THE DEEP STRUCTURE OF CONTINENTS

By

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ABSTRACT

The Lehmann discontinuity at 220 km depth is an important global feature which marks the base of the lithosphere-asthenosphere system under both continents and oceans. It is a barrier to penetration of young lithosphere and marks the base of seismicity in regions of continent-continent collision. The distinction between oceans and continents disappears below this depth. Continental roots, therefore, extend no deeper than about 200 km. The reversible part of plate tectonic convection occurs above about 220 km. The mantle between 220 and 650 km may be predominantly eclogite, the remains of subducted oceanic crust. The basalt-eclogite transformation and eclogite-peridotite separation may be responsible for the mechanisms and geometry of intermediate depth earthquakes. Oceanic and continental geotherms appear to converge above about 200 km and it is proposed that they join the adiabat somewhat deeper. This explains the seismic data and implies a low viscosity channel near 250 km. This would give a decoupling zone of maximum shear beneath continental shields.
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 continentally plates takes place above some 200 km. This basic tenet of plate tectonics is contrary to the idea that continents have roots deeper than 400-500 km (MacDonald, 1963, Jordan, 1975, 1979).

Surface wave studies have shown that there are large differences between oceans and shields above about 220 km (Dorman et al., 1960, Brune and Dorman, 1963, Anderson, 1967a, 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 oceancontinent 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). By comparing free oscillation and body wave models, Jordan (1975) concluded that much of the vertical travel time difference comes from velocity differences between 400-700 km and suggested a difference greater than 0.1 km/sec in this depth range. Newer studies reverse this conclusion. Hart et al. (1977) determined an attenuation corrected free oscillation average earth model, QM2, which is compared with the continental model SHR14 of Helmerberger and Engen (1974) in Fig. 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. Therefore, the large, >0.1 km/sec, difference below 400 km between stable continental interiors and oceans is not a requirement of the data.

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 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 seconds faster than young ocean. Oceanic models, on the average, are about 1.5 to 2 seconds slower than shields but the difference decreases with the age of the ocean.

Fig. 3 displays computed ScS residuals for these oceanic and shield models as a function of crustal age. These models involve differences only above 250 km. Measured values of ScS residuals 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 and no recourse need be made to deeper differences.

The average one-way ScS residual for "average age" ocean (70-90 m.y.) is +0.8 seconds. This can be compared with the average one-way residual of +2.0 seconds 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.

The mean residual for the oldest ocean, >120m.y., is about
-0.7 sec and 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 seconds. The difference between "average" ocean and old ocean is about the same as determined by Duchiens and Solomon (1977) using shear waves from Pacific events, and also with theoretical expectations based on the cooling plate hypothesis.

From the raw ScS data the mean ocean-continent differential time is $+1.5 \pm 2.8$ seconds, 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 one second faster (Fig. 3).

From independent data (Table 1) the travel times above 200-250 km in shields average $1.6 \pm 0.6$ faster than under "average" oceans and $0.9 - 2.0$ seconds faster than 70-100 m.y. old ocean. Therefore the ScS data is in good accord with the surface wave and body wave studies and all differences can be accommodated above 250 km.

The ScS data, when corrected for differences above 50 km, suggest that the mantle under shields is perhaps as much as about 3.4 seconds faster than under old oceans.

At this point it is instructive to estimate the maximum plausible variations in the upper mantle, say above 250 km. The shear velocity in the low-velocity zone in oceans and tectonic regions is about 10% lower than subcrustal velocities. This can be accounted for 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 to 15% lower than the high frequency velocities for mineralologies 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 the maximum plausible variation between shield and ocean mantles, take a travel time of 57 seconds above 250 km and assume that velocities are the same at ~250 km and calculate that 3.4 seconds is a conservative estimate of shear wave vertical travel time variations. This would be the difference between a cold, garnet-rich upper mantle and a warm, relaxed, garnet-poor upper mantle. Relative residuals of this order are common over short distances in both continental and oceanic environments and between young and old ocean. Observed differences of this order therefore do not require deep structure.

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. Fig. 4 gives the difference in phase velocity between oceans and shields as a function of period. The data is from Brune and Dorman (1963), Kanamori (1970), Okal (1978) and Cara (1979). The solid curves show the effect of distributing a 3.4 second difference from 50 km to the depth shown. It is clear that the dispersion data is satisfied if the oceanic delay is placed above 200 km. Putting a 5 second delay between 250 and 400 km is clearly not acceptable.

