

N O T I C E

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ABSTRACT

The basalt, eclogite, and harzburgite that are the differentiation products of the Earth appear to be trapped in the upper mantle above the 670 km seismic discontinuity. It is proposed that the upper mantle transition region, 220 to 670 km, is composed of eclogite which has been derived from primitive mantle by about 20% partial melting and that this is the source and sink of oceanic crust. The remainder of the upper mantle is garnet peridotite which is the source of continental basalts and hotspot magmas. This region is enriched in incompatible elements by hydrous and CO₂ rich metasomatic fluids which have depleted the underlying layers in the L.I.L. elements and L.R.E.E. The volatiles make this a low-velocity, high attenuation, low viscosity region. The eclogite layer is internally heated and it controls the convection pattern in the upper mantle. Material can only escape from this layer by melting. The insulating effect of thick continental lithosphere leads to partial melting in both the peridotite and eclogite layers. Hotspots and ridges mark the former locations of continents. Most of the basaltic fraction of the oceanic lithosphere returns to the eclogite layer.

Plate tectonics is intermittent. The continental thermal anomaly at a depth of 150-220 km triggers kimberlite and carbonatite activity, alkali and flood basalt volcanism, vertical tectonics and continental breakup. Hot spots remain active after the continents leave and build the oceanic islands. Mantle plumes rise from a layer of about 220 km. Mid-ocean ridge basalts rise from the depleted layer below this depth. Material from this layer can also be displaced upward by subducted oceanic lithosphere to form back-arc basins.

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(NASA-CR-162602) A GLOBAL GEOCHEMICAL MODEL
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INTRODUCTION

Although convection plays an important role in plate tectonics and heat transport in the Earth it has not succeeded in homogenizing the mantle. Magmas are still being produced from mantle reservoirs which have remained separate for the order of 1 to 2×10^9 years (e.g., De Paolo, 1979; Sun and Hansen, 1975). Oceanic lithosphere is continuously returned to the mantle but the difference in element ratios in the reservoirs, e.g., Rb/Sr, U/Pb, Th/Pb and Sm/Nd, persists. If the depth of earthquakes in subduction zones can be used as a guide, oceanic lithosphere is presently being delivered to the region of the mantle between about 220 km and 670 km. The isotopic data can be satisfied if this is also the source region for mid-ocean ridge basalts (MORB). This leaves the upper mantle or the lower mantle as the source region for continental flood basalts (CFB), hotspot magmas and ocean island basalts (OIB). The upper mantle low-velocity zone (LVZ), or asthenosphere, is the more likely source region since temperatures there are closest to the melting point.

Ocean floor basalts have comparatively uniform and low $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, and $^{144}\text{Nd}/^{143}\text{Nd}$ ratios whereas continental magmas and basalts from ocean islands not associated with island arcs have less uniform and higher ratios (De Paolo and Wasserburg, 1979). The latter magmas are also enriched in volatiles and the incompatible large-ion lithophile (LIL) elements (White and Schilling, 1978; Frey et al., 1978). The study of isotopes has introduced the time constraint that reservoirs with different element ratios -- Rb/Sr, U/Pb, Th/Pb, and Sm/Nd -- have existed for the order of 1 to 2 b.y. The source region for MORB has been providing uniform composition lavas for long periods of time.

It must therefore be immense in size and global in nature (Schilling, 1975). The reservoir for continental and ocean island magmas also appears to be relatively uniform and global although its products are often mixed with varying amounts of MORB before it is sampled.

There are two competing petrological viewpoints regarding the nature of the source regions. The common view is that all basalts represent various degrees of partial melting of a garnet peridotite. The alternative position is that some basalts represent extensive melting of a deep eclogite source. Both eclogite and garnet peridotite inclusions are common in kimberlite pipes. The eclogite inclusions, although not rare, represent only about 20% of the total. This suggests that eclogite is either a less abundant component of the mantle or it occurs deeper than the garnet peridotite, as befits its higher density. Neither of the two types of fragments can represent primitive mantle (Allsop et al., 1969). They must therefore be a result of a previous differentiation event. The eclogite minerals are depleted in the trace elements which are enriched in peridotite nodules, the plume source region and the continental crust. It is therefore desirable to test the hypothesis that eclogite, peridotite and continental crust are the principle products of mantle differentiation and that xenoliths in kimberlites may be samples from the mantle source regions. If true, this would have considerable impact on our ability to model the composition and evolution of the mantle.

CHEMICAL STRATIFICATION OF THE MANTLE

The mantle is also heterogeneous in its seismic properties. It has not been clear, however, if or how the seismological and geochemical heterogeneities are related. The largest lateral variations in seismic

velocities occur in the outer 200-250 km of the Earth and are related to such surface tectonic features as shields, trenches, rises, and volcanic belts. The mantle is also inhomogeneous radially with the lithosphere, asthenosphere, and transition zone being the main subdivisions of the upper mantle.

A chemically layered upper mantle could provide distinct and isolated reservoirs and is more suitable in many ways than models involving isolated blobs or regional inhomogeneities (Hofmann et al., 1978). It has been proposed that the low-velocity zone is the depleted reservoir and the source of mid-ocean ridge basalts (Schilling, 1973). Plume basalts, i.e., magmas from the L.I.L. enriched reservoir, have been attributed to deeper sources. If the LVZ is enriched in volatiles, as proposed by Anderson and Sammis (1970) on geophysical grounds, then this explanation is untenable. Frey et al. (1978) have discussed other objections to this model. They argue that volatiles should have enriched the upper layers of the mantle.

On the basis of seismic velocities and seismicity patterns, Anderson (1979c) proposed that there were chemical discontinuities in the mantle at 220 and 670 km. The former is the base of the LVZ and near the maximum depth of earthquakes in continental collision zones and regions of subduction of young, <50 Ma, oceanic lithosphere. The latter is a sharp seismic discontinuity and is near the maximum depth of earthquakes. Only old oceanic lithosphere penetrates this deep. The seismic velocities between 220 and 670 km are consistent with eclogite.

The continental lithosphere extends no deeper than about 180 km (Anderson, 1979c). It may terminate at the boundary between granular and sheared lherzolite nodules, ~150 km (Boyd and Nixon, 1975). We

will assume that the sheared nodules are representative of the mantle below the lithosphere and above the Lehmann discontinuity at 220 km. The shallower granular nodules have apparently been subjected to basalt extraction since they contain less CaO and Al_2O_3 than the sheared variety. They may be an important, perhaps major, component of the continental lithosphere. Both varieties of nodules are enriched in the L.I.L. elements compared to oceanic crust, the MORB source region and the minerals of eclogite inclusions. The fertile peridotite presumably rises to shallower depths under the oceans, of the order of 80 km. Thus, the average thickness of the fertile peridotite layer is about 120 km. Volumetrically, this is an adequate source region for continental and hotspot magmas but not for the more voluminous MORB.

We suggest that differentiation of the Earth leads to two layers in the upper mantle, a thick basalt crust over residual peridotite. As the Earth cools the base of the original crust transforms to eclogite which sinks through the upper mantle. The present upper mantle is peridotite overlying a thick (450 km) eclogite section. Partial melting in the eclogite section allows material to escape and to melt extensively upon ascent. This is proposed as the source of oceanic crust.

