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The Deep Structure of Continents

DON L. ANDERSON

Seismological Laboratory, California Institute of Technology, Pasadena, California 91125

The Lehmann discontinuity at 220-km depth is an important global feature which occurs under both oceans and continents. It is a barrier to penetration by young lithosphere and marks the base of seismicity in regions of continent-continent collision. The strong lateral variation in upper mantle velocities occurs mainly above this depth. Continental roots extend no deeper than about 150–200 km. The basalt-eclogite transformation and eclogite-harzburgite separation may be responsible for the geometry of intermediate depth earthquakes. Oceanic and continental geotherms converge above about 200 km and become less steep than the melting gradient at greater depth. This implies a low viscosity channel near 250 km. This would give a decoupling zone of maximum shear beneath continental shields. The Lehmann discontinuity may be the interface between two distinct geochemical reservoirs. The velocity jump, and the inferred density jump, at 220 km are consistent with an increase in garnet content. The mantle may be garnet lherzolite above and eclogite immediately below the Lehmann discontinuity. The transition region may be mainly eclogite and be the source region for oceanic tholeiites.

INTRODUCTION

Continental lithosphere is much thicker than oceanic lithosphere, but the question of how thick a section of continent translates coherently during continental drift has not yet been adequately addressed. The bottom of the low-velocity zone is usually considered to be the bottom of the asthenosphere and it has been presumed that coherent translation of both oceanic and continental plates takes place above some 200 km. This basic assumption of plate tectonics is contrary to the idea that ontinents have roots deeper than 400–500 km [*MacDonald,* 963; *Jordan,* 1975, 1979*a*, *b*].

Surface wave studies have shown that there are large differences between oceans and shields above about 220 km [*Dornan et al.*, 1960; *Brune and Dorman,* 1963; *Anderson,* 1967*a*; *Kanamori,* 1970; *Dziewonski,* 1971]. Recently, *Cara* [1979] has made a detailed study of regional differences using higher mode surface waves. He found strong regional variations between the Pacific Ocean, western U.S., and eastern U.S. above 250 km and no resolvable difference below this depth. *England et al.* [1978] made a direct comparison of upper-mantle structure under the North Atlantic and Arctic oceans and the old shield of the Russian platform. Even when maximum differences between the regions were allowed the *P* wave data could be satisfied by velocity models which were substantially the same below 300 km. Cara's models are shown in Figure 1.

Okal and Anderson [1975] used multiple ScS phases to sample the earth under various geological provinces including oceans and shields. They concluded that the observations were consistent with known differences above about 180 km. There is therefore good agreement between the body wave and surface wave data. *Jordan* [1975], however, proposed that ocean-continent differences extend deeper than 400 km and that the region which translates coherently in the course of plate tectonics may occupy the entire upper 700 km of the mantle. This proposal has reopened the question of the deep structure of continents.

THE DEEP STRUCTURE OF CONTINENTS

Prior to the recognition that attenuation was important in interpreting free oscillation periods, it was thought that average earth shear velocities were appreciably slower than continental values [*Jordan and Anderson,* 1974; *Jordan,* 1975]. *Hart et al.* [1977] determined an attenuation corrected free oscillation average earth model, QM2, which is compared with the continental model SHR14 of *Helmberger and Engen* [1974] in Figure 2. A more direct comparison is the Pacific–eastern U.S. curve which is from *Cara*'s [1979] surface wave study. In this case the major differences are above 250 km.

Okal and Anderson [1975] and *Sipkin and Jordan* [1976] studied the core reflected shear phase, ScS, with conflicting interpretations. Okal and Anderson concluded that all differences could be explained in terms of known effects above ~180 km, while Sipkin and Jordan concluded that differences in velocity must persist to great depths, perhaps extending throughout the entire upper mantle. *Okal* [1977] concluded that surface wave data, regionalized to take account of the age of the oceanic lithosphere, are incompatible with strong, deep lateral inhomogeneity and do not require any substantial structure variation below 240 km.

The ScS phase, of course, averages the velocity throughout the mantle and cannot resolve where the differences occur. There are more direct ways to isolate the effect. The velocity structure of the upper mantle has been studied in many regions by body wave and surface wave techniques. These studies give remarkably consistent results when the average travel times above 200 or 250 km are calculated. Table 1 presents these results. Shields are the fastest, about 3.5 s faster than young ocean. Oceanic models, on the average, are about 1.5–2 s slower than shields, but the difference decreases with the age of the ocean.

