Evidence against a discontinuity at the top of $D''$

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Summary. We investigate a recent suggestion of Lay & Helmberger that there is a world-wide discontinuity in $S$-wave velocity, about 250–300 km above the core–mantle boundary, with an increase of 2.5–3 per cent from above to below. Their model predicts pronounced pulse-form changes of long-period $SH$ body waves in the epicentral distance range 95–120°, a range which has not been studied by them. Similar pulse-form changes are expected in $P$-wave seismograms, under the reasonable assumption of a similar increase in $P$ velocity. Nine long-period seismogram sections of $SH$- and $P$-waves from deep-focus earthquakes below Fiji, West Brazil, the Sea of Japan, the Sea of Okhotsk and the Banda Sea are presented with practically no indication of the complications expected. Also, the decay constants of $P_{\text{diff}}$, as predicted by the new model, differ strongly from observed values. Thus, there is clear evidence against this model. The pulse-form peculiarities between mantle $S$ and $ScS$ in transverse-component seismograms at distances from 70° to 80°, which Lay & Helmberger interpreted as reflections from the discontinuity of their model, must have a different cause.

Introduction

Lay & Helmberger (1983) recently proposed a new model for the depth dependence of the $S$-wave velocity in the lowermost 400–500 km of the Earth's mantle. The main feature is a first-order discontinuity or narrow transition zone 250–300 km above the core–mantle boundary, i.e. roughly at the top of Bullen’s zone $D''$, characterized by a velocity increase from above to below of 2.5–3 per cent. Since Lay & Helmberger found seismic evidence for such a discontinuity below three widely different regions of the globe, they consider it as a world-wide feature, with possible lateral variations in depth. If true, this would be a result of seismological research with far-reaching consequences. For instance, a discontinuity in the lower mantle would imply either chemical inhomogeneity or a phase transition, and mantle-wide convection, if it exists, would probably be influenced seriously. However, before any non-seismological consequence is drawn, the seismological implications of the Lay &
Helmberger model should be investigated in detail. It is shown in the following that this model predicts rather pronounced effects in long-period seismograms which are additional to the seismogram signatures upon which Lay & Helmberger have based their conclusions. It is logical to search in long-period data for these effects; this is the main content of this paper.

One of the S-wave velocity models suggested by Lay & Helmberger, their model SLHO, is shown in Fig. 1, together with the corresponding travel-time curve for a deep-focus earthquake (depth 600 km). Lay & Helmberger state that they have seen the subcritical to critical reflections due to the discontinuity in SLHO. Their data are long-period WWSSN and CSN seismograms for the horizontal transverse component which for a radially stratified earth contain only $SH$-waves. Thus, their evidence comes from arrivals between the mantle $S$-wave and the core reflections $ScS$, seen at distances from 70° to 80° approximately. The reality of some of these arrivals can hardly be disputed, although they are a minor feature in the observed seismograms. Moreover, it is also true that Lay & Helmberger's synthetic seismograms for SLHO agree better with their data than synthetic seismograms for a lower-mantle velocity structure without a discontinuity. However, it is expected from the travel-time curve in Fig. 1 that more pronounced effects should be visible at distances beyond about 90°, where the supercritical reflection from the discontinuity and its continuation beyond the cusp point at about 97° interfere with the end of $ScS$ and the core diffraction $S_{diff}$. Synthetic seismograms, such as those included in Fig. 2, confirm this: long-period $SH$-wave seismograms with dominant periods around 20 s exhibit, for model SLHO, severe pulse-form changes in the distance range 95–120°, whereas for the smooth structure of the PREM model (Dziewonski & Anderson 1981) the pulse form remains the same, apart from a slight lowpass effect due to diffraction which is barely visible. Therefore it is the distance range 95–120° where a serious test of model SLHO is possible.

