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Plutonic xenoliths and their relation to the upper mantle

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The best samples for chemical and petrological studies of the upper mantle are the ultramafic xenoliths carried up during volcanism, and those ultramafic rocks that may be tectonic slices of mantle transplanted into the crust. A geochemical survey of lherzolite xenoliths from a number of different localities indicates that regional differences do occur in upper mantle material and that these differences may be explained by partial fusion and extraction of the liquid phase. In some localities the concentrations of K and U in the xenoliths may be high enough for the xenoliths to represent primitive mantle material.

There is some evidence that isotopic equilibration between co-existing minerals in the mantle is not attained over very long periods. This might lead to isotopic anomalies in rocks derived from the mantle.

Relevance of chemical models for the upper mantle

The chemical composition of the upper mantle is important not only for its own sake but also for its application to problems of petrology and geophysics. For example, the amount and composition of basalt that could form by partial fusion is dependant on the concentration within the upper mantle of water and of the more fusible elements such as Na, K, Ca and Al. A better knowledge of the concentration of K, U and Th and their output of radiogenic heat in relation to the total heat flow from the mantle would tell us more of the distribution of these elements with depth, and hence perhaps the depth of convective recycling and homogenization of mantle material. Geophysical evidence indicates regional heterogeneity in the mantle, while some basaltic suites might also be explained in terms of geochemical provinces or heterogeneity. Chemical evidence for this heterogeneity is lacking. For these and other reasons, a good deal of attention has been given to chemical studies of materials of probable upper mantle origin.

There are three main sources of chemical data for the composition of the mantle, namely meteorites, especially carbonaceous chondrites, ultramafic rocks such as 'Alpine-type' peridotites, and the ultramafic xenoliths often found in volcanic material. About ten years ago, many Earth scientists thought that mantle rocks transported to the Earth's surface were so liable to chemical changes *en route*, and were so variable, that their compositions could not be used in defining a mantle composition. Instead, it seemed that meteorites as samples of planetary or asteroidal interiors provided better information about the interior of the Earth than did terrestrial crustal rocks. This seemed especially so, because at that time a major feature of meteorites was their apparent chemical uniformity, which encouraged belief in a general chemical uniformity between all of the inner, terrestrial planets. On this basis, meteorite analogies were applied to problems of the nature of the mantle, for example, by Lovering (1958) and Ringwood (1966*a*, *b*). Recently, however, it has been shown that there are significant chemical differences between the different groups of chondritic meteorites (see, for example, Van Schmus & Wood 1967), suggesting differences between each of the parent bodies from which the meteorites were formed. The chemical fractionation mechanisms that operated in

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the condensation of the materials of meteorite parent bodies (Larimer & Anders 1970) are likely to have led to even bigger differences between these bodies and the Earth. For example, a body of the composition of types I and II carbonaceous chondrites would have contained from 12 to 20% of water, whereas the primitive Earth probably contained only of the order of 0.1% of water. The Rb/Sr ratio of the Earth is 0.031 or less, while that of chondritic meteorites is 0.25, so that the Earth is depleted in rubidium relative to strontium by a factor of eight (Gast 1968). Similarly, the K/U ratio of the Earth appears to be only about a sixth of that of chondritic meteorites (Gast 1968). It would seem that the only valid quantitative application of meteoritic data to terrestrial problems is in the comparison of ratios of elements such as the rare earths, and of isotopes such as those of strontium, that are unlikely to have undergone much fractionation in the condensation of pre-planetary material and the aggregation and formation of the planets. Even here there may be difficulties (see Gast 1968).

Because of the non-quantitative nature of any meteorite analogy, attention has been refocused on terrestrial rocks. It is now accepted fairly widely that the types of material most likely to have been derived directly from the upper mantle, and therefore the best samples of it, are the ultramafic rocks of non-cumulate type. These include the massive bodies of ultramafic rock, and the ultramafic xenoliths (or foreign blocks) within volcanic debris.