We have computed the differences between the *S* wave travel times above 200 km for the Rayleigh wave models and the Jeffreys-Bullen values. Figure 3 displays the computed ScS residuals for these oceanic and shield models as a function of crustal age. Measured residuals relative to Jeffreys-Bullen, are also shown. It is clear that the ScS times can be explained by known differences in the upper 200–250 km of the mantle.

The average one-way ScS residual for 'average age' ocean (70–90 m.y.) is +0.8 s. This can be compared with the average residual of +2.0 s for all oceanic data combined, including the very slow young oceans [*Sipkin and Jordan,* 1976]. This latter value is a straight average of all data and ignores the variation with age.

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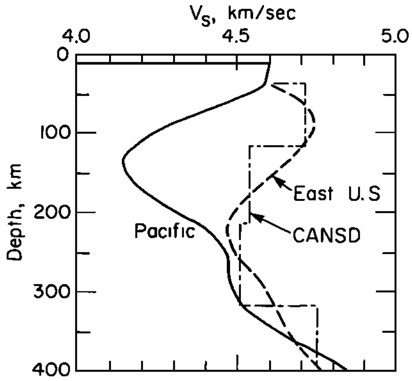

Fig. 1. Shear velocity versus depth for Pacific (average age, 90 m.y.) and eastern U.S. from *Cara* [1979]. CANSD is Canadian shield from *Brune and Dorman* [1963].

The residuals for the oldest ocean, >120 m.y., are scattered, but the mean is −0.7 s. This includes anomalously slow readings from the mid-Pacific mountains and the Bermuda Rise. Therefore the one-way difference in travel times between shields and old oceans is about 1.3 s. The difference between average ocean and old ocean is about the same as determined by *Duschenes and Solomon* [1977] using shear waves from Pacific events.

From the raw ScS data the mean ocean-continent differential time is +1.5 ± 2.8 s, about the same as the upper 250 km alone. Correcting for attenuation reduces the differential time slightly. The ScS shield data overlap the shield models, which differ from the average earth only above 250 km, but average about 1 s faster (Figure 3).

From independent data (Table 1) the travel times above 200–250 km in shields average 1.6 ± 0.6 s faster than under average oceans and 0.9–2.0 s faster than 70–100 m.y. old oceans. Therefore the ScS data are in good accord with the surface wave studies, and it appears that all differences can be accommodated above 250 km.

The ScS data, when corrected for crust and upper mantle effects above 50 km, suggest that the mantle under shields may be as much as about 3.4 s faster than under old oceans.

At this point, it is instructive to estimate the maximum plausible variations in the upper mantle. The shear velocity in the low-velocity zone (LVZ) in oceans and tectonic regions is about 10% lower than subcrustal velocities. This can be accounted for by a difference in chemistry or by high-temperature stress relaxation mechanisms such as dislocation or grain boundary relaxation [*Anderson and Minster,* 1979]. Measured shear wave velocities in the LVZ are 12–15% lower than the high-frequency velocities for mineralogies in assemblages ranging from pyrolite to eclogite [*Anderson,* 1977]. In addition to the temperature effect an additional several percent variation is allowed by variations in mineralogy. We take 12% as a plausible variation between shield and ocean mantles, take a travel time of 57 s above 250 km, and assume that velocities are the same at ~250 km. This gives 3.4 s as a conservative estimate of possible upper mantle shear wave vertical travel time variations. This would be the difference between a cold, garnet-rich upper mantle and a warm, relaxed, garnet-poor

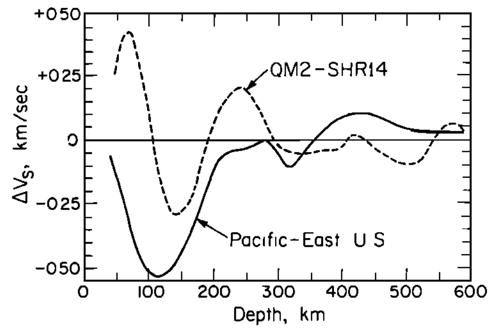

Fig. 2. Ocean-continent shear velocity differences versus depth. Solid line is from *Cara* [1979]. Dashed line is attenuation corrected average earth model QM2 [*Hart et al.,* 1977] minus continental model SHR14 [*Helmberger and Engen,* 1974].

