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The depths of mantle reservoirs

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Abstract—Many petrological studies are concerned with the temperature and pressure of final equilibrium of erupted magmas and residual crystals. The average composition of the source region and its original depth are also of interest but these cannot be determined unambiguously from petrology. Seismic techniques can be used to infer the mineralogy of various regions of the mantle and the probable depth extent of the low–velocity zones (LVZ) associated with high–temperature buoyant upwellings.

Oceanic ridges are characterized by broad deep LVZ's which extend locally to depths in excess of 400 km. Partial melting is implied to depths of at least 300 km. The same is true for some young continental regions such as northeast Africa and western North America, and some midplate regions such as the central Pacific. Hotspots occur on the edges of broad upper mantle low–velocity anomalies, often in regions of thick crust and/or old and thick lithosphere.

Continental shields have thick (150 km), cold, refractory lithospheres which are unlikely sources of voluminous plateau basalt outpourings. The rapid decrease in velocity between 150 and 200 km beneath shields implies a high thermal gradient and a change in mineralogy. From 200 to 400 km the seismic velocities beneath shields fall on the 1400°C adiabat. This suggests that the stable continental plate is 150 km thick and that it is underlain by a thermal boundary layer which grades downward into a convective gradient. Continental and oceanic basalts probably share a common source region which is deeper than 350 km. When hot, this source region becomes buoyant, because of thermal expansion and the reduction or elimination of dense phases, such as garnet, rises into the shallow mantle, adiabatically decompresses, becomes an LVZ and a potential source of magma.

INTRODUCTION

THE DEPTHS and compositions of the midocean basalt (MORB), ocean island basalt (OIB) and continental flood basalt (CFB) reservoirs cannot be inferred unambiguously from petrological and geochemical studies. Some of the minerals which were in equilibrium with observed basalts can be inferred from these studies and therefore minimum depths can be placed on the final equilibration between magmas and residual crystals. The proportions of the various minerals, however, cannot be determined. The source region itself may be much deeper or may itself have risen from greater depths before the melts were expelled.

Seismic studies can place some constraints on these petrogenesis processes. First, seismology can be used as a three–dimensional mapping tool. For example, it can be used to map shallow magma chambers and map the depth extent of the low–velocity region under ridges and other magmatic centers. It can also provide information about the depth extent of continental lithosphere and subduction zones.

Secondly, seismology provides inferential information about the mineralogy and physical state of the various regions of the mantle. In this paper we address both issues: the structure of the upper mantle: and its mineralogy and physical state.

THREE–DIMENSIONAL STRUCTURE OF THE UPPER MANTLE

The low–velocity zone (LVZ) has played a prominent role in most discussions of the location of the basalt source region. In most global seismic models the low–velocity zone occupies the depth interval between about 50 and 200 km. The most plausible explanation of the LVZ involves a partial melt content (ANDERSON and SAMMIS, 1970; ANDERSON and SPETZLER, 1970). In the first global surface wave inversions, it was found that the depth and nature of the LVZ varies from one tectonic province to another (ANDERSON, 1967; TOKSÖZ and ANDERSON, 1966). In shield areas the LVZ is deeper than 120 km and is less pronounced than in oceanic regions. Active tectonic regions also have shallow and pronounced low–velocity zones. These early results are shown in Figure 1.

More recently, high–resolution body wave studies have provided details about upper mantle velocity structures in several tectonic regions. Figure 2 shows some of these results for the Canadian shield (stable continent), western North America–East Pacific

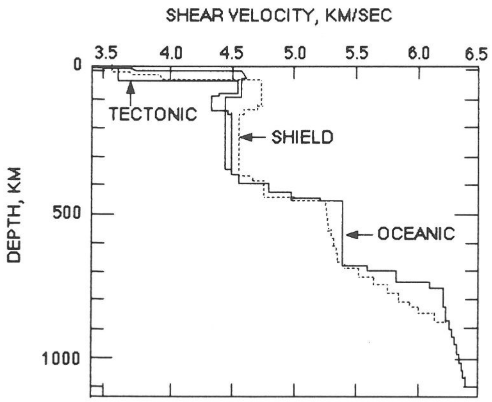

FIG. 1. Shear velocities in upper mantle for different tectonic provinces derived from early surface wave studies (ANDERSON, 1967; TOKSÖZ and ANDERSON, 1966). Note the thick high-velocity shield LID and the shallow low-velocity zones under tectonic and oceanic regions.

Rise (tectonic–young ocean) and the western Atlantic (old ocean). Note that low-velocities extend to depths of about 390 km for the tectonic and oceanic structures. These regional studies confirm the general features of the earlier global studies (ANDERSON, 1967; TOKSÖZ and ANDERSON, 1966).

Shields have extremely high shear velocities extending to 150 km depth. It is natural to assume that this is the thickness of the stable continental plate and that the underlying mantle is free to deform and convect. The high-velocity layer, or LID, under tectonic and oceanic regions is much thinner, of the order of 30 to 50 km, and the shear velocities of the underlying mantle are much lower than under shields, implying higher temperatures and, possibly, the presence of a partial melt phase. The implication is that oceanic plates are much thinner and, possibly more mobile than continental plates.