Figure 1. Left: S-wave velocity distributions in the lowermost mantle for the models SLHO of Lay & Helmberger (1983) and PREM (Dziewonski & Anderson, 1981). Velocity values in parentheses are assumed SLHO $P$-wave velocities. The jumps in $S$ and $P$ velocity at the discontinuity of SLHO are both 2.6 per cent. Elsewhere in the Earth SLHO and PREM agree. Right: S-wave travel times for SLHO and a focal depth of 600 km (reduced with 9 s deg$^{-1}$).
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For P-waves similar results hold. In Figs 4 and 5 synthetic P-wave seismograms are included for SLHO and PREM. The P velocities in the lowermost mantle of SLHO, which have neither been given by Lay & Helmberger nor discussed by them in general terms, were obtained by appropriate modification of the S velocities (see velocity values in parentheses in Fig. 1), and the parameters of the core of SLHO are those of PREM. For simplicity we continue to use the term 'model SLHO' although, of course, its P velocity is based on our assumptions, which may not agree with Lay & Helmberger's ideas. As expected, the synthetic P-wave seismograms also show pronounced differences between SLHO and PREM. Because of travel-time reasons similar interference conditions and hence pulse forms are found at different distances in the P and SH seismogram sections for SLHO; for instance, the SH waveforms at 96—102° (Fig. 2) are repeated in P at 100—110° (Figs 4 or 5).

We conclude that observed P-wave seismograms can and should be used for verification or refutation of SLHO. Of course, the assumption here is implicit that a change in chemical composition or phase in the lowermost mantle, if it exists, influences both wave velocities in about the same way. This assumption can be made more quantitative as follows. From standard formulae for the P and S velocities, \( \frac{\delta a}{a} \) and \( \frac{\delta \beta}{\beta} \), one obtains the relation between the relative velocity changes across the discontinuity in SLHO,

\[
\frac{\delta a}{a} = \frac{\delta \beta}{\beta} + \left( \frac{1}{2} - \frac{2\beta^2}{3a^2} \right) \left( \frac{\delta k}{k} - \frac{\delta \mu}{\mu} \right) \approx \frac{\delta \beta}{\beta} + 0.31 \left( \frac{\delta k}{k} - \frac{\delta \mu}{\mu} \right),
\]

where \( \delta k/k \) and \( \delta \mu/\mu \) are the relative changes in bulk modulus and rigidity. All changes in (1) are assumed to be small, and the density change is included in \( \delta \beta/\beta \). Only to the extent that the change in bulk modulus is significantly less than the change in rigidity, will the velocity changes \( \delta a/a \) be significantly less than \( \delta \beta/\beta \). We consider approximately equal changes \( \delta k/k \) and \( \delta \mu/\mu \) as being most probable under lower-mantle conditions.

Long-period seismograms at 90—115°

In the following we present and discuss observed SH- and P-wave seismogram sections for distances from about 90° to 115° relative to several deep-focus earthquakes below the Fiji region, West Brazil, the Banda Sea, the Sea of Okhotsk and the Sea of Japan. The source parameters of these events and other information are summarized in Table 1, and the epicentres are shown in Fig. 6; it is also indicated in this figure to which geographical region our results about the structure of the lowermost mantle apply. As indicated in Table 1, not all seismogram sections that we have assembled are shown in this paper. Those omitted have only slightly lesser quality than similar sections which are shown; they confirm the general conclusions drawn here.

The synthetic seismograms in Figs 2—5 were calculated with the reflectivity method. In all cases the same source—time function was used: the far-field waveform was chosen to agree with the P pulse, recorded at station ALQ for the event of 1974 March 23 (see Fig. 4). Special modelling would, of course, have been possible for each event, but the main conclusions can also be reached with an average source—time function. Crustal reverberations are not included in the calculations, and the Q structure assumed is that of PREM also in the case of SLHO.

When an average source—time function is used for several different earthquakes, it may appear at first unclear by which criterion agreement or disagreement of theoretical and observed seismograms can be judged. Various possibilities exist. The simplest is to construct and compare theoretical and observed amplitude—distance curves, taken from peak-to-peak amplitudes. This is a useful procedure for the investigation of smooth gradient zones, e.g. the
Table 1. Earthquakes used in this study (coordinates and magnitudes according to ISC).

