be located within an adequate petrological framework; this requires a detailed field map of the ultramafic association and the country rocks.

Additional experimental studies are required to define the mineral facies for appropriate materials at upper mantle conditions. We need to know the effect of variation in composition, with respect to both major and minor elements, upon the positions of the facies boundaries, and also the effect of variations in pressure and temperature upon the compositions of coexisting minerals within each facies. The effects of small proportions of water and carbon dioxide under mantle conditions must also be determined.

We must remind ourselves that it is a long extrapolation to the upper mantle in most locations, and the method of multiple working hypotheses remains as useful in the earth sciences today as when it was first propounded. There is no single relationship between ultramafic rocks and the upper mantle.

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