JORDAN (1975) and SIPKIN and JORDAN (1976) made a radically different proposal. They suggested that the high seismic velocity associated with shields extended to depths in excess of 400 km and perhaps to 700 km and that the continental plates are equally thick. This hypothetical deep continental root was called the "tectosphere" (JORDAN, 1975). OKAL and ANDERSON (1975) and ANDERSON (1979) showed that the large differences in oceanic and continental ScS times (shear waves which reflect off the core), the data used in the development of the continental tectonosphere hypothesis, were mainly caused by differences shallower than 200 km. These waves have very little depth resolution and can only resolve differences below 400 km if the shallower mantle is independently constrained.

Although the largest variations (on the order of 10%) in seismic velocity occur in the upper 200 km of the mantle, the velocities from 200 to about 400 km under oceanic and tectonic regions are slightly less (on the order of 4% on average) than under shields. The question then arises, what is the cause of these deeper velocity variations? Is the continental plate 400 km thick or are the velocities between 150–200 and 400 km beneath shields appropriate for "normal" subsolidus convecting mantle?

ANDERSON and BASS (1984) attempted to answer this question by computing the seismic velocities for several plausible mineral assemblages along a variety of adiabats. The results are shown in Figure 3 along with several recent high-resolution seismic models for different tectonic regions. The heavy lines give the calculated velocities as a function of depth for adiabatic temperature gradients which start at surface temperatures ranging from 600° to 1800°C. The temperature gradient in convecting regions of the mantle is expected to be close to adiabatic. Near the surface and near chemical interfaces, *i.e.*, in thermal boundary layers, heat is transferred by conduction and much higher temperature gradients can be maintained than elsewhere. In such regions the increase of velocity with depth is much

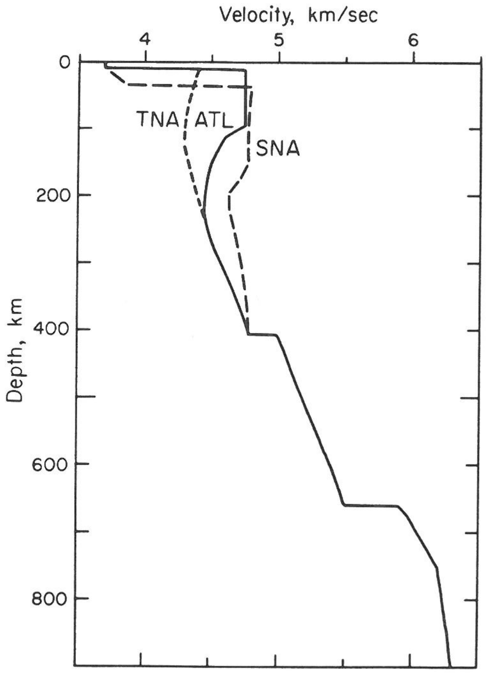

FIG. 2. Recent shear velocity profiles derived from high-resolution body wave studies for the Canadian Shield (SNA), western North America–East Pacific Rise (TNA) and the northeastern Atlantic (ATL). (from GRAND and HELMBERGER, 1984a,b).

less than along an adiabat. In Figure 3 the upper 150 km under shields and the upper 100 km under the North Atlantic (old ocean) have low velocity gradients, implying a rapid increase of temperature or a change in composition with depth.

The intersections of the adiabats with the dry solidus of peridotite are shown with dash–dot curves labelled "solidus" (Figure 3). Below these curves the calculated velocities represent upper bounds because the effect of the melt phase has not been taken into account. Measured seismic velocities which plot below the anhydrous solidus are presumably affected by the presence of melt and can be used only to infer upper bounds on the temperature.