The basalt and eclogite portions of the oceanic lithosphere return to the eclogite layer by subduction. The fertile peridotite layer of the upper mantle can partially melt and provide basalts when the upward convection of mantle heat is prevented by the insulation of continental lithosphere. Thus, both reservoirs are global and underlie both oceans and continents. It appears that the MORB source region can also provide magma to back-arc basins, perhaps when material is displaced by the descending slab.

THE ECLOGITE SOURCE REGION

Pipe eclogites have a strong resemblance to oceanic tholeiites in both the major and trace elements. The Rb/K and other ratios in bimineralic eclogite closely resemble the corresponding ratios in abyssal tholeiite basalts. The similarities are even more pronounced if the eclogites are compared with the average composition of the oceanic crust.

The first column of Table 1 gives a composition which is representative of oceanic tholeiites. More likely estimates of the composition of the primary magma are the total composition of the oceanic crust (column 2) and basaltic komatiites (column 3). These compositions are remarkably similar and have appreciably more MgO and less Al_2O_3 and Na_2O than tholeiites which are considered to be the last crystallizing liquid from a more primary magma. The bimineralic eclogites in kimberlite (columns 4, 5 and 8) are virtually identical to these estimates of the average composition of the oceanic crust. Trace element comparisons between kimberlite eclogites and abyssal tholeiites are given in Table 2. Again, the correspondence is remarkable.

It appears that material similar to eclogite inclusions in kimberlites is a suitable parent for the oceanic crust. The inclusions themselves may represent cumulates from mantle diapirs that were trapped in the continental lithosphere. Diapirs rising from such great depth would melt extensively if their ascent were unimpeded by the continent.

THE GARNET-PERIDOTITE LAYER

The K, Rb, and Sr contents of some kimberlite garnet peridotite inclusions are given in Table 2. Also given are estimates of CFB and

of the "plume" source. Note the agreement between tholeiites and eclogites and between peridotites and the inferred plume source region. Another way to estimate the trace element content of a partial melt from peridotite is to assume that the difference in composition between fertile and barren peridotite is due to basalt removal. The trace element content of the resulting liquid is given in Table 3 and compared with continental and oceanic basalts. The peridotite compositions are from Rhodes and Dawson (1975) and it is assumed that the basalts represent 20% melting. These are extremely fresh peridotite xenoliths from the Lashame tuff-cone in northern Tanzania that have apparently come from a depth of ~150 km. They are chemically and mineralogically similar to peridotite inclusions from kimberlites except that they appear to be relatively less contaminated. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of these samples are about 0.705. The inferred melt is much higher in K, Rb, and Sr than oceanic tholeiites, a characteristic of continental basalts. The K/Rb and Rb/Sr ratios are also much different than abyssal basalts. Fertile garnet peridotite therefore seems a suitable source material for continental flood basalts but not for MORB. It also has the characteristics inferred for the "plume" source region (White and Schilling, 1978). This part of the mantle has probably been subjected to metasomatic enrichment of the incompatible trace elements (Lloyd and Bailey, 1975). Such enrichment has also been proposed for the source region of continental (Boettcher and O'Neil, 1979) and plume basalts (White et al., 1979). The fact that enriched xenoliths are extensively sampled by kimberlites argues for the shallowness of the plume reservoir.

Mid-ocean ridge basalts generally have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between about 0.702 and 0.704 while continental basalts are usually greater than

0.704 and range up to 0.710 (Carter et al., 1978; De Paolo, 1979). Basalts from oceanic islands are intermediate in value and may represent mixtures. The data on kimberlite xenoliths is sparse and equivocal (Allsop et al., 1969; Barrett, 1975; Simazu, 1975). Pipe peridotites have $^{87}\text{Sr}/^{86}\text{Sr}$ values of 0.7060-0.7075 and other characteristics appropriate for the source region of continental basalts. Eclogite xenoliths may have been brought into the continental lithosphere by deeper diapirs and evolved for some time in an environment different from PEL prior to pipe eruption. Whole rock measurements on eclogite xenoliths from S. Africa generally have high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.704-0.711). Allsop et al. (1969) estimated the ratio in "ideal" bimineraltic eclogite as 0.702. A sample from Tanzania has a value of 0.7004; discrete diopside nodules give 0.7029-0.704 (Barrett, 1975).

LOCATION OF THE TWO SOURCE REGIONS

As discussed earlier, at least part of the oceanic lithosphere seems to be returned to depths between 220 and 670 km. The mantle discontinuities at these depths are sharp and they are associated with changes in seismicity, as if they were acting as barriers to slab penetration. This could be due to density jumps caused by changes in mantle chemistry. The isotopic data, although useful in finger printing the source regions and giving age control does not provide information about major element chemistry or intrinsic density. This is where the kimberlite inclusions become useful.

Eclogite is appreciably denser than garnet peridotite and should therefore occur deeper in a gravitationally stable mantle. I have suggested that the Lehmann discontinuity at 220 km is the boundary between garnet peridotite and eclogite, and the discontinuity at 670 km is the

boundary between eclogite in the garnetite assemblage and peridotite in the ilmenite plus spinel assemblage (Anderson, 1979b). The eclogite layer is perched (PEL) in the upper mantle and forms the transition region. In a convecting system composed of two superposed layers there is a thermal boundary layer, i.e., region of high thermal gradient, on each side of the interface. This is where temperatures are most likely to approach or exceed the solidus and where diapirs would originate. There is also a thermal boundary layer associated with the lithosphere-asthenosphere boundary. Temperatures at 670 km and below are likely to be well removed from the melting point.

Differentiation of a silicate planet results in two distinct products, basalt and residual peridotite. The basalt, resulting from low pressure, high temperature partial melting of primordial mantle possibly resembling peridotitic komatiite, would be originally concentrated in a thick layer at the surface. Subsequent cooling, primarily by convection, brings upper mantle temperatures into the stability field of eclogite which is denser than residual mantle. This leads to a massive overturning of the outer layers of the planet, subduction of the eclogite protoplate and destruction of the early geological record. This may explain the rarity of crustal rocks older than 3.8 Ga.

Given that planetary differentiation concentrates basalt in the outer layers and that the Earth has generated and subducted massive amounts of oceanic lithosphere what is the likely distribution of basaltic material in the interior? To answer this I have estimated the density as a function of depth for basalt and peridotite (Anderson, 1979a). Basalt below about 50 km converts to eclogite which is denser than normal mantle even after olivine has converted to spinel. Pyroxene

and garnet react at higher pressures to form a garnet solid solution. Normal mantle also undergoes a series of phase changes but remains less dense than garnetite until ilmenite and perovskite structures become stable below 670 km. The eclogite cannot sink below this level. The addition of Al_2O_3 to CaSiO_3 and $(\text{Mg}, \text{Fe})\text{SiO}_3$ expands the stability field of garnet and increases the pressure required for transformation to such dense lower mantle phases as perovskite and ilmenite. This means that eclogite cannot sink into the lower mantle. Whole mantle convection, which may have been possible prior to the establishment of the eclogite layer and the chemical discontinuities in the upper mantle, would be replaced by separate convection systems in the lower mantle, the eclogite layer and the upper mantle above 220 km.