Amongst the ultramafic rocks, the peridotites and serpentinites of 'Alpine' type and those dredged from the deep ocean floor are probably the most representative of mantle material. However, there are problems of interpretation. 'Alpine' peridotites range in composition from dunites consisting almost completely of olivine through to pyroxenites, and from almost anhydrous types through to serpentinites with 14% or more of water. To what extent do these chemical and mineralogical changes represent differences in the parent source material, or are they imposed on the rocks during their emplacement into the crust? In particular, have these rocks undergone some degree of partial fusion, and loss of the more fusible constituents as basaltic liquid? Ringwood (1966*a*), in his 'pyrolite' model, suggests that primitive unaltered mantle is equivalent to three parts of peridotite (infusible residue) and one part of basalt (extracted liquid). On the assumption that the wide range in composition of ultramafic rocks represents that from primitive mantle through to the infusible residue most depleted in fusible constituents, White (1967) has prepared histograms based on the published analyses of ultramafic rocks. The preliminary comparison of these with selected analyses of ultramafic xenoliths, suggests that both sources of data are of value, and that ultramafic materials, with high contents (about 3% or more) of CaO and Al_2O_3 are likely to be closest to the original mantle material (Harris, Reay & White 1967). However, although a number of studies of the geochemistry of ultramafic rocks have been made (e.g. by Stueber & Murthy 1966; Carswell 1968; Griffin & Murthy 1969), there has been little systematic sampling and analysis of large bodies of ultramafic rocks to find their variability and the changes that may have occurred during their derivation from the mantle.

ULTRAMAFIC XENOLITHS AS FRAGMENTS OF MANTLE

Plutonic blocks or xenoliths of ultramafic rock occurring in basaltic lavas and pyroclasts are recorded from more than 200 localities in the world (Forbes & Kuno 1965, 1967) and have been widely studied (e.g. by Ross, Foster & Myers 1954; Wilshire & Binns 1961; White 1966; Jackson & Wright 1970; Kuno & Aoki 1970). In some localities, such as the Massif Central in

France, the xenoliths are almost entirely of the rock-type lherzolite (olivine with lesser amounts of orthopyroxene and clinopyroxene, and often with small amounts of spinel), but elsewhere may be more variable. In Hawaii for example, although lherzolite is dominant in some occurrences, in others the xenoliths include such types as dunite (almost pure olivine) and websterite (ortho and clinopyroxene). Some apparent relationship between xenolith type and the nature of the host basalt was found by White (1966) in Hawaii, but later work there by Jackson & Wright (1970) shows that the different types of xenolith have an areal zonation which does not coincide with the areal zonation of the lava type, in the Honolulu volcanic series of Oahu. The conspicuous feature of ultramafic xenoliths is that the lherzolitic ones from all localities have a remarkable uniformity in mineralogy, as shown by Ross *et al.* (1954). The olivine normally contains about $91 \pm 2\,\%$ of the forsterite molecule. For example, in olivines from lherzolite xenoliths from the Dreiser Weiher in Germany and from five eruptive centres in the Massif Central of France, Paul (1970) found the composition of nine analysed olivines to fall in the range Fo_{89}–$Fo_{92.5}$, and the compositions of 33, determined by the less reliable methods of X-ray diffraction, to fall entirely in the range Fo_{87}–F_{94} (22 in the range Fo_{89}–Fo_{93}). This very restricted compositional range is in marked contrast to the wide range in the composition of the olivines of the host basalts and of ultramafic rocks of cumulate origin. Other features of the lherzolite xenoliths are the unusually high contents of Al_2O_3 in the ortho- and clinopyroxenes, often taken to indicate high pressures of formation. The textures of the xenoliths are those of metamorphic rather than cumulate rocks (see, for example, Ernst 1967).