TABLE 1. Upper Mantle Vertical Shear Wave Travel Times Above 200 and 250 km

	Travel Time, s		
	200 km	250 km	Reference
Shield	45.2 ± 0.3	56.1 ± 0.3	(1)
Continent	45.9 ± 0.5	57.0 ± 0.2	(2)
Ocean	46.7 ± 0.4	57.7 ± 0.3	(3)
Minimum	49.2	60.6	(4)
15 m.y.	48.7	59.8	(5)*
70 m.y.	47.1	58.1	(5)*
100 m.y.	46.4	57.5	(5)*
150 m.y.	45.8	56.8	(5)*
Maximum	46.0	57.2	(4)

(1) *Anderson* [1967*a*], *Anderson and Harkrider* [1968], *Brune and Dorman* [1963], *Wickens* [1971], and *Massé* [1973].

(2) *Helmberger and Engen* [1974], *Anderson and Julian* [1969], *Cara* [1979].

(3) *Anderson* [1967*a*], *Kanamori* [1970], *Schlue and Knopoff* [1976], and N. R. Burkhard and D. D. Jackson (unpublished manuscript, 1979) (average ocean).

(4) *Yoshida* [1978]; minimum and maximum age groups.

(5) N. R. Burkhard and D. D. Jackson (unpublished manuscript, 1979).

**SV* velocities from Rayleigh waves. Horizontally traveling *SH* waves (Love waves) give travel times 1.6 s (15 m.y.) to 0.5 s (150 m.y.) shorter. ScS waves should be compared with Rayleigh wave, not Love wave, velocities. 'Average' ocean (~70 m.y.) is about 2 s slower than shield. If Love waves are used in the comparison oceans would appear to be about 3 s slow.

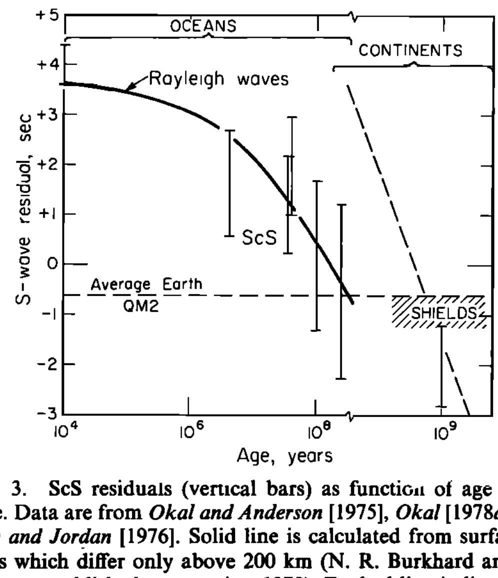

Fig. 3. ScS residuals (vertical bars) as function of age of lithosphere. Data are from *Okal and Anderson* [1975], *Okal* [1978*a, b*], and *Sipkin and Jordan* [1976]. Solid line is calculated from surface wave models which differ only above 200 km (N. R. Burkhard and D. D. Jackson, unpublished manuscript, 1979). Dashed line indicates range of continental residuals [*Hales and Roberts,* 1970; *Poupinet,* 1979]. 'Average earth' is free oscillation model of *Hart et al.* [1977] corrected for attenuation. Region marked shields is calculated from body wave and surface wave shield models.

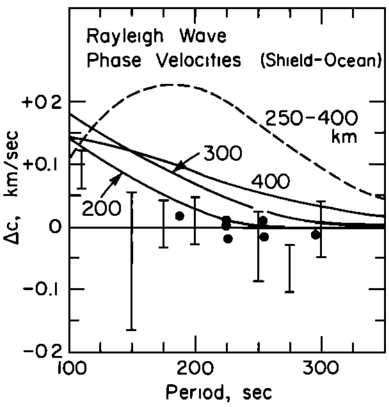

Fig. 4. Rayleigh wave differential phase velocities (shield minus ocean). Data from *Kanamori* [1970], *Brune and Dorman* [1963], *Okal* [1978*a*, *b*], and *Cara* [1979]. Solid lines show the effect of distributing a 3.4-s difference in travel time, as implied by the ScS data, between 50 km and the depth shown. The dashed line shows the effect of placing a 5-s difference between 250 and 400 km. The ScS and Rayleigh wave data are compatible if the difference is above 200 km.

upper mantle. In this case the suboceanic upper mantle can be interpreted as residual or depleted material relative to subshield mantle.