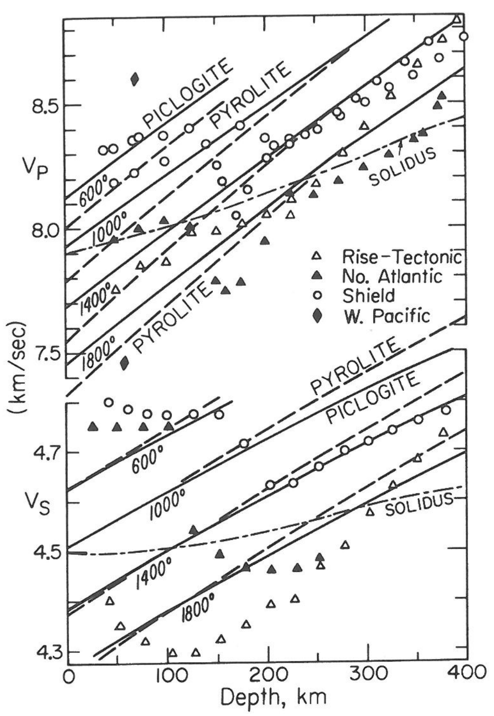

FIG. 3. Theoretical compressional velocity (V_p) and shear velocity (V_s) profiles along various adiabats for two mineral assemblages, pyrolite (mainly olivine and orthopyroxene) and piclogite (an olivine eclogite). A melt phase is probably present on the low–velocity side of the anhydrous solidus curves and this would lower the calculated velocities. The symbols are seismic results for various depths and tectonic provinces. The data are mostly from GRAND and HELMBERGER (1984b), WALCK (1984), L. ASTIZ and V. LEFEVRE (personal communications). Data which fall below the solidus curves are probably in the partial melt field. Note that the shield data fall near the 1400°C adiabat below 200 km depth. The low temperatures inferred for the shield lithosphere can be raised if the lithosphere is richer in olivine, particular forsterite–rich olivine (modified from ANDERSON and BASS, 1984).

In the upper 150 km of the shield mantle the derived temperature is low, 400–1000°C, and the inferred temperature gradient is higher than adiabatic, as appropriate for a conductive thermal boundary layer. Consistent temperatures can be obtained for both V_p and V_s if the shield lithosphere is more olivine–rich and more forsteritic than pyrolite (ANDERSON and BASS, 1984). The shield lithosphere therefore appears to be cold depleted peridotite or harzburgite. The rapid decrease of velocity between 150 and 200 km implies a high thermal gradient probably accompanied by a change in mineralogy toward a more fertile peridotite. By a change in mineralogy we mean a change in the relative proportions of the major mantle minerals; olivine, orthopyroxene, clinopyroxene and garnet. A garnet–poor depleted shield lithosphere is buoyant relative to fertile peridotite or pyrolite (BOYD and MCCALLISTER, 1976) unless offset by the colder temperatures. The presence of such a cold, thick, compositionally distinct layer extends the depth of the surface thermal boundary layer. Below 200 km the sub–shield velocities follow the 1400°C adiabat. This observation implies convection below this depth, and this is consistent with the flow behavior of olivine–rich rocks (*e.g.*, KOHLSTEDT and GOETZE, 1974). Olivine flows easily at these temperatures and at stress levels thought to be appropriate for the upper mantle. Even though the seismic velocities under shields are higher than average mantle velocities to depths approaching 400 km, there is no reason to suppose that this material is not participating in mantle convection. As mentioned earlier, conductive layers are characterized by high thermal gradients. Stagnant regions of the mantle, regions where convection is not operational, will therefore have low or negative seismic gradients unless offset by a chemical or mineralogical gradient. The fact that seismic velocities in the sub–shield mantle fall along an adiabat either implies a convective gradient or a fortuitous combination of thermal and chemical gradients. In the latter case temperatures would rise more rapidly than in the former. The implied nature (temperature and composition) of the 150 km–thick shield LID make this an unlikely source for continental flood basalts or rift volcanics. Evidence from kimberlite inclusions suggests that the upper 150 km of the shield mantle has been stable for some time (RICHARDSON *et al.*, 1984; BOYD *et al.*, 1985). A refractory, chemically distinct shield lithosphere underlain by a thermal boundary layer is also consistent with the chemistry of mantle nodules and the "kinked geotherm" derived therefrom (*e.g.*, BOYD, 1979). The deeper nodules are sheared fertile peridotites, consistent with flow below 150 km, and a change in chemistry near this depth. The shield sub-lithospheric temperature inferred from Figure 3 is consistent with the temperature between 150–200 km estimated from present-day heat flow (POLLACK and CHAMPMAN, 1977) and at 3 b.y. from thermobarometry (BOYD *et al.,* 1985).

Seismic velocities which are lower than those inferred for any plausible mineral assemblage provide evidence for anelastic phenomena such as partial melting, grain boundary relaxation or dislocation relaxation (ANDERSON and SAMMIS, 1970; MINSTER and ANDERSON, 1980, 1981). These are all high temperature phenomena but it is essentially impossible to distinguish between partial melting and the subsolidus mechanisms. On the other hand there is abundant evidence from other considerations that a large part of the upper mantle is near or above the melting point, particularly if volatiles are present, so we have cast our discussion in terms of the partial melt mechanism. Seismic velocities which fall well below the calculated curves are almost surely affected by anelastic processes, but even the higher velocities may also be affected to some extent, by processes such as partial melting. We have placed conservative estimates on the depth of inferred partial melting.

There are three obvious mechanisms for providing hot material to the base of the stable continental lithosphere, a process required to initiate volcanism. (1) The thermal boundary layer below 150 km can become unstable, allowing hot material to impinge the base of the lithosphere. (2) A diapir from greater depth can be brought into the shallow mantle. Such diapirs may initiate in a thermal boundary layer at 400 or 650 km, if these are chemical boundaries, or at a depth where the geotherm crosses the solidus. And finally, (3) the continent may drift over a hotter region of the mantle.

It is not clear whether continental rifting is a result of these deep processes or if rifting initiates the instability at depth. Continental convergence and collision may also trigger instability of the thermal boundary layer. Large and deep (>300 km) seismic anomalies are associated with the Red Sea–East African Rift and western North America as well as with midoceanic ridges and back–arc basins (See later sections.) This observation might suggest that none of these areas are entirely a result of passive rifting and pressure–release melting, although these effects can contribute to further melting and intrusion.