Ultramafic xenoliths occur both in pyroclastic debris and in lava, but, as with other types of xenolith, there appears to be a general tendency that the more pyroclastic the eruption, the greater the possibility of xenolithic material being present. Presumably magmas of pyroclastic tendencies have higher gas contents and greater eruptive speeds through their conduits, than normal flow basalts. The host basalts are normally alkalic, e.g. alkali-olivine basalts, basanites, nephelinites etc., but occurrences have been reported also from tholeiites (White 1966), phonolite (Wright 1966) and trachyte (Wright 1969). However, the lherzolite xenoliths have uniform mineral compositions, irrespective of the host rock type.

In addition to these occurrences in host rocks of basaltic type, ultramafic xenoliths also occur in diatremes and dykes of kimberlite and related lamprophyric rocks (see, for example, Nixon, von Knorring & Rooke 1963; Carswell & Dawson 1970; Mathias, Siebert & Rickwood 1970). Although in these occurrences the range of rock-type among the xenoliths may be much greater, including eclogites and granulitic crustal rocks, of the ultramafic xenoliths the dominant type is garnet lherzolite (Mathias *et al.* 1970; Carswell & Dawson 1970). Again, there is a remarkable uniformity in mineral composition in ultramafic xenoliths from different localities (Mathias *et al.* 1970). The bulk chemical compositions of the garnet lherzolites (Carswell & Dawson 1970) overlap the compositions of lherzolites from basalt, and are virtually indistinguishable from them. The presence of garnet, the low content of Al_2O_3 in the orthopyroxene and other minor differences in mineralogy between the garnet lherzolites and lherzolites are attributed to the higher pressures or lower temperatures of equilibration of the garnet-bearing rock (O'Hara 1967, 1968; Boyd 1969). In both garnet-lherzolites from kimberlite, and spinel lherzolites from basalt, it seems that the xenoliths represent solid mantle torn from the conduit walls by the rapidly ascending magma, the differences in mineralogy being the result of different mineral facies and pressure–temperature environments within those regions of the upper mantle sampled by the volcanic erupta.

This view that ultramafic xenoliths are fragments of mantle transported to the surface, although widely supported (e.g. by Ross *et al.* 1954; Wilshire & Binns 1961; Ernst 1967; Harris *et al.* 1967; Kuno & Aoki 1970), has not been undisputed. Others have suggested that the xenoliths represent cumulates, the result of crystals settling from a body of magma, presumably related to the host basalt (see, for example, Frechen 1948; O'Hara & Mercy 1963). Some support for this appeared to come from the work of White (1966), which showed for Hawaii that the rock type of the xenolith varied with the type of host basalt. However, this relationship between xenolith and host rock has now been discounted by Jackson & Wright (1970). It is probable that certain types of xenoliths, e.g. of pyroxenite or of dunite of more iron-rich type (Fo_{87} or less) may be cumulates. However, the lherzolites and garnet lherzolites are so uniform in mineralogy that it is difficult to attribute them to crystallization from magmas similar to the host rocks, especially in view of the very wide range in composition of the latter. Also, the differences in strontium and lead isotope ratio between host basalt and lherzolite xenolith indicate that there can be no genetic relationship between them (Leggo & Hutchison 1968; Cooper & Green 1969; Paul 1970). So we are forced to the conclusion that lherzolite xenoliths are unrelated to the host rock, i.e. are accidental, and yet being so uniform in mineralogy are derived from some relatively uniform, ultramafic Earth-layer beneath both continents and oceans. By definition, this must be the upper mantle. However, even though we accept that the xenoliths are from the upper mantle, we still have to decide whether they represent complete mantle or an infusible residue depleted in its basaltic components. This problem is the same as that outlined for 'Alpine' peridotites—can a meaningful composition for the upper mantle be derived from material which may have undergone some unknown degree of selective removal of fusible components.

Of the more recent models for the composition of the upper mantle, the best known is the 'pyrolite' one of Ringwood (1966*a*, *b*), a hypothetical mixture of three parts of Alpine peridotite and one part of basalt. Although these proportions are somewhat empirical, the pyrolite model appears to receive some confirmation because its composition can be calculated independently from a meteorite model. This calculation requires the following initial assumptions (Ringwood 1966*a*):

(*a*) The primitive Earth had the same composition as type I carbonaceous chondrites, but subsequently underwent a high degree of reduction and oxygen loss.