The shorter period Rayleigh wave data (20-60 seconds) including higher modes can be satisfied by a 3.2 second delay between 50-250 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 (1966) 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 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, Schule and Knopoff, 1976, Burkhard and Jackson, 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 wave 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.
THE LEHMANN DISCONTINUITY

The major seismic discontinuities in the mantle are near 400 and 650 km. These have been interpreted as phase changes although the deeper one may involve a composition change as well (Anderson, 1967b, 1968). A composition change would be an effective barrier to convection and explain the termination of seismic activity at ~700 km. The sharpness of this discontinuity (Adams, 1971, Whitcomb and Anderson, 1970) 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, at or near 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, 1967a) and Knopoff et al (1966). Additional evidence has accumulated since these summaries.

Hales et al. (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) using surface waves found high velocity gradients near 220 km. The increase in velocity is of the order of 3.5 to 4.5 percent. Using the seismic equation of state (Anderson, 1967c) the associated density increase is about 3 percent.
Niazi (1969) demonstrated that the Lehmann discontinuity in California-Nevada is a strong reflector and found a depth of $227 \pm 22$ 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, N. Atlantic, Atlantic-Indian Rise, Antarctica and the Ninetyeast 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 is 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 surface with undulations (Lambeck, 1976, Marsh and Marsh, 1976). This gives additional evidence for the discontinuity and indicates 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 below which there are no differences between continental shields and oceans and no earthquakes in continental collision zones. It is likely, therefore, that the motion of both oceanic and continental plates occurs above this level and that the reversible part of convection associated with plate tectonics is confined to above about 220 km.
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, N. Chile, Peru, S. 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 an impediment 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 a mechanical barrier near the Lehmann discontinuity. 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 Wortel, 1976) suggesting that warm lithosphere cannot subduct to great depth. Cold lithosphere on the other hand is not only intrinsically denser but can penetrate to great enough depth so that phase changes in the slab can provide the negative buoyancy required for penetration down to 650 km.

We believe that the seismicity patterns may be controlled by a 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 suggests that there is a chemical change. An increase in $\text{Al}_2\text{O}_3$ content would increase the
amount of garnet available, thereby increasing the density and the
seismic velocities.

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 vertical V-shaped 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 major portion of the slab get assimilated
above ~250 km and only very old slabs penetrate deeper. The crustal
fraction, ~5 km, however as eclogite is much denser than its
underlying residual peridotite. We propose that it is a density
barrier rather than a strength barrier that is responsible for the
distribution and stresses of intermediate depth earthquakes. The
viscosity also increases below 250 km but it increases at shallower
depths as well (see following section). With this model the mantle
between 220 and ~650 km is mainly eclogite and represents the burial
ground for subducted 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 Fig. 5. They converge above 200 to 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) pointed out that Jordan's (1975) conclusion that the thermal contrast between oceans and continents must persist to 400 km is based entirely on his particular choice of geotherms. Jordan's shield geotherm is 100-200°C below the corrected continental pyroxene geotherms. His oceanic geotherms, which are strictly hypothetical, are high by about the same amount. Solomon concluded that systematic continent-ocean differences do not appear to extend below 200 km, a conclusion consistent with Fig. 5.

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

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 olivine. This is the condition for minimum viscosity.

The temperature structure between shields and ocean basins leads to substantial differences in viscosity. The viscosity, \( \eta \), depends on stress, temperature and pressure. For dislocation climb (Nabarro, 1967)

\[
\eta = \frac{\pi kT G^2}{D b \sigma^2}
\]

where \( G, D, b \) and \( \sigma \) are, respectively, the shear modulus, diffusivity, Burger's vector and stress. The diffusivity, \( D \), is a strong function of temperature and pressure,

\[
D = D_0 \exp\left(-\frac{gT_m}{T}\right)
\]

where \( T_m \) is the liquidus temperature of the major phase. For the upper mantle it is usually assumed that olivine controls the rheology. From the geotherms of Fig. 1 and constants from Ashby and Verrall (1978) and Goetz (1978), we calculate the viscosity profiles of Fig. 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 fluid-like near 230 km and this is the mostly likely horizon for a shear boundary layer between the continental plate and the underlying mantle. Pyroxene and garnet comprise at least 40% of the upper mantle. Since they have lower melting points than olivine above 200 km and can form a continuous matrix, the viscosities may be lower than calculated. The actual value of the minimum viscosity is important if the return flow is confined to the upper mantle. This would be the case for a chemically stratified mantle.
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 fully accounted for by known 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 Englad 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 adiabat at relatively shallow depths, ~150 km. Temperatures under older oceans and shields converge above about 200 km and join the 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. Oceanic crust, as eclogite, however, can penetrate this barrier. Separation of eclogite and residual peridotite causes the seismic zone to steepen. The rising peridotite possibly initiates back-arc spreading.
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REFERENCES


Akaogi, N. and S. Akimoto, Pyroxene-garnet solid solution equilibria in the systems Mg$_4$Si$_4$O$_{12}$-Mg$_3$Al$_2$Si$_3$O$_{12}$ and Fe$_4$Si$_4$O$_{12}$-Fe$_3$Al$_2$Si$_3$O$_{12}$ at high pressures and temperatures, Phys. Earth Planet. Int., 15, 90-106, 1977.