The geochemical nature of continental flood basalts, rift basalts, ocean island basalts and kimberlites suggests that they have either originated in, or at least have passed through or evolved in, a mantle that is enriched in the more incompatible elements. The shallow mantle may be the source of this enriched or hotspot signature (or the "metasomatic" fluids), even if the bulk of the magma comes from greater depth. Rising diapirs from a deep, depleted (low Rb, Sr, U, Th, LREE, Rb/Sr, Nd/Sm etc.) yet fertile (high CaO, Al_2O_3, Na_2O) reservoir will tend to expel their melts at shallow depths which, in turn, pond beneath the cold continental lithosphere where they cool, fractionate and ecome contaminated with shallow mantle and lower lithosphere melts prior to eruption. A similar mechanism may operate under old oceanic lithosphere, such as Hawaii, and thick crustal regions such as Iceland. Cooling, fractionation and contamination will also result if the parent magma must flow laterally for large distances before it can escape to the surface. On the other hand, basalts from the uplifted depleted reservoir which can flow directly to the surface through thin crust and lithosphere, as at mid-ocean ridges, will be relatively uncontaminated and will have experienced less cooling and fractionation and the fractionation (*e.g.,* olivine, plagioclase) may be mainly shallow. It is necessary to distinguish between the trace element characteristics and the major element chemistry of basalts because MORBs are trace element depleted and fractionated, relative to chondrites, but obviously come from a fertile source whereas OIB and CFB are trace element enriched (sometimes called undepleted) and also come from a fertile source. The use of "depleted", "enriched", "fertile" etc. to refer to both the major element and minor element or isotopic chemistry can be confusing. A fertile, yet depleted, source could be for example, a garnet–clinopyroxene–rich cumulate or a garnet–clinopyroxene–rich layer that has had its incompatible elements stopped out in a prior stage of partial melting (ANDERSON, 1982, 1985).

THE OCEANIC MANTLE

Figure 3 shows that the seismic velocities in most of the upper mantle, down to at least 300 km, are so low that partial melting is implied for the tectonic and oceanic mantles. The high negative velocity gradients in the shallow mantle are probably the combined effects of a high temperature gradient and an increasing melt fraction with depth. At greater depth the high positive gradients are consistent with a decreasing melt content with depth. It appears that geographically large areas of the upper mantle are partially molten. Melting is not restricted to narrow, shallow zones associated with midoceanic and continental rifts although the pressure release associated with such near surface phenomena, and the resulting adiabatic ascent, may increase the amount of partial melting. We show later that large volumes of the shallow mantle down to 200 or 300 km in the vicinity of back–arc basins are also very slow. Thus, much of the upper mantle above 300 km is close to or above the melting point. Even the 1400°C adiabat characterizing the shield mantle crosses the dry solidus near 100 km. In the absence of a cold shield lithosphere, material advecting up the 1400°C adiabat would melt at ~100 km depth and even deeper if the wet solidus is more appropriate. The implied temperatures between 150 and 200 km are close to the wet solidus of peridotite.

The suboceanic geotherms may be on higher temperature adiabats. The velocities will not follow the calculated adiabats if the melt content varies with depth. The effect of melting on the seismic velocities is not included in the calculations of Figure 3.

In the next section we show that higher than average velocities between 200 and 400 km are not confined to stable shield areas. Old oceans, on average, many convergent regions and some tectonic regions also have fast velocities in this depth interval.

GLOBAL SURFACE WAVE TOMOGRAPHY

A global view of the lateral variation of seismic velocities in the mantle can now be obtained with surface wave tomography (NAKANISHI and ANDERSON, 1983, 1984a,b; TANIMOTO and ANDERSON 1984, 1985; NATAF *et al.* 1984, 1986; ANDERSON and DZIEWONSKI, 1984; DZIEWONSKI and ANDERSON, 1984; WOODHOUSE and DZIEWONSKI, 1984; TANIMOTO, 1984, 1985). There are two basic approaches. One is the regionalization approach (ANDERSON, 1967; TOKSÖZ and ANDERSON, 1966) which assumes that the velocities of surface waves are linearly dependent on the fracton of time spent in various tectonic provinces. The inverse problem then states that the velocity profile depends only on the tectonic classification. For example, all shields are assumed to be identical at any given depth. This assumption appears to be valid for the shallow structure of the mantle (NATAF *et al.*, 1984, 1986; WOODHOUSE and DZIEWONSKI, 1984) but becomes increasingly tenuous for depths greater than 200 km. However, it probably provides a maximum estimate of the depth of tectonic features and it also provides a useful standard model with which other kinds of results can be compared.

The second approach subdivides the Earth into cells or blocks or by some smooth function such as spherical harmonics. NATAF *et al.* (1984, 1986) used spherical harmonics for the lateral variation and a series of smooth functions joined at mantle discontinuities for the radial variation. In this approach no *a priori* tectonic information is built in.