(*b*) The mantle has an $FeO/(FeO+MgO)$ ratio of 0.12.

(*c*) The core contains sufficient silicon (11% by mass) to give a metal/silicate ratio in the Earth of 31:69.

Ringwood's 'chondritic' model gives a composition of the whole mantle; his 'pyrolite' model for the upper mantle. The similarity of the two models, if it is significant, requires that upper and lower mantle are virtually identical in chemistry.

An alternative approach was that of Harris *et al.* (1967) who assumed that ultramafic xenoliths and rocks represented a range in composition from undepleted upper mantle material through to material depleted in Na, K, Ca, Al, etc., by partial fusion and loss of the liquid phase. Those xenoliths highest in CaO and Al_2O_3 were thought to be closest to undepleted mantle. However, this preliminary study was based on relatively few analyses quoted from the literature, and complete descriptions of the analysed material were not always available. Following this, an investigation of the chemistry of ultramafic xenoliths from a wide range of localities was initiated at Leeds University, partly to provide more data on the geochemistry of ultramafic xenoliths, in their relevance to models of mantle composition, and partly to look for regional variations in composition (see Hutchison, Paul & Harris 1970).

Regional variation in xenolith composition

There is evidence of regional variation in the physical or chemical nature of the upper mantle. For example, there are differences from area to area, in seismic velocities. These could be due to regional variations in temperature, with higher velocities in areas of lower geothermal values, where the deeper the isogeotherms, the shallower will be the boundaries between mineral facies and phase-transformations. Alternatively, some of the variation in seismic velocity could be due to regional chemical differences which in turn would control the mean atomic mass, mineral assemblage and rock density, and proportion of interstitial liquid phase, all of which influence seismic velocity. Regional differences in chemistry might also be suggested on petrological or geochemical grounds. Some unusual suites of basic volcanic rocks, such as the potassic lavas of central Italy or the western rift in East Africa, could be the result of regional geochemical anomalies in the magma source zone of the upper mantle. Even the differences between oceanic and circumoceanic basalts (Chayes 1964) or within individual basalt provinces could be ascribed to mantle heterogeneity, although in practice other explanations are preferred (Jamieson & Clarke 1970).

Very recently, mantle heterogeneity has been demonstrated in Hawaii (Jackson & Wright 1970), from the zonation of xenolith type. This zonation is ascribed to the effects of partial fusion and basalt eruption.

To study this possibility of regional chemical variation in the mantle, up to a hundred samples of ultramafic xenoliths were collected from each of seven localities in the Massif Central in France (by R. H.) and from the Dreiser Weiher area of the Eifel district in Germany (by D. K. P.). Further collections have been made from Lanzarote, Canary Islands (R. H.), the Red Sea, Fernando Po, Australia and elsewhere by other collaborators. So far, about 200 specimens from France and Germany have been analysed by chemical and X-ray fluorescence techniques. The intention was to study the variation in composition within a single locality, and then to see if there were significant chemical differences between different localities of a single province. A preliminary account of the results is given by Hutchison *et al.* (1970).

The results (Paul 1970) indicate a reasonable degree of coherence in composition within a single locality, and significant chemical differences between localities. Statistical analysis of the variance showed TiO_2 to have the most conspicuous differences between localities, with all other elements except SiO_2 and Cr_2O_3 to have highly significant differences. Bivariate analysis showed a strong positive correlation between the pairs Al_2O_3–SiO_2, Al_2O_3–TiO_2, Al_2O_3–CaO, Al_2O_3–Na_2O, NiO–MgO, FeO–MnO, CaO–Na_2O, TiO_2–Na_2O, CaO–TiO_2, and a negative correlation between the pairs MgO–SiO_2, MgO–TiO_2, MgO–Al_2O_3, NiO–CaO, MgO–CaO, NiO–TiO_2, NiO–Al_2O_3, MgO–Na_2O. These differences and correlations are consistent with the idea of a relatively uniform original upper mantle on which have been superimposed varying degrees of extraction or depletion in the fusible or liquid components. For example, the concentrations of CaO and Al_2O_3 in the xenoliths vary proportionately with each other (see Fig. 6.5*a* of Paul 1970), as would be expected if CaO and Al_2O_3 were extracted in almost equal proportions in basalt.