The oceanic part of the plate tectonic cycle in this scheme is very simple (Figure 1). Heating of the eclogite layer causes partial melting and the rise of eclogitic diapirs. The latent heat for complete melting is provided by adiabatic decompression. Oceanic crust forms from this eclogite liquid. MORB forms the surface veneer and represents the latest freezing fraction. Subduction causes the crust to reinvert to eclogite and it sinks back to the PEL. The harzburgite part of the lithosphere remains in the upper mantle. The continuous recycling and remelting of the material in the oceanic crust depletes it in the LIL and, in particular, the LREE.

PRIMITIVE MANTLE

The isotopic data indicates that the two source regions are the results of an early differentiation event. If we accept the 220 and 670 km discontinuities as its boundaries, the eclogite layer represents about 20% of the mass of the mantle. By assuming that the whole mantle

was involved in this early differentiation we can obtain a spectrum of estimates of primitive mantle composition. Several of these are given in Table 4.

There are other approaches that have been used for estimating primitive bulk Earth chemistry. Ganapathy and Anders (1974) have provided a cosmochemical mixing estimate which is also given in Table 4. There is surprisingly good agreement between these estimates and the resulting compositions are distinct from any modern rock type. Peridotite komatiites are widespread in early (>3.5 Ga) Precambrian terrains. Viljoen and Viljoen (1969) propose that these approximate the composition of primitive mantle. Indeed, these have Mg/Si ratios in the range of whole Earth estimates. They may represent primitive mantle that has left some garnet in the source region. I propose that a material similar to those in Table 4 was the parent from which the current mantle reservoirs were derived. These reservoirs are a shallow peridotitic layer and a deeper eclogite layer. In this scheme pyrolite would not represent primitive mantle but mantle which has already been depleted in a basaltic component.

It has long been recognized that the source region of MORB is depleted in LIL elements compared to alkali basalts, continental flood basalts, and hot-spot magmas. One would expect, however, that the original primary differentiation would enrich the basalt/eclogite fraction relative to the residual peridotite. Whole rock analyses of pipe eclogite indeed show such enrichment. The major phases, omphacite and garnet are, however, depleted and the enrichment occurs in the intergranular material (Allsop et al., 1969). The intergranular material LIL content is similar to that of the continental crust (Table 2). The eclogite layer may have become depleted and the peridotite layer enriched

by the upward transport of fluids as discussed by Frey et al. (1978), Doettcher et al. (1979) and Mysen (1979).

Tatsumoto (1978) and Hedge (1978) proposed a model, based on lead and strontium isotopes that is similar to the present result, i.e., the LVL ("asthenosphere") is undepleted or enriched and supplies "hot-spot" magmas; the underlying mantle ("mesosphere") is depleted and provides abyssal tholeiites. This is the opposite of Schilling's (1973) model. Continental and hotspot related magmas represent a wide range of partial melting, from about 4% to 25% (Frey et al., 1978; White et al., 1979). This suggests that they come from a wide range of (shallow) depths. MORB's are invariably tholeiitic, indicating extensive (>25%) melting and a consistently deep origin.

The incompatible trace elements in both source regions are enriched relative to recent estimates of bulk Earth composition (Ganapathy and Anders, 1974). For example, if the lower mantle is identical in composition to the peridotitic upper mantle and if the silicate portion of the planet is 21% eclogite and 0.5% continental crust then the major elements are in agreement with bulk Earth estimates such as Ganapathy and Anders (1974) but such elements as K, Rb, and Sr are about a factor of 2-1/2 higher. This can be accounted for if the mantle below 670 km has transferred its incompatible trace elements to the crust and upper mantle. This presumably occurred during the early differentiation of the Earth and accompanied basalt extraction from primitive mantle. The calculated Rb/Sr ratio of the continental crust plus upper mantle (peridotite plus eclogite) is 0.028. This is also the value inferred for the bulk Earth (Ganapathy and Anders, 1974; DePaolo and Wasserburg, 1976).

This plus the complementary nature of the two source regions suggests that the material above 670 km may exhibit bulk Earth patterns of the LIL elements, and have a 2-3 times enrichment. The lower mantle is extremely depleted in the LIL elements.

EVOLUTION OF THE MANTLE

The evolution of the Earth's mantle according to the present scheme is shown in Figure 2. The primitive mantle has roughly the composition given by a 1:4 mix of eclogite and garnet peridotite. Early differentiation processes lead to the development of a thick enriched basalt crust and a residual peridotite mantle. As the outer layer cools it converts to eclogite which settles through the upper mantle and becomes perched near 670 km by a phase boundary in the peridotite mantle. Incompatible trace elements are removed into the continental crust and the upper mantle garnet peridotite regions by metasomatic fluids. The eclogite layer therefore becomes depleted in those components which are concentrated in the overlying layers including the continental crust.

The highest temperatures in the mantle, relative to melting temperatures, are in the thermal boundary layer near the top of the PEL, ~220 km depth. This is where the density contrast between eclogite

and peridotite can be overcome by partial melting and where eclogite diapirs originate. Peridotite diapirs originate from the top of the thermal boundary layer. Their shallower depth and the broad melting interval of peridotite leads to relatively small amounts of partial melting, a requirement of alkali basalt petrology (Frey et al., 1978). This is a persuasive but overlooked argument for a shallow location of the plume source region relative to the source region for tholeiites.

Continental and alkali basalts are usually emplaced at greater elevations than the oceanic tholeiitic basalts. This is sometimes taken as evidence that the MORB source region is shallower than the plume source region. Alternatively, the alkali basalts are emplaced at higher elevations because they are intrinsically less dense than tholeiites.

The parameters of the thermal boundary layer depend on the thermal properties, viscosity and heat flow at the interface. The thickness is calculated to be about 20 km and the temperature rise is 300-600°C. This is comparable to the near surface gradient and brings the average temperature close to the melting point of mantle silicates at 220 km. The thermal perturbation by a stationary continent, or a large continent moving slowly, causes a further temperature rise and may be the trigger that initiates partial melting.

Depleted peridotite, the refractory product of partial melting of the garnet peridotite layer, is lighter than any other component of the mantle and becomes part of the continental lithosphere and the suboceanic harzburgite layer. The lower mantle need not be involved at all in the current magmatic cycle.

MANTLE METASOMATISM AND THE REDISTRIBUTION OF TRACE ELEMENTS

Calculation of the partitioning of the rare Earth elements among the various regions indicated in Figure 2 for a dry planet give results which are contrary to observations (Schilling, personal communication). In particular the eclogite layer being the result of partial melting of primitive mantle should be LREE enriched even after removal of the continental crust. The peridotite layer should be depleted relative to primitive mantle and be LREE deficient. This suggests that there has been upward transport of a LREE phase which serves to deplete the PEL and enrich the overlying layers, including the peridotite plume source and the continental lithosphere. An H_2O -rich vapor phase strongly concentrates the REE and, in particular, the LREE (Mysen, 1979). Evidence that the mantle has experienced such metasomatism prior to the transport of samples to the Earth's surface by kimberlites has been provided by Ridley and Dawson (1975), Erlank and Shimazu (1977) and summarized by Mysen (1979). Parallel evidence from alkali basalts is given by Boettcher and O'Neil (1979). Carbonatites, kimberlites, "depleted" granular peridotites, alkali basalts and the interstitial phase in eclogite xenoliths all show extreme LREE enrichment. The source region of MORB on the other hand is LREE depleted. The degree of LREE enrichment in a water-rich fluid increases with garnet content (Mysen, 1979). These observations all suggest that a vapor or fluid phase removes the REE from the eclogite layer and deposits them in the peridotite layer and the continental lithosphere. The other LIL elements also show a complimentary pattern between MORB and plume basalts.