In both of these approaches the number of parameters that one would like to estimate far exceeds the information content (*i.e.*, number of independent data points) of the data. It is therefore necessary to decide which parameters are best resolved by the data, what is the resolution, or averaging length, which parameters to hold constant and how the model should be parameterized (*e.g.*, layers or smooth functions, isotropic or anisotropic). In addition, there is a variety of corrections that might be made (*e.g.*, crustal thickness, water depth, elevation, ellipticity, attenuation). The resulting models are as dependent on these assumptions and corrections as they are on the quality and quantity of the data. This is not unusual in science. Data must always be interpreted in a framework of assumptions and the data are always, to some extent, incomplete and inaccurate. In the seismological problem the relationship between the solution, or the model, including uncertainties, and the data can be expressed formally. The effects of the assumptions and parameterizations, however, are more obscure but these also influence the solution. The hidden assumptions are the most dangerous. For example, most seismic modelling assumes perfect elasticity, isotropy, geometric optics and linearity. To some extent, all of these assumptions are wrong, and their likely effects must be kept in mind. Petrologists, of course, also design and interpret their experiments in terms of an array of prejudices ("the paradiam") about what the source region is like and what processes may or may not be important. The data itself may be consistent with quite a different scenario, *e.g.*, the actual composition and evolution of the Earth.

NATAF *et al.* (1986) made an attempt to evaluate the resolving power of their global surface wave dataset and invoked physical *a priori* constraints in order to reduce the number of independent parameters which needed to be estimated from the data. For example, the density, compressional velocity and shear velocity are independent parameters but their variation with temperature, pressure and composition show a high degree of correlation, *i.e.*, they are coupled parameters. Similarly, the fact that temperature variations in the mantle are not abrupt means that lateral and radial variations of physical properties will generally be smooth except in the vicinity of phase boundaries, including partial melting. Changes in the orientation of crystals in the mantle will lead to changes in both the shear wave and compressional velocity anisotropies. These kinds of physical considerations can be used in lieu of the standard seismological assumptions which are generally made for mathematical convenience rather than physical plausibility.

The studies of WOODHOUSE and DZIEWONSKI (1984) and NATAF *et al.*, (1984, 1986) give upper mantle models which are based on quite different assumptions and data analysis techniques. WOODHOUSE and DZIEWONSKI (1984) inverted for shear velocity, keeping the density, compressional velocity and anisotropy fixed. They also use a very smooth radial perturbation function that ignores the presence of mantle discontinuities and tends to smear out anomalies in the vertical direction. They corrected for near-surface effects by assuming a bimodal crustal thickness, continental and oceanic.

NATAF *et al.* (1986) corrected for elevation, water depth, shallow mantle velocities and measured or inferred crustal thickness. They inverted for shear velocity and anisotropy but included physically plausible accompanying changes in density, compressional velocity and anisotropy. Corrections were also made for anelasticity. The radial perturbation functions were allowed to change rapidly across mantle discontinuities, if required by the data. In spite of these differences, the resulting models of NATAF *et al.* (1986) and WOODHOUSE and DZIEWONSKI (1984) are remarkably similar above about 300 km. The main differences occur below 400 km. These differences seem to arise from differences in the assumptions and parameterizations (crustal corrections, radial smoothing functions) rather than the data. The choice of an *a priori* radial perturbation function can degrade the vertical resolution intrinsic to the dataset. The solution, in this case, is overdamped or oversmoothed.

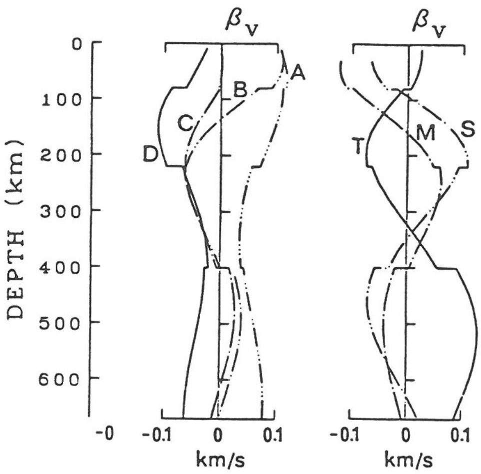

FIG. 4. Shear velocity (βv) vs. depth, expressed as differences from average Earth model PREM (DZIEWONSKI and ANDERSON, 1981). The oceanic profiles are in order of increasing age (A is oldest ocean, D is youngest). The other profiles are shields (S), mountain–tectonic (M) and trench–marginal seas (T) (after NATAF *et al.*, 1986).