The xenoliths from Dreiser Weiher appear to be more depleted in Ti, Ca, Al, K and perhaps

Na, and higher in Mg than those from the Massif Central. Among the French localities, Monistrol d'Allier in Haute-Loire, and to a lesser extent, Tarreyres in Haute-Loire appear to have the least depleted xenoliths, while from Montboissier, Puy-Beaunit and Volcan de Zaniere, all in Puy-de-Dome, the xenoliths are more strongly depleted.

If the xenoliths represent fresh, unaltered mantle from beneath each locality, then the upper mantle beneath Monistrol d'Allier appears to be relatively undepleted, compared with elsewhere. At the same time, one would need a greater number of sample localities before it would be possible to confirm that the mantle of the Haute-Loire area in general was less depleted than that of the Puy-de-Dome area. To date it has not been possible to compare oceanic and continental regions or cratonic and non-cratonic occurrences, although some collections have been made with this in mind.

If the variations in composition between individual xenoliths, and between xenolith populations from different localities are due largely to varying degrees of depletion in the more fusible constituents, it is desirable to know if this depletion immediately preceded eruption or had no time relationship. That is, does the xenolith represent the refractory residue left by the extraction of a liquid phase and the host basalt this extracted liquid, or is there no genetic relationship between xenolith and host basalt? As mentioned earlier, Leggo & Hutchison (1968) found the strontium isotopic ratios to be different for xenoliths and host basalts from the Massif Central, while Paul (1970) found a similar difference in material from Dreiser Weiher. As will be discussed in a later section, Paul (1970) found, in a xenolith from Dreiser Weiher, an apparent isochron of 500 ± 200 million years, indicating no isotopic equilibrium between the co-existing minerals of the xenolith for hundreds of millions of years. These two features indicate that the Quaternary volcanic episodes in which the xenoliths were erupted were not accompanied by melting in those higher zones of the mantle sampled by the ascending magma; i.e. the xenoliths represent real and complete upper mantle material. No doubt like all mantle material this had undergone some earlier changes by removal or addition of liquid components in the past, so is not representative of primitive mantle, but the xenoliths are fragments of actual modern mantle.

If the xenolith populations represent different aspects of a regionally heterogeneous mantle, it is instructive to look at the material least depleted in the fusible elements, to show how this conforms with ideas of mantle composition.

The composition of this undepleted material can be chosen in two ways, either the average for the least depleted locality, or by the selection of some analyses from all localities. Among the French localities, there are internal inconsistencies (Paul 1970). For example, samples from Montboissier are relatively high in K_2O and low in Na_2O, while those from Tarreyres are low in K_2O and high in Na_2O. The samples that are consistently least depleted in Ti, Ca, Al, Na and K are from Monistrol d'Allier. The average of the 20 analyses of spinel lherzolite from this area are given in table 1. Alternatively, following a similar procedure to that of Harris *et al.* (1967), one could select those analyses highest in CaO and Al_2O_3 (85 of the 200 analyses, with both CaO and Al_2O_3 in excess of 2 %), or those analyses highest in Na_2O and K_2O (25 of 200 analyses with Na_2O 2000 parts/10^6 and K_2O 300 parts/10^6). Of the alternative compositions given in table 1, the average of the analyses of the xenoliths from Monistrol d'Allier seems preferable in that it should represent a regional composition of upper mantle material, relatively undepleted in fusible constituents.