Since the enrichment of REE in general, and LREE in particular, in the fluid increases rapidly with pressure as well as garnet content, we expect an upward increase in LREE enrichment in the rocks as we proceed

from the eclogite to the fertile peridotite to the "depleted" peridotite (continental lithosphere) and finally to the crustal layers. Mass balance calculations suggest that the continental crust is not the only repository of LREE enriched material. The remainder we propose is in the continental lithosphere and the LVZ. This is supported by the evidence from material which has been sampled from these regions. The depleted nature of the MORB source region is an argument that it lies below the REE enriched regions, i.e., below the depth of generation of kimberlites and alkali basalts and the xenoliths they contain.

PLATE TECTONICS

A major source of mantle heat flow is the Perched Eclogite Layer (PEL). With tholeiitic concentrations it would provide $0.7 \mu\text{cal}/\text{cm}^2\text{sec}$, about $1/3$ of the global average surface heat flow. The scale length of convection in the PEL will be of the order of the layer thickness, ~ 450 km. Convection in the overlying peridotite layer would be driven by this non-uniform heating from below and would have a similar or smaller scale length and be characterized by narrow ascending plumes. The high thermal gradient in the boundary layer leads to a large decrease in viscosity at the interface and the two convecting systems may be thermally coupled rather than coupled by viscous drag forces. That is, cold descending regions would occur in close proximity in the two layers. A descending slab, for example, may trigger detachment of the cold boundary layer in the PEL. This is shown schematically in Figure 3. Seismic waves will see a continuous cold region with high velocities. There will also be other regions of cold descending plumes in the PEL which are not directly related to slabs. Likewise, there may be regions

of rising currents in both the PEL and the overlying layer, which do not express themselves in surface features such as mid-ocean ridges. The convective pattern, however, may be evident in detailed analyses of topography, gravity and seismicity.

Earthquakes do not extend below about 250 km at most convergent plate boundaries. In other regions there is a gap in the seismic zone between about 250 and 500 km. Even where the zone is continuous it is usually contorted near 250-350 km. In many cases the deeper zone is more-or-less spatially continuous with the upper zone but in Chile, Peru and New Zealand the two zones appear to be displaced. This is suggestive of the type of two-layered convection considered here.

There are several ways to estimate the lateral extent of the convection cells. We assume that convection in the eclogite layer controls convection in the thinner and shallower peridotite layer. Thiessen et al. (1979) suggested that the distribution of high spots in Africa reflects the underlying convection pattern. By comparison with laboratory data they inferred a vertical extent of convection of about 500 km, slightly greater than the thickness of the eclogite layer. They proposed that this pattern could only be observed through a stationery continent.

Jordan (1978) showed that terrain, crustal thickness, and Bouguer gravity anomalies have correlation distances of the order of 550 km, remarkably similar to African highspot distances. This again suggests a scale length of convection comparable to the thickness of the transition layer.

Menard (1973) attributed depth anomalies in the eastern Pacific and the bobbing motion of drifting islands to convection cells in the upper mantle of half-wavelength 250-500 km. The depth anomalies, having amplitudes of ± 300 meters can be explained by temperature

differences in a 200 km thick layer of about 200°C. The depth anomalies seem to be fixed relative to hot spots. Menard believes that motion of plates over these bumps explain many aspects of vertical tectonics. The "bumpy" asthenosphere envisaged by Menard is a natural consequence of the convection pattern proposed here.

The pattern of convection in the PEL is probably fairly complicated. The migration of trenches and continents may change the locations of the descending plumes in an individual cell or groups of cells in the PEL but it seems likely that the cells themselves cannot move far relative to one another. This provides a rationale for a fixed hot-spot reference frame and a mechanism for allowing the surface expression of a hot spot to wander on the order of 5°. The proposed upper mantle convection pattern is shown schematically in Figure 4.

HOT SPOTS AND PLUMES

Morgan (1972) suggested that island and sea-mount chains are produced by plate motion over convective plumes extending from near the core-mantle interface to the base of the lithosphere. Anderson (1975) proposed that plumes came from a distinctly different source region than midocean ridge basalts and that kimberlites, carbonatites and continental flood basalts were all related to hot-spot or plume activity. I suggested that plumes were a result of a thermal perturbation due either to continental insulation or a diapir rising from the deep mantle. In either case the plume source region differed in chemistry from mantle providing abyssal tholeiites and was the source region for distinctive ocean island magmas and undersaturated continental magmas. It was proposed that this source region was rich in Ti, Ba, Sr, Y, La, Zr, Nb, Rb, CO_2 , P_2O_5 , H_2O , etc., compared to "normal" mantle (the source

region of MORB). These characteristics have since been found to be typical of plume chemistry (Unni and Schilling, 1978; Schilling et al., 1976; Bonatti et al., 1977; White and Schilling, 1978).

Although I initially favored a deep mantle origin for plumes it now appears that they originate above the 220 km mantle discontinuity in a region of the mantle that has been enriched in incompatible components by metasomatic processes that have depleted the source region of MORB.

Since hot spot activity is restricted both in space and time we need a thermal anomaly to initiate partial melting in the peridotite layer. One possibility is thermal blanketing by the thick conductive continental lithosphere (Anderson, 1975). A large stationary continent may thereby cause its own break-up and a subsequent period of rapid plate motion. The temperature anomaly and partial melt zone remains after the continent leaves and it becomes an oceanic hot spot. Hot spots will have a finite lifetime which appears to be at least 200 Ma after the continent starts to move off.

With a mantle heat flow of $0.7 \mu\text{cal}/\text{cm}^2\text{sec}$ there are 2×10^9 cal/cm^2 delivered to the base of subcontinental lithosphere in 10^8 years. Even if only 20% of this is trapped this is enough to heat a 50 km thick section of mantle by 200°C and to melt it to the extent of 20%. Continental blanketing therefore seems to be an adequate mechanism for turning on melting spots below the continental lithosphere.

The hot spot tracks in the Atlantic and Indian Oceans can be traced back to continental interiors. The timing of Mesozoic and Cenozoic continental flood basalts in North and South American, Africa, Europe and Siberia is appropriate for their location over hot spots when they formed (Morgan, 1979). If our hypothesis concerning the origin of

hot spots under stationary continents is correct then we would expect that a continent would have been over Hawaii some 200 m.y. ago. The Hawaiian-Emperor seamount chain disappears into the Aleutian Trench so we cannot trace it beyond about 70 m.y. By backing up the Pacific plate we can infer that some of these continental fragments may have been incorporated into northwestern North America. The 210 m.y. old flood basalts from central Alaska to Northern Oregon, so-called Wrangallia (Jones et al., 1977), originated near the equator ($\sim 15^\circ$) and subsequently moved north to become attached to North America. Greenstones in central Japan were formed near the equator in the late Paleozoic (Hattori and Hirooka, 1979). Other fragments are elsewhere around the Pacific Margin (Nur and Ben-Avraham, 1979).