The near-vertical ScS data cannot isolate the depth range responsible for the variation. For this purpose we can use the dispersion of Rayleigh waves. Figure 4 gives the difference in phase velocity between oceans and shields as a function of period. The data are from *Brune and Dorman* [1963], *Kanamori* [1970], *Okal* [1978*a*, *b*], and *Cara* [1979]. The solid curves show the effect of distributing a 3.4-s difference from 50 km to the depth shown. It is clear that the dispersion data are satisfied if the oceanic delay is placed above 200 km. Putting a 5-s delay between 250 and 400 km is clearly not acceptable.

The shorter period Rayleigh wave data (20–60 s) including higher modes can be satisfied by a 3.2-s delay between 50 and 250 km under oceans compared to the eastern U.S. [*Cara*, 1979]. Cara's eastern U.S. model is similar to the Canadian shield model of *Brune and Dorman* [1963] and gives similar travel times above 250 km.

Jordan [1975] noticed that the difference between oceanic and continental Love wave phase velocities was much less than would be predicted from the observed differences in ScS travel times if the variations were restricted to shallow depths, for example, above 400 km. He tacitly assumed that horizontally and vertically traveling *SH* waves had the same velocity. In a transversely isotropic media these velocities are different, but vertically traveling *SH* has the same velocity as horizontally traveling *SV*, i.e., ScS times should be compared with Rayleigh wave velocities, or with models based on Rayleigh wave data, as we have done above. The Love wave and ScS observations are consistent if the oceanic upper mantle in the vicinity of the low-velocity zone has a shear wave anisotropy of about 5%. This is about the same as required to reconcile Love wave and Rayleigh wave data [*Anderson*, 1966; *Anderson and Harkrider*, 1962; *Schlue and Knopoff*, 1976; N. R. Burkhard and D. D. Jackson, unpublished manuscript, 1979]. Thus this seems to be a reasonable alternative to deep (>200 km) continental roots.

There is therefore good consistency between the free oscillation, surface waves and ScS observations when the effects of anelasticity and ocean age are taken into account. ScS and Love waves do not average the upper mantle in the same way. They should not be compared unless anisotropy is taken into consideration. An apparent transverse isotropy can be the result of fine layering which will not be detected directly if the seismic wavelengths are longer than the layer thicknesses [*Anderson*, 1966]. For example, a thin low-rigidity layer under the oceanic lithosphere, possibly due to melt accumulation by drainage from the upper mantle, will give an apparent transverse isotropy if not allowed for in the modeling. Variations in the thickness of such a layer could also explain the large spread in residuals of ScS in older ocean basins.

The Lehmann Discontinuity

The major seismic discontinuities in the mantle are near 400 and 670 km. These have been interpreted as phase changes, although the deeper one may involve a composition change as well [*Anderson*, 1967*b*, 1968]. A composition change would be an effective barrier to convection and explain the termination of seismic activity at ~670 km. The sharpness of this discontinuity [*Adams*, 1971; *Whitcomb and Anderson*, 1970] together with the large increase in velocity argues for a change in chemistry as well as a change in crystal structure. There is another important mantle discontinuity at a depth near 220 km, the base of the low-velocity zone.

A discontinuity at 232-km depth was proposed in 1917 by Galitzin. The most detailed early studies indicated the presence of a discontinuity under North America and Europe near 215–220 km [*Lehmann*, 1959, 1961, 1967] and we shall henceforth refer to it as the Lehmann discontinuity. The early work is summarized in *Anderson* [1966, 1967*a*] and *Knopoff et al.* [1966]. Additional evidence has accumulated since these summaries.