This mantle composition differs only slightly from that of other models such as that of the 'pyrolite' of Ringwood (1966*a*). The main difference is in the amount of basalt that could be extracted. In the 'pyrolite' model, by definition 25% is basalt. In a mantle of the composition of the Monistrol d'Allier xenoliths, only 6% of basalt (oceanic tholeiite) could be removed before the residue was completely depleted in TiO_2, and 9% for Na_2O. Although it is often assumed that K_2O is the limiting element in deciding how much basalt could be formed in the mantle, this is not so if the basalt composition is that of oceanic tholeiite. If a mantle composition is assumed, equivalent to that of garnet lherzolite xenoliths from kimberlite (average of 15, calculated from Carswell & Dawson's data), TiO_2 and Na_2O again are the critical elements, the content of Na_2O permitting less than 7% of basalt to be extracted.

TABLE 1

	(1)	(2)	(3)	(4)	(5)	(6)
SiO_2	44.9	45.0	44.7	46.2	45.16	49.21
TiO_2	0.08	0.06	0.07	0.12	0.71	1.39
Al_2O_3	3.15	2.80	2.57	2.16	3.54	15.81
Cr_2O_3	0.43	0.45	0.43	0.35	0.43	0.04
Fe_2O_3	1.28	1.47	1.43	6.82	0.46	2.21
FeO	6.69	6.63	6.88		8.04	7.19
NiO	0.24	0.25	0.25	0.31	0.20	0.01
MnO	0.11	0.11	0.12	0.11	0.14	0.16
MgO	39.7	40.1	40.7	42.0	37.47	8.53
CaO	3.14	2.93	2.53	1.64	3.08	11.14
Na_2O	0.24	0.20	0.26	0.17	0.57	2.71
K_2O	0.042	0.025	0.085	0.15	0.13	0.26

(1) Mean of 20 analyses of spinel-lherzolite xenoliths (19 for Na_2O) from Monistrol d'Allier, Haute Loire.
(2) Mean of 85 analyses of spinel-lherzolite xenoliths with CaO > 2% and Al_2O_3 > 2%.
(3) Mean of 25 analyses of spinel-lherzolite xenoliths with Na_2O > 2000 parts/10^6 and K_2O > 300 parts/10^6.
(4) Mean of 15 analyses of garnet lherzolite xenoliths from kimberlites (excluding 1 high value for TiO_2) calculated from the data of Carswell & Dawson (1970).
(5) Pyrolite mantle (Ringwood 1966).
(6) Oceanic tholeiite (Melson, Thompson & Van Andel 1968).
Analyses 1 to 4 have been recalculated to total 100%.

The garnet lherzolite analyses are of some interest in that in the past it has been suggested that the uppermost parts of the upper mantle are likely to be more depleted in fusible constituents, than the lower zones. Some support for this appeared to be found in earlier analyses, which showed the garnet peridotites from kimberlite, representing the deeper, less depleted layers, to be higher in Ca, Al, Na, etc., than the xenoliths from basalt. Although this may be true for some areas and some collections of analyses, the generalization now seems less likely, and there appears to be a complete overlap in composition between spinel-lherzolites from basalt and garnet-lherzolites from kimberlite. Indeed, at first sight, table 1 would suggest that spinel-lherzolite was less depleted than garnet-lherzolite. However, this is a spurious impression from the comparison of unlike material—specially selected analyses of spinel-lherzolite and an average of all analyses of garnet-lherzolite. More analyses of garnet-lherzolite are required to see if significant differences between the two groups exist.

HEAT FLOW AND MANTLE COMPOSITION

One possible check on mantle composition is through the level of radioactive elements required to maintain the heat flow from the mantle. If we knew the value for a mean heat flow from the mantle, and assuming reasonable ratios of K:U:Th abundances, we could calculate average contents of K, U and Th necessary in the mantle. This would require the further assumptions that upper and lower mantle had the same radioactive content, and that heat could escape from all levels. Both these assumptions are unlikely to be true.

Alternatively, if the lower mantle has lost its original content of K, U and Th, these now being concentrated in the outer, convecting and recycled layer, and given actual analytical values for K, U and Th in this mantle material, it should be possible to calculate the thickness of this outer layer.