In the present scheme it is the thermal perturbation caused by the deep (~ 150 km) continental lithosphere that is responsible for the onset of hotspot activity. The hotspots generate kimberlites, carbonatites, alkali basalts, and lead to continental breakup. There follows a period of rapid continental drift and seafloor spreading. Continental igneous activity wanes as the continents drift off their hotspots but ridge and ocean island volcanism increase. In this scenario hotspots play an important, perhaps dominant, role in breaking up and driving the plates. At the end of an interval of rapid spreading they would tend to be centrally located in the oceans, much as they are today. An important force in plate tectonics may be the "hotspot fleeing force". The thermal perturbation by continents may also control locations of mid-ocean ridges (Nur and Ben-Avraham, 1979).

HOT SPOT PROPULSION

Elder (1976) has considered the propulsion of continents by a horizontal temperature gradient. In the absence of resisting forces at plate boundaries the velocity is

$$u = \alpha g \Delta T h^3 / 3 \nu l$$

where α is the coefficient of thermal expansion ($3 \times 10^{-5}/K$), g is acceleration due to gravity (10^3 cm/sec^2), ΔT is the temperature anomaly (200K), h is the layer thickness (100 km), ν is the viscosity (10^{20} P) and l is the horizontal scale of continent (1000 km). The assigned values are just for the purpose of obtaining an order of magnitude estimate for u which turns out to be about 6 cm/yr.

Periods of extensive continental magmatism are correlated on a global basis and seem to last of the order of 0.3 to 0.4 Ga (Windley, 1977). They are separated by periods on the order of 0.7 to 1.0 Ga. Periods of rapid plate motion, or at least of rapid apparent polar wander, last for about 30-60 Ma (Gordon et al., 1979). At a velocity of 10 cm/yr. this would lead to total displacements of 3000 to 6000 km which are of continental and inter-continental distances. The continents, on the average, then would come to rest far from their own or other continents' hot spots. The gestation period for forming a hot spot appears to be 200 to 400 Ma and the lifetime estimated from the duration of continental magmatism and the duration of the subsequent hot spot track may be as much as 500 Ma. Since convection is more efficient through the oceanic mantle we expect that hotspots will start to dissipate as soon as their continents move off and that regions of high heat flow in the oceans would mark the previous locations of stationery or slowly moving continents.

Reconstruction of the continents indicate that most of the Atlantic and Indian ocean hotspots were beneath continents from about 100 to >350 m.y. ago. The present African hotspots were apparently beneath Europe at the earlier time.

SUMMARY AND DISCUSSION

Isotopic evidence indicates that there are at least two source regions of basaltic magma in the mantle which have remained separate for the order of 1 to 2 Ga. The sub-continental and hotspot source region is enriched in incompatible components compared to the source region for midocean ridge basalts. Eclogite and garnet peridotite xenoliths in kimberlite pipes seem to have the appropriate characteristics to provide mid-ocean ridge basalts, and continental basalts, respectively. The eclogite layer, the source of mid-ocean ridge basalts, is denser, and therefore deeper than the enriched layer. Melting in both layers may result from the thermal insulation provided by the thick continental lithosphere.

The latent heat for extensive melting of eclogite diapirs is available if rapid ascent of 160 km or greater is possible (Yoder, 1976). This is less likely under thick continental lithosphere than in the oceanic asthenosphere. Therefore, oceanic tholeiites occur in the oceans and eclogite xenoliths occur under continents. The temperature rise required for the initiation of garnet peridotite diapirs may require a long period of continental insulation. Therefore, hotspots start under continents and lead directly to uplift and breakup and provide the initial driving force for continental drift.

It is possible that continental insulation is also required to initiate melting in the eclogite layer. The difference in geometric style between ridges and hotspots would then reflect the difference

in convection in the eclogite and peridotite layers; rising sheets in the internally heated eclogite layer and rising plumes in the overlying peridotite layer.

One would expect the basalts formed in the primary differentiation of the earth to be LREE enriched and enriched in the incompatible elements. Present MORB is coming from a source region which has been depleted in these elements. Continental and hotspot magmas are enriched in H_2O , CO_2 , P, Cl, F, Rb, Sr, Ba, Ti, etc. Ultramafic mantle xenoliths are also enriched in these components, even those which have been depleted in a basaltic component. The minerals in eclogite xenoliths are low in the incompatible elements but the intergranular material is enriched in a material having abundances similar to the continental crust. We suggest that LIL elements were removed from the MORB reservoir not only by extraction of the continental crust but also by removal of a fluid or vapor phase which has enriched the continental lithosphere and the upper mantle peridotite layer.

A schematic continental geotherm with a thermal boundary layer is shown in Figure 5. The ascent path of an eclogite diapir is also shown. Yoder (1976) has estimated that eclogite must rise about 160 km from its source region in order to completely melt if the heat of melting is obtained by adiabatic rise. For a 220 km deep source region complete melting would be achieved at 60 km. Note that completely molten eclogite, i.e., basaltic magma, can be delivered to the surface without melting dry garnet peridotite. With the geotherm shown, wet peridotite diapirs can rise from the thermal boundary layer but they will be only partially molten even after ascent to the surface because of the broad melting interval. Extensively molten peridotitic diapirs are not possible today because of their high liquidus temperature and their relatively shallow origin. This suggests that komatiites

could only form when peridotite diapirs could rise from depths greater than some 300 km, i.e., prior to the establishment of the thick eclogite layer. Only such deep diapirs could extensively melt ($>60\%$) before they reach the surface. This assumes that the geotherm approaches the adiabat only at greater depths.

The gradual heating associated with continental insulation will mobilize mantle fluids before extensive melting occurs. Therefore, a metasomatic precursor and a redistribution of LIL can be expected prior to continental magmatism. Alkali basalt activity can also be expected to precede and accompany tholeiitic eruption from the deeper levels.

We have suggested that a thermal perturbation may be required to initiate the rise of diapirs from the MORB source region. Tholeiitic volcanism is also associated with back-arc spreading. This suggests that subduction can also trigger the rise of diapirs from the MORB source. They will not have been preheated to the extent of normal MORB magmas and may be more volatile rich. The spatial relationships of marginal basin and island arc volcanics, relative to the Benioff zone, suggest that the latter come from shallower depths than the back-arc tholeiites. Since island arcs are generally less than about 150 km above the Benioff zone (Ringwood, 1975) this suggests that marginal basin basalts come from somewhat deeper, consistent with our interpretation of the location of the MORB source region.