Davies, G. F., Thickness and thermal history of continental crust and root zones (in press).


Poupmet, G., Relation entre le temps de parcours vertical des ondes sismigues et l'âge de la lithosphère continentale, Bul. de la Societe Geologique de France (in press).


Table 1
Upper Mantle Vertical Shear Wave Travel Times
above 200 km and 250 km

<table>
<thead>
<tr>
<th></th>
<th>200 km</th>
<th>250 km</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shields</td>
<td>45.2 ± 0.3</td>
<td>56.1 ± 0.3</td>
<td>(1)</td>
</tr>
<tr>
<td>Continent</td>
<td>45.9 ± 0.5</td>
<td>57.0 ± 0.2</td>
<td>(2)</td>
</tr>
<tr>
<td>Ocean</td>
<td>46.7 ± 0.4</td>
<td>57.7 ± 0.3</td>
<td>(3)</td>
</tr>
<tr>
<td>MIN</td>
<td>49.2</td>
<td>60.6</td>
<td>(4)</td>
</tr>
<tr>
<td>15 m.y.</td>
<td>48.7</td>
<td>59.8</td>
<td>(5)*</td>
</tr>
<tr>
<td>70 m.y.</td>
<td>47.1</td>
<td>58.1</td>
<td>(5)*</td>
</tr>
<tr>
<td>100 m.y.</td>
<td>46.4</td>
<td>57.5</td>
<td>(5)*</td>
</tr>
<tr>
<td>150 m.y.</td>
<td>45.8</td>
<td>56.8</td>
<td>(5)*</td>
</tr>
<tr>
<td>MAX</td>
<td>46.0</td>
<td>57.2</td>
<td>(4)</td>
</tr>
</tbody>
</table>

(1) Anderson (1967a), Anderson and Harkrider (1968), Brune and Dorman (1963), Wickens (1971), Massé (1973)
(3) Anderson (1967a), Kanamori (1970), Schlue and Knopoff (1976), Burkhard and Jackson (1978) ("average"ocean)
(4) Yoshida (1978); minimum and maximum age groups.
(5) Burkhard and Jackson (1978)

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

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

Figure 2 Ocean-continent shear velocity differences vs. 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).

Figure 3 ScS residuals (vertical bars) as function of age of lithosphere. Data from Okal and Anderson (1975), Okal (1978) and Sipkin and Jordan (1976). Solid lines are oceanic surface wave models from Burkhard and Jackson (1978) which differ only above 200 km. Dashed line indicates range of continental residuals from Hales and Roberts (1970) and by converting P-residuals from Poupinet (1979) using the Hales - Roberts relation between P and S residuals. "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 models. Note that continent residuals overlap oceanic residuals.
Figure 4  Rayleigh wave differential phase velocities (shield minus ocean). Data from Kanamori (1970), Brune and Dorman (1963), Okal (1978) and Cara (1979). Solid lines show the effect of distributing a 3.4 sec. 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 sec. difference between 250-400 km. The ScS and Rayleigh wave data are compatible if the difference is above 200 km.

Figure 5  Oceanic and continental pyroxene geotherms from Carter and Mercier (1975) and theoretical oceanic geotherms at 50 m.y. and 150 m.y. from Schubert et al. (1978). The critical gradient (Kumazawa and Anderson, 1969) is for constant shear velocity vs. 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).

Figure 6  Viscosity vs. depth for old ocean (cold ocean geotherm in Fig. 5) and shields. Dislocation climb in olivine is assumed. The viscosities will be less if pyroxenes control the rheology. The viscosity increases as the square of the assumed stress.
Fig. 1
Fig. 2
The graph illustrates the relationship between the age of different geological features and the observed S-wave residuals. The x-axis represents the age of the Earth in years, ranging from $10^4$ to $10^9$. The y-axis indicates the S-wave residual, measured in seconds. Several key features are highlighted:

- **ScS**: A specific seismic phase indicated by a central dot.
- **OCEANS** and **CONTINENTS**: Two broad categories distinguishing oceanic and continental crust.
- **Rayleigh waves**: A curved line denoting changes in S-wave residuals over time.
- **Average Earth QM2**: A horizontal line representing the average Earth's properties.
- **SHIELDS**: An area shaded to denote a particular geological condition or feature.

**Fig. 3**
Rayleigh Wave Phase Velocities. (Shield-Ocean)

\[ \Delta c, \text{ km/sec} \]

Period, sec

\[ 100, 200, 300 \]

\[ 250-400 \text{ km} \]

\[ 200, 300, 400 \]

Fig. 4
Fig. 5
Upper Mantle Viscosity

\[ \sigma = 100 \text{ bars} \]

Viscosity, poise

Depth, km

Fig. 6