The mean surface heat flow is about 63 mW m^{-2} (1.5 μcal cm^{-2} s^{-1}) for both oceanic and continental areas. In continental areas, much of this heat is of crustal origin, while in oceanic areas, much of the heat flow will be the result of convective upwell accompanying sea-floor spreading. Perhaps the modal heat flow value in oceanic areas of 46 mW m^{-2} (1.1 μcal cm^{-2} s^{-1}) is close to the equilibrium one. In this paper, a value of 42 mW m^{-2} (1.0 μcal cm^{-2} s^{-1}) is assumed as a reasonable world average, across the mantle-crust boundary.

Analyses by Paul (1970) for the Monistrol d'Allier material have an average of 420 parts/10^6 K_2O (say 350 parts/10^6 K). Analyses by Dr N. Gale (Oxford) for uranium in 15 xenoliths supplied by R. H. and D. K. P. give widely variable results, ranging from 0.0007 parts/10^6 up to 0.18 parts/10^6, and the K/U ratios are equally variable from 0.6×10^3 to 191×10^3). However, a reasonable median value from the 11 specimens from the Massif Central is about 0.06 parts/10^6 U. This is higher than most values for ultramafic rocks. Wakita, Nagasawa, Uyeda & Kuno (1967*a*, *b*) found only 0.018 parts/10^6 U and 54 parts/10^6 K in lherzolite xenoliths from oceanic regions, but higher values of about 0.04 parts/10^6 U and 200 parts/10^6 K (median values) in xenoliths from continental regions. Green, Morgan & Heier (1968) in six lherzolite xenoliths from Australia found median values of about 0.02 parts/10^6 U and 50 parts/10^6 K.

Assuming a mantle containing 350 parts/10^6 K, 0.06 parts/10^6 U and 0.24 parts/10^6 Th, a surface heat flow of 42 mW m^{-2} (1.0 µcal cm^{-2} s^{-1}) could be provided by a mantle layer 900 km thick, with no radioactivity below this.

If the assumption is made that convective stirring and re-homogenization of mantle occurs down to the bottom of the Benioff Zone at 700 km, with the material below that depleted in radioactive elements, then this is close to the calculated value. The uncertainties are great, but the similarity in the two values suggests that the analytical values assumed for average upper mantle, of 350 parts/10^6 K and 0.06 parts/10^6 U, are within a factor of two of the real values.

In a similar calculation, Gast (1968) finds an original mantle composition of 0.04 parts/10^6 U, and a present uranium distribution of 0.0913 parts/10^6 in the upper mantle (35 to 400 km) and 0.0289 parts/10^6 from 400 to 1500 km (assuming a K/U ratio of 1×10^3, and Th/U of 4.0).

A comparison of K/U ratios for lherzolite xenoliths from different laboratories (Wakita *et al.* 1967*a*, *b*; Green *et al.* 1968; Paul 1970) and of K/U ratios in basalts and other crustal rocks suggests that further analytical data on the abundance of potassium and uranium, and on the K/U ratios in mantle materials are required.

STRONTIUM ISOTOPE RATIOS IN THE MANTLE

It is believed that basalts come from the mantle, and that the strontium isotope ratios of primitive basalts of about 0.702 to 0.705 ($^{87}Sr/^{86}Sr$) are inherited from the mantle. But alleged samples of mantle often have distinctly higher strontium isotope ratios. For example, Stueber

& Murthy (1966) found that ultramafic xenoliths had ratios of up to 0.7083, with bimodal peaks at 0.704 and 0.706, while 'Alpine' type ultramafic rocks ranged from 0.706 to 0.729. Other workers report equally high values for 'Alpine' type ultramafics. The most recent values for ultramafic rocks from the mid-Atlantic ridge vary from 0.706 to 0.723, whereas associated basalts and gabbros range only from 0.702 to 0.704 (Bonatti, Honnorez & Ferrara 1970). These results are of course extremely embarrassing—how can one derive basalt from a mantle of apparently different isotopic ratio? More precisely, since the basalt value is probably correct, being consistent also with meteoritic data, how does one explain why many of the strontium isotope values for apparent upper mantle materials are anomalously high.