A. L. Hales et al. (unpublished manuscript, 1979), *Steinmetz et al.* [1974], *Lukk and Nersesov* [1965], and *Wiggins and Helmberger* [1973] have all found evidence for a discontinuity between 190 and 230 km from body wave data. *Cara* [1979] found high velocity gradients near 220 km. The increase in velocity is the order of 3.5–4.5%. Using the seismic equation of state [*Anderson*, 1967*c*], the associated density increase is about 3%.

Niazi [1969] demonstrated that the Lehmann discontinuity in California-Nevada is a strong reflector and found a depth of 227 ± 22 km.

Sacks et al. [1977] and *Jordan and Frazer* [1975] found converted phases from a discontinuity at a depth of 200–250 km under both the Canadian and Baltic shields. Reflections from a similar depth have been reported from *P′P′* precursors [*Adams*, 1971; *Whitcomb*, 1973; *Whitcomb and Anderson*, 1970] for Siberia, western Europe, North Atlantic, Atlantic-Indian Rise, Antarctica, and the Ninety-east Ridge. Evidence now exists for the Lehmann discontinuity in eastern and western U.S., Canadian Shield, Baltic Shield, oceanic ridges, normal ocean, the Hindu Kush, the Alps, and the African rift.

The V_p/V_s ratio of recent global earth models [*Hart et al.*, 1977] reverses trend at 220 km. This is indicative of a change in composition, phase, or temperature gradient.

There are not yet enough seismic data to map the variability in depth of the '220-km' discontinuity. Most reported reflections occur at depths between 190 and 230 km. Part of this variation is due to assumptions about the mantle velocity above the reflector.

The geopotential power spectrum yields a depth of 200 km for a density discontinuity [*Lambeck*, 1976; *Marsh and Marsh*, 1976]. This gives additional evidence for the interface and in-

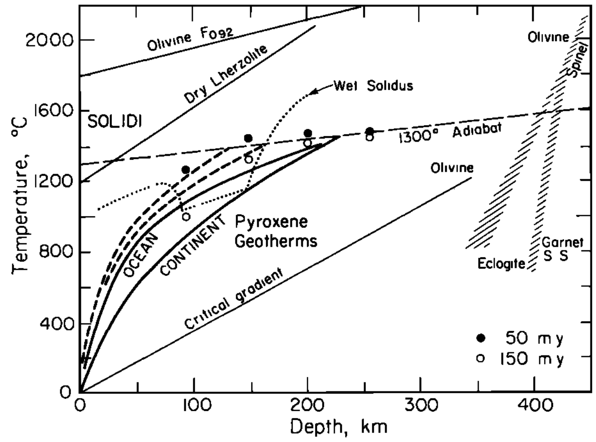

Fig. 5. Oceanic and continental pyroxene geotherms from *Mercier and Carter* [1975] and theoretical oceanic geotherms at 50 and 150 m.y. from *Schubert et al.* [1978]. The critical gradient [*Kumazawa and Anderson,* 1969] is for constant shear velocity versus depth in a homogeneous olivine mantle. The critical gradient for garnet is not much different. The olivine-spinel and garnet + pyroxene to garnet solid solution boundaries are from *Akaogi and Akimoto* [1977].

dicates that it is variable in depth. The depth range depends on the density contrast but need only be a few kilometers.

Thus there is a variety of evidence of support of an important discontinuity near 220 km. This discontinuity affects seismicity and may be a density or mechanical impediment to slab penetration. It marks the depth above which there are large differences between continental shields and oceans. Few earthquakes occur below this depth in continental collision zones and in regions where the subducting lithosphere is less than about 50 m.y. old.

Seismicity

In most seismic regions, earthquakes do not occur deeper than about 250 km. This applies to oceanic, continental, and mixed domains. The maximum depths are 200 km in the South Sandwich arc, Burma, Rumania, the Hellenic arc, and the Aleutian arc; 250 km in the west Indian arc; and 300 km in the Ryukyu arc and the Hindu Kush. There are large gaps in seismicity between ~250 km and ~500–650 km in New Zealand, New Britain, Mindanao, Sundu, New Hebrides, Kuriles, North Chile, Peru, South Tonga, and the Marianas [*Isacks and Molnar,* 1971]. In the New Hebrides there is a concentration of seismic activity between 190 and 280 km that moves up to 110 and 150 km in the region where a buoyant ridge is attempting to subduct [*Chung,* 1979]. In the Bonin-Mariana region there is an increase in activity at 280–340 km to the south and a general decrease in activity with depth down to about 230 km. Where earthquakes reach as deep as 400 km, there is a pronounced gap below 150 km. In the Tonga-Kermadec region, seismic activity decreases rapidly down to 230 km and, in the Tonga region, picks up again at 400 km. In Peru most of the seismicity occurs above 190–230 km, and there is a pronounced gap between this depth and 500 km. In Chile the activity is confined to above 230 km and below 500 km.