<table>
<thead>
<tr>
<th>Date</th>
<th>Epicentre</th>
<th>Source region</th>
<th>Depth (km)</th>
<th>mb</th>
<th>Focal mechanism*</th>
<th>Phase, profile</th>
<th>Record section</th>
</tr>
</thead>
<tbody>
<tr>
<td>1969 Feb 10</td>
<td>22.75°S, 178.76°E</td>
<td>Fiji</td>
<td>635</td>
<td>6.0</td>
<td>SH, US</td>
<td>B, R</td>
<td>Fig. 2</td>
</tr>
<tr>
<td>1973 Dec 28</td>
<td>23.88°S, 180.00°E</td>
<td>Fiji</td>
<td>517</td>
<td>6.2</td>
<td>B, R</td>
<td>SH, US</td>
<td>Fig. 4</td>
</tr>
<tr>
<td>1975 June 22</td>
<td>24.98°S, 178.88°W</td>
<td>Fiji</td>
<td>333</td>
<td>6.1</td>
<td>B, R</td>
<td>SH, US</td>
<td>Not shown</td>
</tr>
<tr>
<td>1967 Feb 15</td>
<td>9.05°S, 71.34°W</td>
<td>West Brazil</td>
<td>598</td>
<td>6.1</td>
<td>IM</td>
<td>P, Europe</td>
<td>Fig. 5</td>
</tr>
<tr>
<td>1965 Nov 3</td>
<td>9.04°S, 71.32°W</td>
<td>West Brazil</td>
<td>587</td>
<td>5.9</td>
<td>IM</td>
<td>P, Europe</td>
<td>Not shown</td>
</tr>
<tr>
<td>1966 June 22</td>
<td>7.21°S, 124.69°E</td>
<td>Banda Sea</td>
<td>523</td>
<td>6.1</td>
<td>FM</td>
<td>pP, Europe</td>
<td>Fig. 5</td>
</tr>
<tr>
<td>1970 Aug 30</td>
<td>52.36°N, 151.64°E</td>
<td>Sea of Okhotsk</td>
<td>643</td>
<td>6.5</td>
<td>S</td>
<td>P, Central America</td>
<td>Fig. 5</td>
</tr>
<tr>
<td>1975 June 29</td>
<td>38.79°N, 130.09°E</td>
<td>Sea of Japan</td>
<td>549</td>
<td>6.1</td>
<td>SH, US</td>
<td>Figs 2, 3</td>
<td></td>
</tr>
</tbody>
</table>

† Own determination: nodal plane 1 strikes 20° and dips 60°SE, nodal plane 2 strikes 126° and dips 65°SW (essentially different from the solution in B).
‡ Own determination: nodal plane 1 strikes 47° and dips 75°NW, nodal plane 2 strikes 47° and dips 15°SE.

main part of the lower mantle or the outer core, or of discontinuities and thin transition zones, e.g. the inner-core boundary. In more complicated cases, such as the present one with possible interference of several arrivals, one may use the ratio of the largest positive and negative amplitudes which is more strongly influenced by interference than a peak-to-peak amplitude. Then, however, weaker arrivals, producing deflection in pulse flanks and additional oscillations, are disregarded, and they may actually be the clearest indication of structural complexities. Therefore, in the following we do not pay attention to peak-to-peak amplitudes or amplitude ratios as a function of distance, but to pulse-form peculiarities especially at the end of the SH or P wavegroup. Considering the synthetic seismograms in Figs 2–5, we can characterize the PREM pulses as stable throughout all distances, whereas the SLHO pulses show strong variations at their ends. Similar differences between PREM and SLHO would be present in synthetics that are calculated for the individual source-time function of each event (which actually is well represented by the average source-time function in almost all cases). Hence, the main conclusions can indeed be drawn from a comparison of the observations displayed in Figs 2–5 with the special synthetics given there.

We begin the discussion of Figs 2–5 with the SH-wave seismograms of Fig. 2. It is quite obvious that the records for the events of 1969 February 10 and 1974 March 23 are characterized by stability of the pulse forms for all distances; this is in agreement with the synthetic seismograms for PREM, but not with those for SLHO. In particular, the observations beyond 95° do not show any of the complications at the end of the SLHO pulses; the best examples are the records of OXF for both events and those of ATL, BLA and SCP for the event of 1969 February 10.