Stueber (1969) and Paul (1970), on examining the isotope ratios in the constituent minerals separated from lherzolite inclusions, found that the olivine and orthopyroxene had significantly higher $^{87}Sr/^{86}Sr$ ratios than the clinopyroxene. The reason is that strontium enters the clinopyroxene structure preferentially and is at about 100 times the level found in orthopyroxene and olivine. The rubidium, on the other hand, has a more uniform distribution. The result is a relatively high Rb/Sr ratio in olivine and orthopyroxene and a low one in clinopyroxene. Therefore, the ratio of radiogenic ^{87}Sr to the non-radiogenic isotopes will increase more rapidly in olivine and orthopyroxene than in clinopyroxene.

Paul (1970) found that the olivine, orthopyroxene and clinopyroxene in a lherzolite xenolith gave a rough isochron of 550 ± 200 Ma, although the host basalt had an age of less than 1 Ma. Similarly, Roe (1964) found for Pacific dunites apparent ages of 700 Ma, although the eruptive ages were 200 Ma. Paul's (1970) data indicate that isotopic equilibrium between the co-existing minerals of the xenolith was not attained in 350 Ma, at least, even though the xenolith as a mantle fragment must have been at temperatures of the order of 600 to 1000 °C, for all that time. If isotopic equilibrium between coexisting minerals is so much more difficult to attain than we imagine, then perhaps the same difficulty will occur between partial melt and residual crystals. If one melts partially a lherzolite xenolith, clinopyroxene and orthopyroxene are the chief minerals to enter the melt, normally the clinopyroxene disappearing first, leaving a harzburgite residue. The liquid will have an initial $^{87}Sr/^{86}Sr$ ratio close to that of the clinopyroxene, because this contains so much strontium that it swamps the contribution from orthopyroxene and olivine. Therefore the $^{87}Sr/^{86}Sr$ ratio of the initial liquid will be low, and the ratio of the residual solid will be high. The question remains unanswered, whether this initial liquid will re-equilibrate isotopically with the coexisting crystals, or can be removed with its $^{87}Sr/^{86}Sr$ ratio unaltered. If isotopic equilibration is not attained, then this could explain the anomalies between basalts and ultramafic rocks. The explanation might also be applied to the inheritance of rare earth patterns from individual mineral species, during partial fusion within the mantle.

Hydrous minerals in the mantle

In this review, there has been no mention of the possibility of hydrous minerals within the mantle, and only a brief mention of these can be made now.

The mantle possibly contains about 0.1% of water, perhaps in hydrous minerals like amphibole or mica, or in interstitial liquid. Oxburgh (1964) and Griffin & Murthy (1969) have discussed why hydrous mineral phases should occur; Varne (1970) and Aoki (1970) have described amphibole-bearing xenoliths, and Dawson & Powell (1969) phlogopite-bearing xenoliths. Some of these xenoliths differ from lherzolite ones in their chemistry, e.g. having relatively high contents of Na_2O and K_2O (Aoki 1970), or of water (Varne 1970).

The possibility of hydrous minerals in parts of the mantle is of considerable petrological interest. Bose (1967), Carmichael (1967), Griffin & Murthy (1969), Dawson (this volume, p. 297) and others have invoked the selective fusion and removal of these hydrous minerals to explain the genesis of some alkalic magmas, such as nephelinites, leucite-basalts and kimberlite.

In the selective fusion of hydrous minerals, the mineral disappears, so that the remaining assemblage is liquid–olivine–pyroxene. If equilibration between liquid and crystals occurs, then the liquid should be the same as that formed directly from olivine and pyroxenes of the same total chemistry. On this basis, the liquid composition should be controlled by the bulk chemistry of the system, and by the minerals remaining and in equilibrium with the liquid, and not be affected by the prior existence of minerals now disappeared. However, as the previous section emphasizes, our knowledge of the rate of equilibration between liquid and crystals is so incomplete that no firm conclusions are possible.

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