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Table 1
POSSIBLE COMPOSITIONS OF THE TRANSITION ZONE ECLOGITE LAYER

	<u>(1)</u>	<u>(2)</u>	<u>(3)</u>	<u>(4)</u>	<u>(5)</u>	<u>(6)</u>	<u>(7)</u>	<u>(8)</u>
SiO ₂	50.3	47.8	46.2	45.7	49.5	46.6	45.5	47.2 ± 2.4
TiO ₂	1.2	0.6	0.7	0.4	0.5	0.8	1.9	0.6 ± 0.3
Al ₂ O ₃	16.5	12.1	12.6	17.9	8.5	13.7	12.4	13.9 ± 4.5
FeO	8.5	9.0	11.4	11.2	8.8	9.1	9.5	11.0 ± 3.6
MnO	0.1	0.1	0.2	0.2	0.2	0.2	0.2	0.2 ± 0.1
MgO	8.3	17.8	16.6	11.9	16.2	16.1	18.8	14.3 ± 3.0
CaO	12.3	11.2	10.5	7.4	10.6	11.8	9.7	10.1 ± 2.2
Na ₂ O	2.6	1.3	1.2	2.0	1.7	1.3	1.6	1.6 ± 1.1
K ₂ O	0.2	0.03	0.02	0.4	1.1	0.02	0.1	0.5 ± 0.4

- (1) Oceanic tholeiite (Kay et al, 1970, Engel and Engel, 1964)
- (2) Oceanic crust, calculated from ophiolite section (Elthon, 1979)
- (3) Basaltic "komatiite", Gorgona Island (Gansser et al, 1979)
- (4) } Mantle eclogites (Nixon, 1973)
- (5) }
- (6) Possible eclogite extract in fractionation of primary magma in upper mantle (O'Hara et al, 1975)
- (7) Picrite (Ringwood, 1975)
- (8) Average biminerallitic eclogite in kimberlite (Ito and Kennedy, 1974)

Table 2
ESTIMATES OF LITHOPHILE ELEMENT CONCENTRATIONS
(ppm) IN BULK EARTH, THE ECLOGITE AND PERIDOTITE
SOURCE REGIONS AND VARIOUS PRODUCTS OF THESE
SOURCE REGIONS

	K	Rb	Sr	Rb/Sr	U	Ref.
Abyssal tholeiite	732	0.75	92	0.008	0.16	(1)
Kimberlite eclogite	820	0.7	95	0.007	0.17	(2)
Kimberlite peridotite	617	3.4	55	0.061		(2)
Kimberlite peridotite	483	2.0	59	0.035		(2)
"Plume" source	>468	2.5	60	0.042		(1)
Continental flood basalts	6400	17	320	0.053	0.3	(3,4)
Ocean island basalts	3160	5.3	231	0.023		(1)
Continental crust	15 000	33	370	0.089	0.7	(3)
Intergranular material in eclogites	16 000	48	550	0.087		(2)

TABLE 2

(LEGEND)

- (1) White and Schilling (1978)
- (2) Allsop et al (1969)
- (3) Jacobsen and Wasserburg (1979)
- (4) Carmichael et al (1974)

Table 3
TRACE ELEMENTS IN INFERRED MID-OCEAN RIDGE BASALTS,
ECLOGITE XENOLITHS, PARTIAL MELT OF "FERTILE" GARNET PERIDOTITE
AND CONTINENTAL FLOOD BASALTS

	<u>Oceanic</u>			<u>Continental</u>		
	(1)	(2)	(3)	(4)	(5)	(6)
K	820	732	700	2000-3600	4000	6400
Rb	0.7	0.75	1.1	11-20	20	17
Sr	95	92	134	30-500	400	320
K/Rb	1170	976	640	100-300	400	376
Rb/Sr	0.01	0.008	0.01	0.04-0.3	0.005	0.053

- (1) "Ideal" eclogite xenolith (Allsop et al, 1969). Potential source region for oceanic tholeiites.
- (2) Oceanic tholeiite (White and Schilling, 1978)
- (3) Oceanic tholeiite (Table 1)
- (4) Inferred partial melt (20%) product from fertile garnet peridotite xenoliths with sterile peridotite xenoliths as residual (Rhodes and Dawson, 1975)
- (5) Karoo basalts (Carmichael et al, 1974)
- (6) Continental flood basalts (Jacobsen and Wasserburg, 1979)

Table 4

INCLUSIONS IN KIMBERLITES AND ESTIMATES
OF PRIMITIVE MANTLE COMPOSITION

	Xenoliths		Mantle Compositions				
	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO ₂	47.2	47.3	47.3	47.3	48.0	46.6	44.8
TiO ₂	0.6	0.05	0.2	0.2	0.3	0.3	0.2
Al ₂ O ₃	13.9	1.6	4.1	5.3	5.2	3.0	5.3
FeO	11.0	5.8	6.8	7.4	7.9	10.4	10.3
MgO	14.3	43.8	37.9	35.0	34.3	34.2	34.3
CaO	10.1	1.0	2.8	3.7	4.2	4.8	4.4
Na ₂ O	1.6	0.2	0.5	0.6	0.3	0.2	0.4
K ₂ O	0.5	0.2	0.2	0.3	0.2	0.03	0.03

(1) Average eclogite nodules (Ito and Kennedy, 1974)

(2) Average garnet lherzolite in kimberlite (O'Hara et al, 1975)

(3) 20% eclogite, 80% garnet lherzolite.

(4) 30% eclogite, 70% garnet lherzolite.

(5) Primitive mantle (Ganapathy and Anders, 1974)

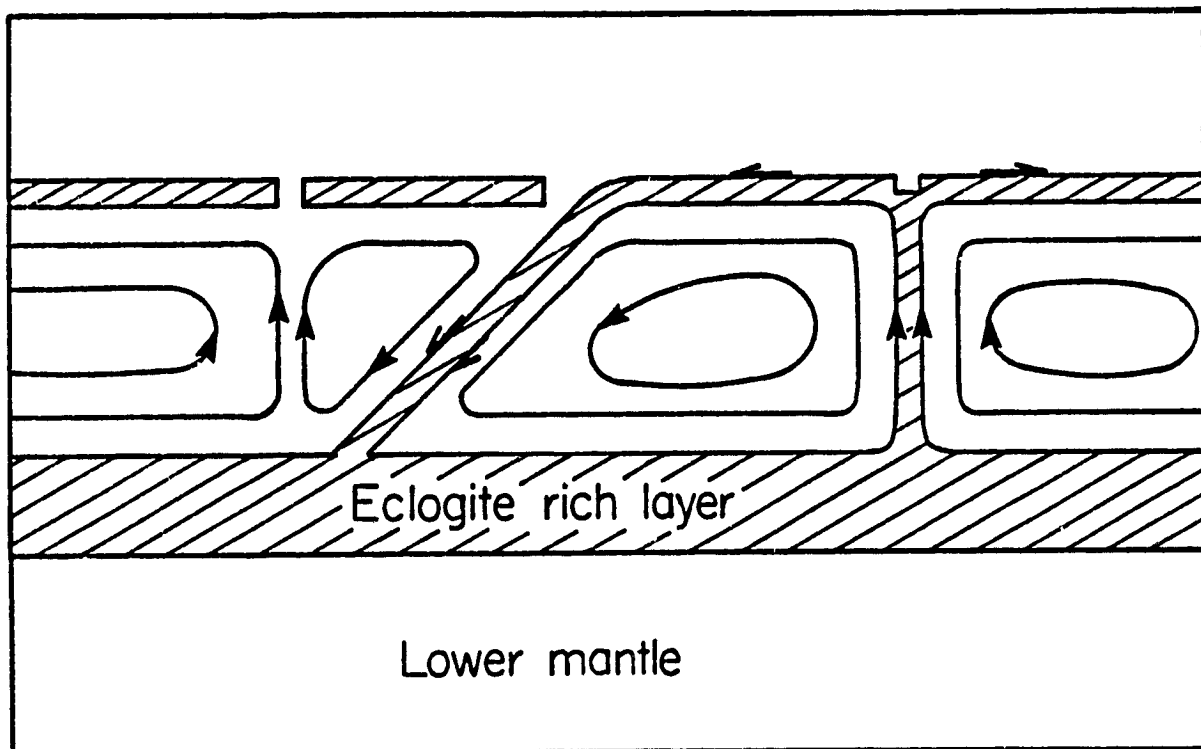
(6) Average peridotitic komatiite, S. Africa (Viljoen and Viljoen, 1969)