Figure 2. Observed transverse-component seismogram sections, constructed from long-period WWNSS records, and comparison with synthetic SH-wave seismograms for the models SLHO and PREM. All seismograms for one event have the same amplitude scale. Numbers after station codes are station azimuths.

The synthetic seismograms correspond to a focal depth of 600 km, the focal mechanism is of vertical dip-slip type, and the profile runs along the vertical P nodal plane where SH-waves are strongest. Times are reduced with 8 s deg⁻¹. Clipping of observed records is indicated by trace interruption.
The event of 1975 June 29 is a more difficult case. The records shown in Fig. 2 vary considerably with increasing distance, such that they probably should not be used for rejecting SLHO (except the record of LPS). In particular, the seismogram of SHA with its high frequencies and large amplitudes is problematic. However, because of the great station density between 93° and 100° seismogram development could be studied in detail for these distances (Fig. 3). Since on average this event radiated relatively high frequencies, the input pulse for the synthetics are made more high frequent by compressing its time-scale by 10 per cent, compared with all other calculations, without changing the pulse amplitudes. The SLHO synthetics beyond 96° have as the main characteristic a broadening of their minimum with increasing distance, which is not seen in the PREM synthetics. It is also not visible in the observed seismograms at DAL, JCT, OGD, OXF and GEO. The record of BLA, however, is more similar to the SLHO than to the PREM synthetics. Taking all observations together, we think that also the event of 1975 June 29 provides evidence against SLHO. This result is of interest since it casts doubt on the validity of model SLHO for just that geographical region, Alaska, for which it should apply according to Lay & Helmberger (see Fig. 6).

The P-wave sections of Fig. 4 also show stability of the pulse forms over all distances. In particular, the almost noise-free seismograms in the four largest distances for the event of
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Figure 4. The same as Fig. 2 for P-waves (vertical-component records). The focal mechanism for the synthetic seismograms is of thrust type with the T-axis vertical, and the profile is perpendicular to the strike of the P nodal planes. Times are reduced with 4.4 s deg⁻¹.
Figure 5. The same as Fig. 4 for three more earthquakes.
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1974 March 23, as observed in the United States, show no indication whatsoever of the pulse-form peculiarities of the SLHO synthetics. This result is of particular relevance, since these synthetics have been calculated for an input pulse which agrees with the seismogram of station ALQ for this same event. Also the other record sections of Fig. 4 each contain two or three seismograms at distances greater than 100° with clear evidence against SLHO.

The same conclusions have to be drawn from the data presented in Fig. 5. The seismograms in the three largest distances for the event of 1967 February 15 clearly disagree with the SLHO synthetics; very similar records, not shown here, are found at the same stations for the deep-focus earthquake of 1965 November 3 which had almost the same hypocentre. The event of 1970 August 30 had rupture complications which are visible in the first maximum of the P pulses, and the Central American stations LPS, SJG and BHP show a pronounced P-wave coda of uncertain origin. However, those parts of the P waveforms which are essential for the present discussion are not influenced by these complications. Also this event provides no evidence for and much evidence against SLHO. Finally, the pP arrival of the event of 1966 June 22 has a very stable pulse form for all distances. The input pulse used in the calculation of the synthetics is least suitable for modelling this event, but it is clear that closer modelling would have produced SLHO seismograms which also show pronounced pulseform changes with increasing distance, in disagreement with the observations. Hence, it is safe also to consider the event of 1966 June 22 as presenting evidence against SLHO.

Discussion

The long-period data shown in this paper which are representative of widely different regions of the globe (Fig. 6) do not support model SLHO. The stability of the SH and P pulses with increasing distance in the range 95–115°, that is found here for practically all events studied, is well explained by PREM or other models with similarly smooth velocity–depth functions in the lowermost mantle. Thus, layering of this depth range on a global scale appears improbable; we recall that 10 to 15 years ago other layered models have been suggested which were also in disagreement with long-period seismograms (Müller, Mula & Gregersen 1977).