REGIONALIZED RESULTS

Figure 4 shows vertical shear velocity profiles, expressed as differences from the average Earth, using the regionalization approach. Young oceans, Region D, have slower than average velocities throughout the upper mantle and are particularly slow between 80 and 200 km, in agreement with the higher resolution body wave studies. Old oceans, Region A, are fast throughout the upper mantle. Intermediate age oceans are intermediate in velocity at all depths. Most of the oldest oceans are adjacent to subduction zones and the subduction of cold material may be partially responsible for the fast velocities at depth. Notice that velocities converge toward 400 km but differences still remain below this depth (Figure 4). The continuity of the low velocities beneath young oceans, which include mid-ocean ridges, suggests that the ultimate source region for MORB is below 400 km. Shields are faster than average and faster than all other tectonic provinces except old ocean from 100 to 250 km. Below 220 km the velocities under shields decrease, relative to average Earth, and below 400 km shields are among the slowest regions. At all depths beneath shields the velocities can be accounted for by reasonable mineralogies and temperatures without any need to invoke partial melting. Trench and marginal sea regions, on the other hand, are relatively slow above 200 km, probably indicating the presence of a partial melt, and fast below 400 km, probably indicating the presence of cold subducted lithosphere. The large size of the tectonic regions and the long wavelengths of surface waves require that the anomalous regions at depth are much broader than the sizes of slabs or the active volcanic regions at the surface. This suggests very broad upwellings under young oceans and abundant piling up of slabs under trench and old ocean regions.

Maps of the regionalized results are shown at 250 and 350 km in Figure 5. Shields and young oceans are still evident at 250 km. At 350 km the velocity variations are much suppressed. Below 400 km, most of the correlation with surface tectonics has disappeared, in spite of the regionalization, because shields and young oceans are both slow, and trench and old ocean regions are both fast. Most of the

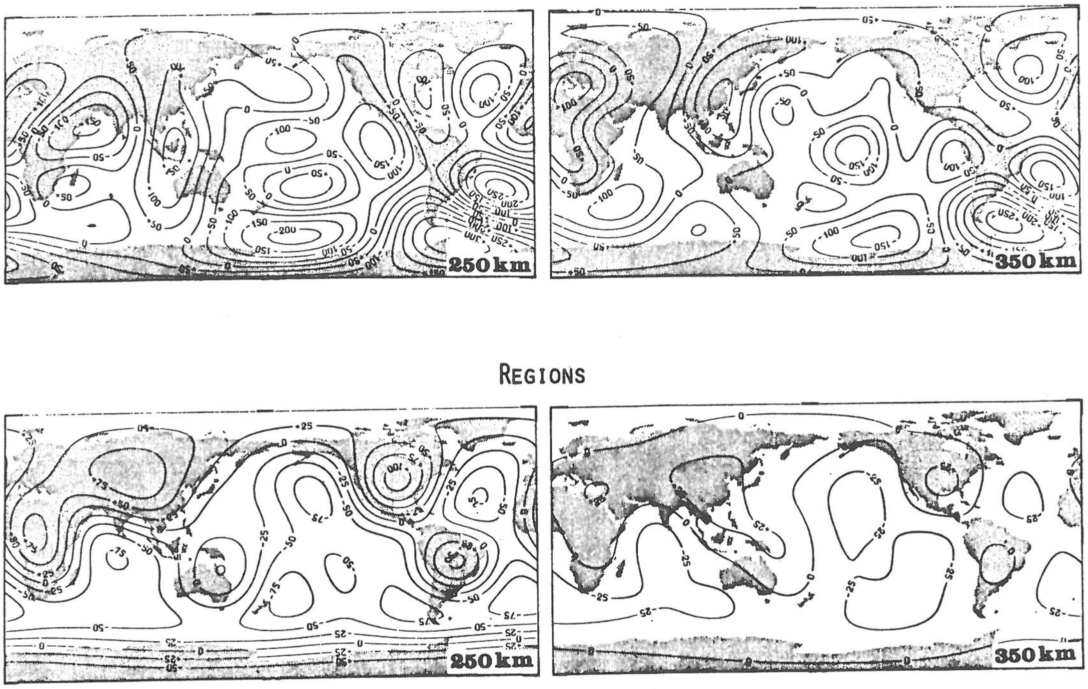

FIG. 5. Shear wave velocities (β_v, m/s) at depths of 250 and 350 km for spherical harmonic model (upper panels) and regionalized model (lower panels). The spherical harmonic model contains no *a priori* tectonic information but shields and platforms, in general, fall in high velocity regions and young oceans fall in low–velocity regions at 250 km and shallower depths. At 350 km the lowest velocity areas are under young oceans and in the central Pacific. Most hotspots lie on the periphery of these low–velocity regions. In the regionalized model all shield areas are assumed to be the same, all trench and marginal sea areas are assumed to be the same and oceanic profiles are assumed to be a function of age alone. The shields and stable platforms are still evident at 250 km in this parameterization but there is little lateral variation at 350 km (after NATAF *et al.*, 1986).