(7) Peridotite with quench texture, W. Australia (Nesbitt, 1972)

FIGURE CAPTIONS

- Figure 1. Schematic of the primary plate tectonic cycle with the transition zone eclogite layer serving as the source and sink of the oceanic lithosphere. The harzburgite portion of the lithosphere remains in the upper mantle.
- Figure 2. Flow chart of mantle differentiation. The primitive differentiation results in basalt and peridotite. Partial melting of the basalt layer concentrates the LIL elements into the continental crust.
- Figure 3. Schematic illustration of isotherms for convection in a stratified system showing boundary layer detachment. The locations of descending plumes in the lower layer will be controlled by the locations of the cold isotherms in the upper layer. Material penetrating into the lower layer may initiate diapiric uprise from this region.
- Figure 4. Illustration of flow in superposed convecting layers. The lower layer is internally heated and is characterized by broad upwellings. The upper layer is mainly heated from below and is characterized by narrow ascending plumes.

Figure 5. Geotherm in a chemically stratified mantle showing the thermal boundary layer and the path of ascent of an eclogite diapir that starts to melt at about 250 km. Temperatures at the top of the eclogite layer must appreciably exceed the solidus before the diapir has sufficient buoyancy to rise through the peridotite layer. It may heat further if its upward escape is prevented by an overlying continental lithosphere. Extrusion temperatures and depth of complete melting may therefore be greater than shown here. Dry melting curves are from Wyllie (1971) and Yoder (1976).

*Fig. 1*

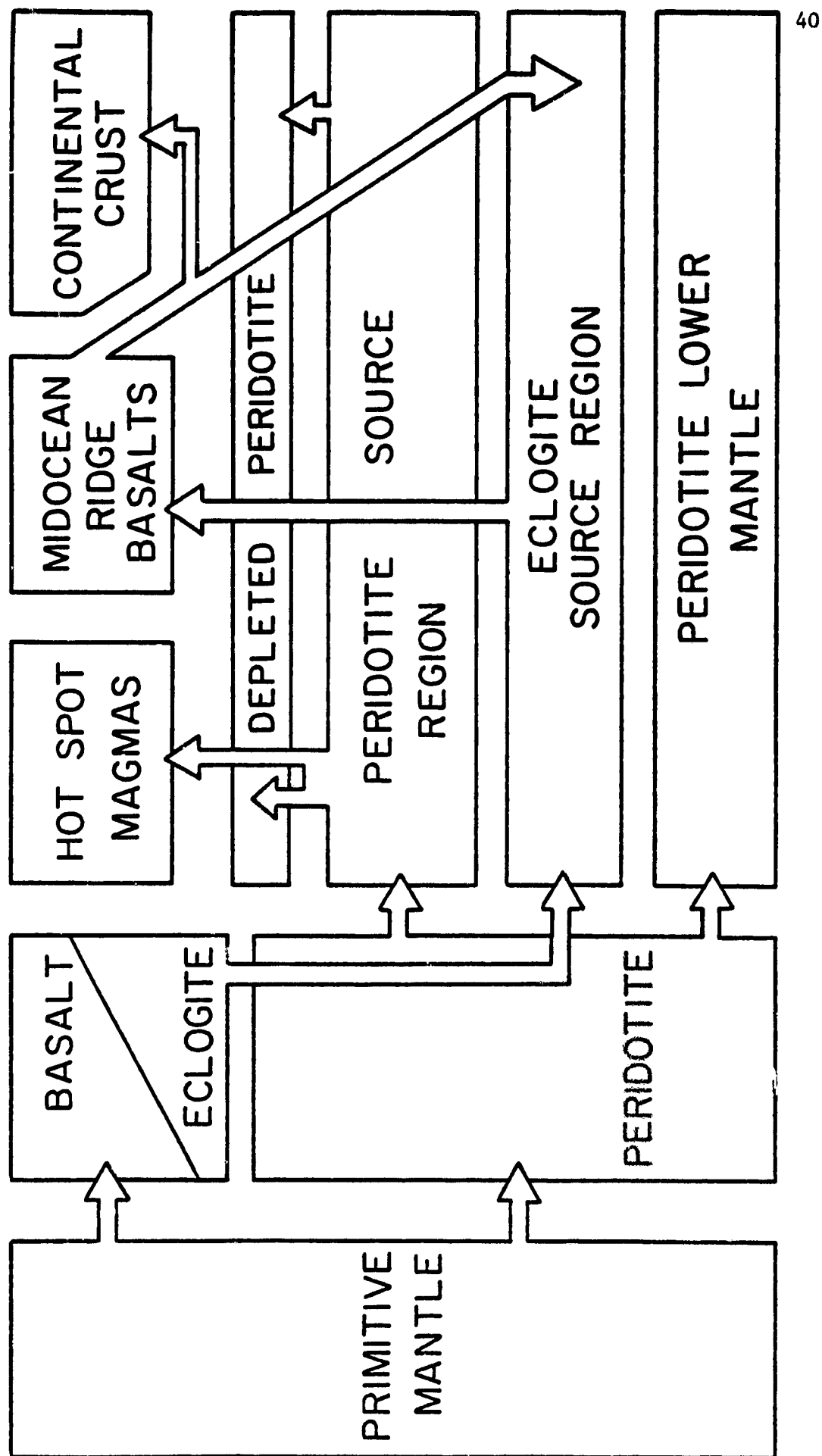
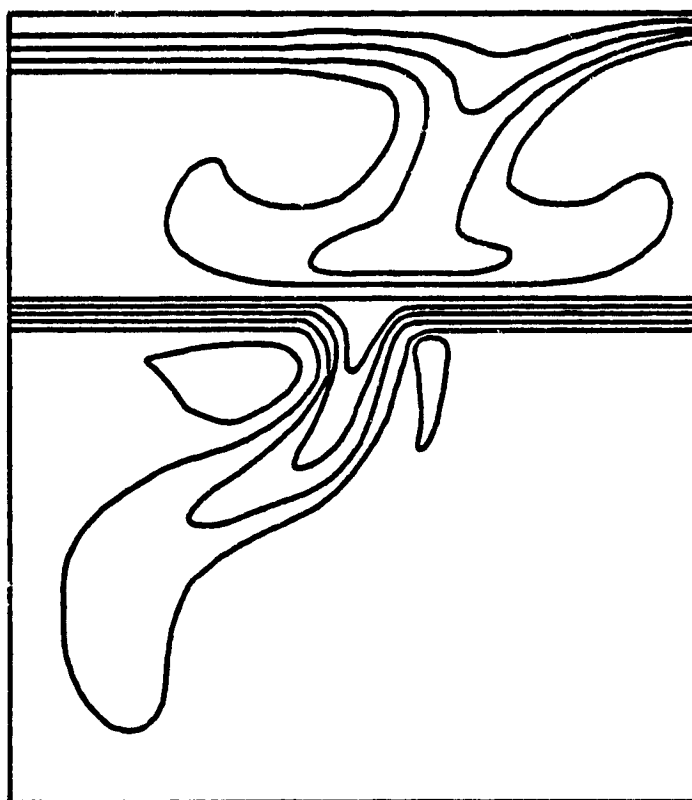