Cross sections of seismicity in these regions suggest impediments to slab penetration at depths of about 230 and 600 km. Oceanic lithosphere with buoyant ridges seems to penetrate only to 150 km.

Compressional stresses parallel to the dip of the seismic zone are prevalent everywhere that the zone exists below about 300 km, indicating resistance to downward motion below about this depth [*Isacks and Molnar,* 1971]. Actually, between 200 and 300 km about half the focal mechanisms indicate down dip compression, and most of the mechanisms below 215 km are compressional. *Isacks and Molnar* [1971] suggested that the slabs encounter stronger or denser material which resists their sinking.

We propose that all these observations are consistent with mechanical barriers near the 670 km and Lehmann discontinuities. A small intrinsic increase in density, due to a change in chemistry, is a very effective brake to penetrative convection. For example, a 3% difference in intrinsic density can be offset only by a large temperature differential of 1000°C. A similar decrease in temperature is required to elevate the olivine-spinel phase boundary to 250 km in the slab.

There is a relationship between age of subducted plate and penetration depth [*Vlaar and Wortzel,* 1976], suggesting that thin lithosphere cannot subduct to great depth. Old lithosphere on the other hand is not only colder but may be intrinsically denser if it grows by freezing eclogite onto its base.

We believe that the seismicity patterns may be controlled by the mantle discontinuity near 220 km. There are no important first-order phase changes in the mantle near this depth [*Ringwood,* 1975]. This plus the sharpness of the discontinuity suggest that there is a chemical change. An increase in garnet content is the most reasonable way to increase the density. The Lehmann discontinuity may be the boundary between depleted and fertile lherzolite or between peridotite and eclogite.

In regions of continent-continent collision the distribution of earthquakes should define the shape and depth of the collision zone. The Hindu Kush is characterized by a seismicity pattern terminating in an active zone at 215 km [*Santo,* 1969]. A pronounced minimum in seismic activity occurs at 160 km. Again, the Lehmann discontinuity appears to mark the lower boundary of the moving plates.

We propose that the harzburgite portion of the slab remains above ~250 km and only old slabs can penetrate deeper. The eclogite part of the lithosphere is denser than spinel or garnet peridotite. We suggest that it is a density barrier rather than a strength barrier that is responsible for the distribution and stresses of intermediate depth earthquakes. With this model only the eclogite portion of the lithosphere can penetrate below 220 km and the uppermost mantle will be rich in olivine and orthopyroxene. The upper mantle below the Lehmann discontinuity is richer in garnet.

The composition of the mantle between 220 and 670 km is the subject of a separate paper. The seismic data are consistent with eclogite in this region. This may be the source, as well as the sink, of oceanic crust.

Temperatures and Viscosity in the Mantle

Mercier and Carter [1975] have reanalyzed xenolith data and derived the continental and oceanic pyroxene geotherms shown in Figure 5. They converge above 200–250 km. The colder oceanic curve is their preferred solution for normal ocean. The points for 50 and 150 m.y. oceanic mantle are from a theoretical discussion of *Schubert et al.* [1978]. There is good agreement between the estimates of temperature using geophysical and petrological techniques, and no evidence for deep, >200 km, differences correlated with shields. *Solomon* [1976] also made this point.