oceanic regions have similar velocities at depth. This is a severe test of the continental tectosphere hypothesis. Shields do not have higher velocities than some other tectonic regions below 250 km and definitely do not have "roots" extending throughout the upper mantle or even below 400 km. Results for other depths are given in NATAF *et al.* (1986). The maps shown (Figure 5) are spherical harmonic expansions of the regionalized results so that they can be compared with those in the following section. The minimum resolvable features with presently available data have a half–wavelength of about 2500 km. Only the long wavelength components of smaller features show up. In the high resolution studies discussed earlier we saw that subshield velocities dropped rapidly at 150 km depth although velocities remained relatively high to about 390 km. These high velocities could represent "roots" physically attached to the shield lithosphere, overridden cold oceanic lithosphere, or, simply, "normal" convecton mantle weakly coupled to the overlying shield lithosphere via a boundary layer at 150–200 km depth. We argued that the velocities below 200 km were on an adiabat and therefore probably represented normal convecting mantle. Therefore, it is the slow mantle under ridges and tectonic regions that is anomalous and, if anything, these are the regions with the roots. If the mantle under shields is convectively stagnant, as implied by the deep tectosphere hypothesis, a high thermal gradient would maintain over a large depth interval. This could lead to partial melting and a depression of the olivine–spinel phase boundary under shields. We therefore prefer the 150 km thick plate hypothesis, *i.e.*, a correspondence of the thickness of the plate with the seismic high–velocity layer.

SPHERICAL HARMONIC RESULTS

An alternate way to analyse the surface wave data is through a spherical harmonic expansion which ignores the surface tectonics. This provides a less biased way to access the depth extent of tectonic features. The results using this technique are shown in Figure 5. At 150 km (not shown) all the major shield areas fall near the centers of high–velocity anomalies and the ridges are in low–velocity regions. The main differences at this depth, compared to the regionalized models, are the very low velocities in eastern Asia, the Red Sea region, and New Zealand. At 250 km, the shields are less evident than at shallower depths and also less evident than in the regionalized model. On the other hand, the areas containing ridges are more pronounced low–velocity regions. The central Pacific and the Red Sea are also very slow. The highest velocity anomalies are in the far south Atlantic, northwest Africa to southern Europe and the eastern Indian Ocean to southeast Asia and are not confined to the older continental areas. At 350 km, most ridges are still evident although the slow region of the eastern Pacific has shifted off the surface expression of the East Pacific Rise. A central Pacific slow-velocity region persists throughout the upper mantle. Below 400 km there is little correlation with surface tectonics and in many areas the velocity anomalies are of opposite sign from those in the shallow mantle, in agreement with the regionalized results. The net result is that shields, on average, have very high velocities to 150–200 km and ridges, on average, have low velocities to 350–400 km. The Red Sea anomaly appears to extend to 400 km but the very slow velocities associated with western North America die out by 300 km.

CONCLUSIONS AND DISCUSSION

The high temperatures associated with midocean ridges appear to extend to depths of the order of 400 km. The low temperatures, or compositional differences, associated with stable continents attenuate rapidly below 200 km. The effects of temperature and composition on density must nearly compensate each other because shields are not evident in the geoid. Many shields, however, tend to fall in or near geoid lows but deglaciation effects may be partly responsible. Many hotspots occur in regions that overlie faster than average parts of the transition region. This throws doubt on the hypothesis that hotspots are deeply rooted. On the other hand, the most geochemically distinctive hotspots occur in regions of thick crust, thick lithosphere or where the upper 50 km are faster than average, and in regions where velocities are low between 150 and 400 km. Hotspots are generally on the edges of these low–velocity anomalies. Hawaii, for example, sits on old, cold Pacific lithosphere, which may be locally thinned and is on the northern boundary of the Pacific low–velocity anomaly. The local crust is also thick. The Cretaceous seamounts and plateaus in the western Pacific were located over the central Pacific velocity anomaly when they were formed. The extensive volcanism may have been shut off by thick overriding Pacific lithosphere or by migration of the East Pacific Rise to its present location. Although separation of melt and residual crystals may occur at depths of the order of 100 km or shallower, the ultimate source region, and the instability that initiates adiabatic ascent into the shallow mantle, appear to be deeper than 350 km and, possibly, in the transition region. The effect of pressure on melt density and viscosity removes the necessity for postulating rapid removal of melt and small degrees of partial melting in deep reservoirs. At high pressure melt densities may approach the densities of olivine and pyroxene (RIGDEN *et al.*, 1984). If so, the melt will not separate from the matrix, particularly if the melt viscosity is also high at high pressure or if the rock permeability is low. On the other hand, because the dense phase garnet is reduced or eliminated, upon partial melting, the partially molten rock will be less dense than it was before heating and melting. A deep garnet–rich source region can therefore become lighter than adjacent or overlying material when it is heated, even if the melt density is comparable to the density of olivine. Garnet exsolution at high temperature has a similar effect. These processes can be thought of as exaggerated forms of thermal expansion, which can initiate and substain buoyancy driven convection. A thermal boundary layer near 400 or 650 km or the intersection of the geotherm with the solidus near 400 km may cause the instability. If hotspots and ridges share the same deep fertile source, the contrasting geochemical characteristics of the basalts may reflect different degrees of contamination and crystal fractionation, including clinopyroxene and garnet, at shallow mantle depths (ANDERSON, 1985). Although the lateral extent of the high–velocity regions in the lower part of the upper mantle cannot be determined with the available resolution, they would not show up at all if they had the dimensions of slabs. The piling up of cold material over long periods of geological time seems to be implied. This in turn is consistent with stratified mantle convection.