Short-period synthetic seismograms for P-waves at distances from 70° to 110° that we have calculated show pronounced differences between SLHO and PREM, which is of course expected. Thus, high-quality short-period P-wave data will offer another means to study the reality of the discontinuity in SLHO. Array data that we have at hand for distances from 70° to 85° show no indication of an arrival between P and the onset time of PcP (Schlittenhardt 1984; manuscript in preparation). This is in agreement with the results of Ruff & Helmberger (1982) who investigated short-period P-waves in the distance range 80–105° and inferred a smooth velocity–depth function for the lowermost mantle below the Arctic.

The amplitude decay of $S_{\text{diff}}$ (SH component) and $P_{\text{diff}}$ into the shadow zone for SLHO is so different from the decay for PREM (see Figs 2 and 4), that observed decay constants offer another means to test SLHO. Fig. 7 illustrates this for $P_{\text{diff}}$, whose decay constants have been measured with sufficient accuracy (Doornbos & Mondt 1979; Mula 1981; Doornbos 1983). Since the observed values followed from averaging over many earthquakes and station azimuths, they are true global averages, and the disagreement with the predicted values for SLHO is another piece of evidence against this model.

The SH- and P-wave data presented in this paper argue against a world-wide increase of both S and P velocity by 2–3 per cent at the top of D'''. Since Lay & Helmberger did not comment on P velocities, we had to assume them; our assumption of a similar increase in P and S velocity appears most plausible for the rocks of the mantle. If the P velocity increase is reduced from 2.6 to 1.3 per cent, with all other parameters kept unchanged, the long-period SLHO seismograms for P waves are considerably closer to the PREM seismograms, such that
Figure 6. Epicentres of the earthquakes, used in this study, and average profiles to the stations which are in the distance range 90–115°, approximately. The profile endpoints are at 115°. Hatched areas in the middle of the profiles mark the regions to which the result of this study, no evidence for a discontinuity at the top of zone D°, applies.
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Figure 7. Observed decay constants for the decay of $P_{\text{diff}}$ into the shadow zone (triangles and squares) and comparison with theoretical decay constants for SLHO and PREM which are determined by spectral analysis of synthetic seismograms in the distance range 100–130°.

The $P$-wave observations collected here can no longer be used to disprove SLHO. Values of the $P$ velocity increase of 1.3 per cent and lower can only be disproved by short-period data. However, even if the velocity increase is restricted to the $S$ velocity alone, our $SH$-wave data argue against such a model, although the evidence comes only from two profiles, Fiji – United States and Sea of Japan – United States. Then, a conservative conclusion (less far reaching than ours), which accommodates Lay & Helmberger’s interpretation and our results, would be that there are pronounced lateral variations in $S$ velocity at the top of $D''$.

A question that is not answered here concerns the cause of the seismogram signatures from which Lay & Helmberger derived model SLHO. Other causes than a reflector in the lowermost mantle are conceivable, for instance

- coda arrivals of mantle $S$ due to reverberations in the crust and lithosphere or due to complications at the end of the rupture process;
- $SKS$ arrivals which are visible on the transverse component because of lateral heterogeneities in the Earth, an explanation which is favoured by Cormier (1985);
- $P$ to $ScS$ conversion produced at a scattering inhomogeneity somewhere below the focus, leading to $ScS$ precursors also on the transverse component.

Also, a superposition of such arrivals could be possible. We mention these points without claiming that they actually explain Lay & Helmberger’s observations. However, the fact that we must leave these observations unexplained does not question our refutation of model SLHO.

A still open question about zone $D''$ is whether in the sense of a globally averaged earth model the vertical velocity gradients are generally positive, as in the models favoured by us so far (Mula & Müller 1980; Mula 1981), or zero as in PREM, or partly also negative (Doornbos & Mondt 1979; Doornbos 1983). The true complications of $D''$, however, appear to reside rather in lateral velocity and $Q$ variations. If there are significant horizontal velocity gradients, the question of correct globally averaged vertical velocity gradients loses some of its importance.
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