The shear velocity under shields increases with depth, or is constant, to about 100 km in spite of the fact that the shield

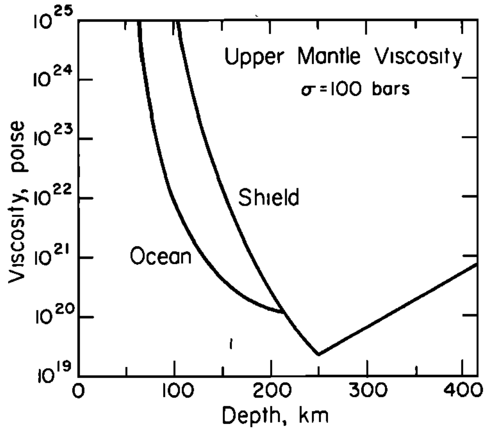

Fig. 6. Viscosity versus depth for old ocean (cold ocean geotherm in Figure 5) and shields. Dislocation climb in olivine is assumed. The viscosity varies as the square of the assumed stress.

geotherm is steeper than the critical gradient for a low-velocity zone down to about 200 km. This requires that the mineralogy and/or composition change with depth. This suggests that the garnet content increases, in agreement with the petrology of kimberlite pipes [*Boyd and McCallister,* 1976; *Jordan,* 1978]

The stable continental upper mantle shear velocities above ~250 km are higher than oceanic shear velocities even when corrected for the difference in temperature. The difference, ~8%, can be due to partial melting or to dislocation relaxation [*Anderson and Minster,* 1979], or it may indicate a difference in chemistry.

The adiabatic gradient is less than the critical gradient. Therefore below some 200 km the shear velocity should increase with depth and the K/μ ratio should reverse, as observed. This is also the depth at which the geotherms are closest to the melting point of mantle minerals. This is the condition for minimum viscosity.

The temperature structure between shields and ocean basins leads to substantial differences in viscosity. The viscosity η depends on stress, temperature, and pressure. For dislocation climb [*Nabarro,* 1967],

$$\eta = \pi kTG^2/Db\sigma^2$$

where G, D, b, and σ are, the shear modulus, diffusivity, Burger's vector, and stress, respectively, and kT has the usual meaning. The diffusivity D is a strong function of temperature and presure,

$$D = D_0 \exp(-gT_m/T)$$

Where T_m is the pressure dependent liquidus temperature of the major phase. For the upper mantle it is usually assumed that olivine controls the rheology. From the geotherms of Figure 5 and constants from *Ashby and Verrall* [1978] and *Goetz* [1978], we calculate the viscosity profiles of Figure 6. The shield and ocean values are the same below 200 km and exhibit a minimum at about 230 km where the geotherms join the 1300°C adiabat. Below this depth the viscosity increases because the adiabat diverges from the melting curve. Note that above 150 km, shields are at least $1\frac{1}{2}$ orders of magnitude more viscous than oceans. The mantle will be most fluidlike near 230 km, and this is the most likely horizon for a shear boundary layer between the continental plate and the underlying mantle. A thermal boundary layer associated with a chemical change at the Lehmann discontinuity would make an even more pronounced minimum viscosity channel at this depth, and would make this region the most likely source for mantle diapirs, the precursors of basaltic volcanism.

SUMMARY AND DISCUSSION

There has been to date no seismic study which has detected resolvable ocean-shield differences in velocity below about 250 km. The large observed variation in ScS times can be accounted for by changes above this level which also corresponds to a seismic and seismicity discontinuity. Deeper variations also exist but there is no evidence that they are rigidly coupled to shallower plate motions. The study of deep lateral variations by combining seismic data of different types is complicated by the necessity of allowing for anelasticity, anisotropy and lithospheric aging.

The most detailed recent studies are *England et al.* [1978] for body waves and *Cara* [1979] for surface waves, including higher modes. In these studies the geometry and analysis techniques were particularly favorable for detecting such differences if they exist. Known variations above ~250 km are consistent with observed ScS times.

The geotherms under young oceans join the 1300°C adiabat at relatively shallow depths, ~150 km. Temperatures under older oceans and shields converge above about 200 km and join the 1300°C adiabat near 220 km. This will be a minimum viscosity channel and the decoupling zone for continental plates. Continental collision earthquakes will be mainly above this zone. The Lehmann discontinuity is probably due to a change in chemistry and represents a density barrier to slab penetration. The eclogitic portion of the oceanic lithosphere, however, can penetrate this barrier. Separation of eclogite and residual harzburgite may be responsible for intermediate depth earthquakes and the steepening of the seismic zone.

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