The fact that subshield velocities lie on the 1400°C adiabat below 200 km indicate that the shield plate is of the order of thickness of the high–velocity LID (150 km) plus a 50–km thick thermal boundary layer. This is consistent with the thickness estimated from the geoid by TURCOTTE and MCADOO (1979).

There is no support from either high–resolution body wave studies or global tomography for a thick continental tectosphere. The relatively high velocities between 200 and about 400 km under shield regions indicate the absence of partial melt rather than a compositionally distinct buoyant region that is permanently attached to the shield. A small perturbation in temperature, or an instability in the thermal boundary layer, would raise the temperatures between 150–200 km above the wet solidus. This is perhaps the trigger for kimberlite intrusion. At depths below some 300 km, the fastest regions of the mantle are not all located beneath the shields. Below 400 km much of the Canadian Shield is slow (see also GRAND, 1986).

PETROLOGICAL IMPLICATIONS

There has always been cross–fertilization between seismology and petrology. For example, the standard petrological view that the source region of basalts is olivine–rich is partially based on the early observation that the compressional velocities in the upper mantle are similar to olivine. Hidden assumptions were that no other combination of minerals would satisfy the seismic data and that basalts sampled the same part of the mantle as the seismic data. More recent seismological studies and studies of mantle xenoliths have strengthened the conclusion that the shallow mantle at least, above 200 km, is olivine–rich. The seismic data include velocities and anistrophy. The continental lithosphere, the top of the oceanic lithosphere and the upper mantle below the lithosphere and above 400 km are consistent with an olivine–rich peridotite. The main unsampled region is the lower oceanic lithosphere. Unfortunately, the velocities of olivine and orthopyroxene, on the one hand, and clinopyroxene and garnet, on the other hand, both bracket the observed velocities in unmelted regions of the upper mantle. Therefore, one cannot discriminate between peridotitic and eclogitic assemblages on the basis of seismic velocities alone and the seismic constant is lost. The seismic anisotropy of the upper 200 km slightly favors the peridotitic hypothesis. In a large portion of the shallow mantle the velocities are too low to be explained by any combination of minerals. The presence of a small melt fraction can explain these results but it removes the possibility of inferring mineralogy.

Seismological evidence for a low–velocity zone and its interpretation as a partial melt zone seemed to provide petrologists with a source region for basalts. Most petrologists now assume that not only do basalts come from the shallow mantle but that the ultimate source is also shallow. The LVZ is equated with "the source region" generally for depleted MORB–type basalts. The alternative is that basalts or their source rocks are brought into the shallow mantle by deeper processes and that basalts only record the final stages of their evolution.

The presence results show that the LVZ is not a global layer of uniform thickness or velocity nor does it invariably terminate at some shallow depth such as 150 or 200 km. It can be traced to much greater depths under some ridges and tectonic regions suggesting that petrologists should start to be concerned about deeper processes. The broad areal extent of low–velocity regions, compared to the area of volcanic centers, suggests that sublithospheric cooling and fractionation, fractionation that may include garnet and clinopyroxene, may be important petrological processes. The possible great depth of the initiation of adiabatic ascent suggests that much larger degrees of partial melting might be involved than have previously been considered plausible. Large degrees of partial melting at depth, followed by large amounts of crystal fractionation at modest depth must be considered as a viable alternative to small degrees of partial melting, at shallow depth, of a primitive pyrolite–like reservoir.

In a way, we have taken a step backwards. The recognition that peridotites and eclogites can have similar seismic velocities, at least above 400 km, and that the lowest velocity regions of the mantle have velocities lower than any subsolidus assemblage mean that we cannot constrain the mineralogy of most of the upper mantle. The anisotropy of the upper oceanic lithosphere and the very high velocities and V_s/V_p ratio of the shield mantle are evidence that these regions are probably peridotitic but these regions are not pertinent to basalt petrogenesis. The anisotropy of the upper 200 km of the mantle is weak evidence that this region, on average, is olivine–rich. Olivine, being a refractory residual phase and being less dense than fertile peridotite, should concentrate in the shallow mantle as a result of basalt separation throughout Earth history. In contrast to garnet– and clinopyroxene–rich assemblages, olivine does not undergo any temperature or pressure induced phase changes at depths less than 400 km. It is therefore permanently trapped in the shallow mantle. Basalts rising from a deeper source must, of course, traverse this olivine–rich region and may, to some extent, be trapped in it, converting it to a fertile peridotite. Basalts, however, are less dense than olivine in the shallow mantle and they are more likely to be trapped below the lower density crust or colder impermeable, high–viscosity lithosphere, if they cannot proceed directly to the surface. If only a fraction of this melt escapes to the surface then there may be a shallow basalt–rich layer which will cool to eclogite at depths greater than about 50 km. Such a layer could contribute to the eventual instability of the oceanic lithosphere and, possibly, to midplate volcanism.

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