Fig. 2

*Fig. 3*

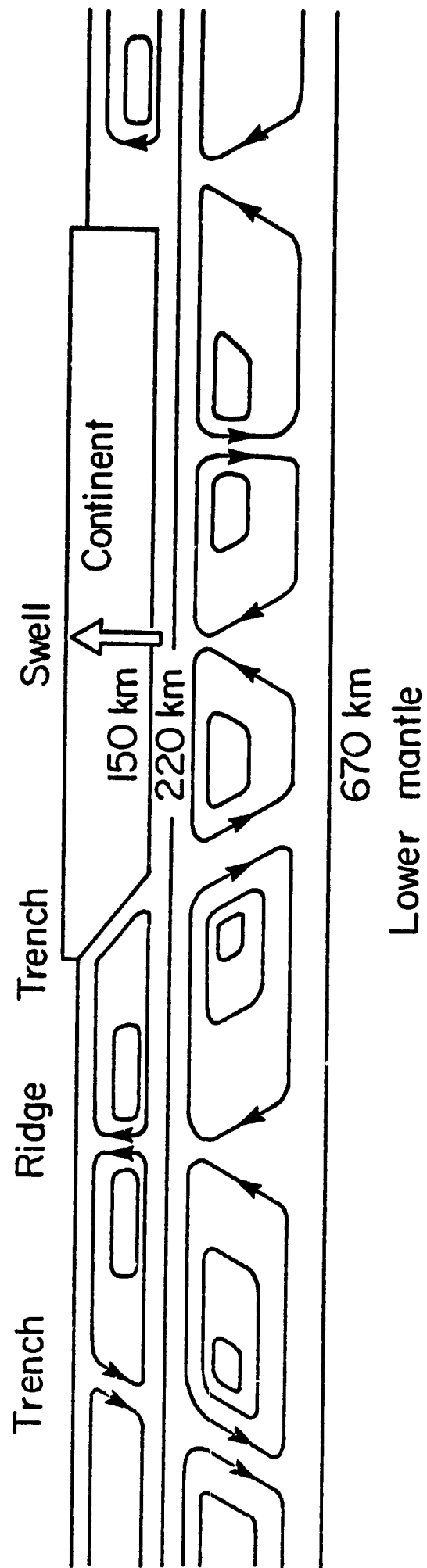
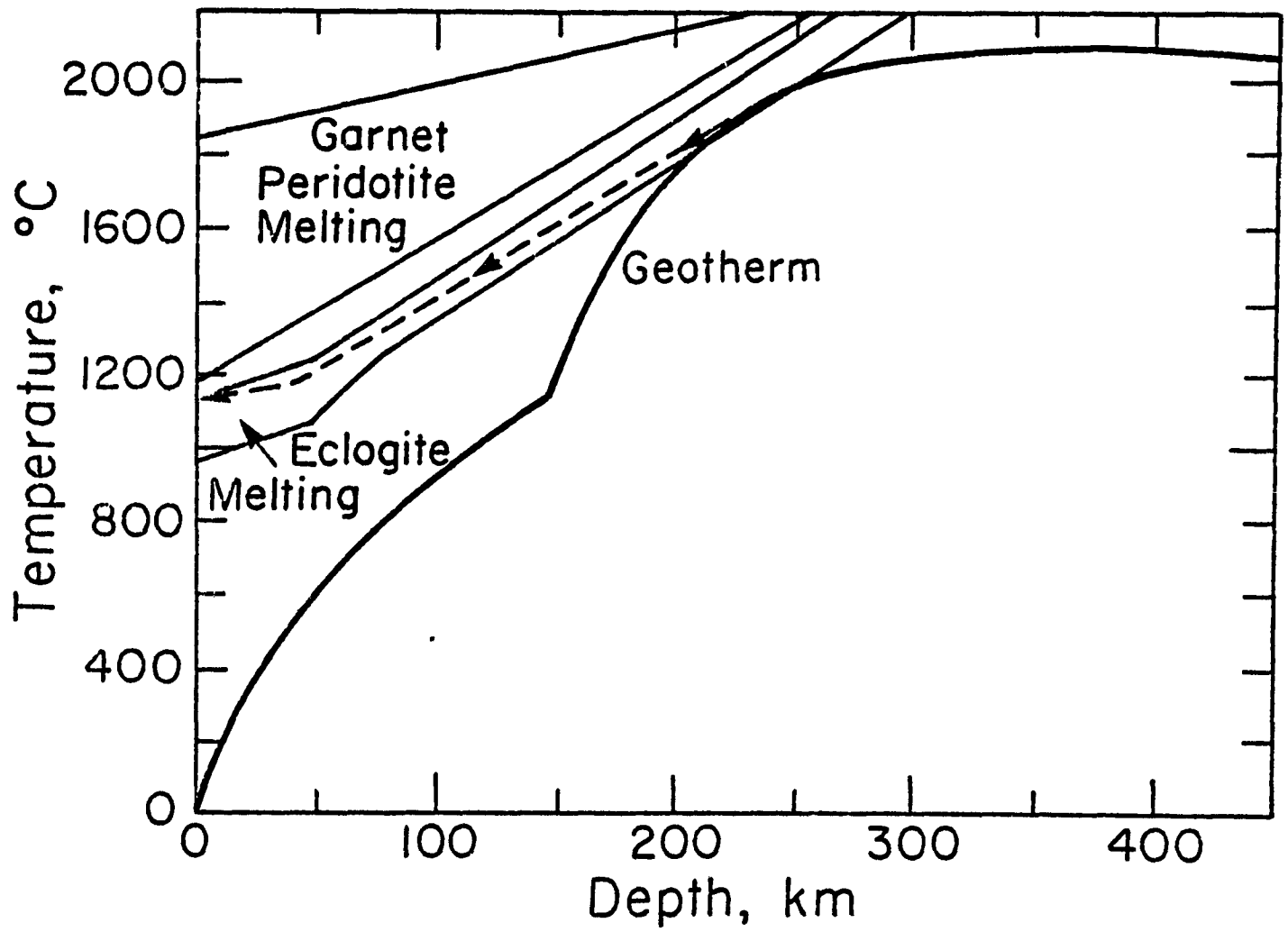


Fig. 4

*Fig. 5*