ELSEVIER

Physics of the Earth and Planetary Interiors 92 (1995) 85–98

PHYSICS OF THE EARTH AND PLANETARY INTERIORS

On the structure of the lowermost mantle beneath the southwest Pacific, southeast Asia and Australasia

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Received 22 July 1994; accepted 19 October 1994

Abstract

The region of the lowermost mantle beneath the southwest Pacific, Australasia and southeast Asia (50°S–50°N and 80–190°E) has been studied using a wide variety of seismic techniques. We complement these studies with results obtained from long-period Global Digital Seismograph Network (GDSN) data using a recently developed phase-stripping technique that permits the isolation of D″ reflections from the stronger neighbouring S and ScS signals. We identify patches with D″ reflections and areas where we cannot confidently determine the presence or absence of a D″ reflection. A synthesis of our results with other studies suggests that D″ varies dramatically through the region, generally thickening from 100–150 km in a central zone to about 300 km at the eastern and western margins. However, there are irregularities in this overall pattern, including areas where there seems to be little evidence for a D″ discontinuity. Inspection of waveform amplitudes shows considerable scatter in not only the D″ reflected phases, but also core–mantle boundary reflected phases. Experiments with synthetic seismograms for a variety of D″ models and the observed lateral variability through the region suggest that this is to be expected. Furthermore, ray-theory calculations for large-scale 3D Earth models predict significant variations (±30%) in S-wave amplitudes of lowermost mantle turning rays. Finally, we investigate possible correlations between lower-mantle velocity, flow and D″ thickness. We find some correlation between D″ thickness and lower-mantle velocities obtained from tomographic inversions, with a thin D″ layer in high-velocity regions and a thickening of the layer toward slower regions. The relationship between predicted lower-mantle flow and D″ thickness is less clear. These results are qualitatively consistent with thermal boundary layer predictions for D″, but do not preclude the possibility of compositionally distinct material in the layer.

1. Background

In this paper we investigate the structure of the lowermost mantle in a region beneath the southwest Pacific, Australasia and southeast Asia (50°S–50°N and 80–190°E). Tomographic models of the large-scale structure of the mantle show that this region has some of the largest lateral variations in velocity in the lower mantle (see, e.g. models SH10C of Masters et al. (1992) and WM13 of Woodward et al. (1993)). Furthermore, a variety of more localized studies of this area have indicated even stronger lateral variations in

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SSDI 0031-9201(95)03063-8

smaller-scale structure of the D″ region (e.g. Revenaugh and Jordan, 1991; Wysession et al., 1994). D″ serves as a thermal boundary layer between the liquid outer core and the crystalline lower mantle, and convection and thermal instabilities are expected to cause some degree of lateral heterogeneity in the region (McKenzie et al., 1974; Yuen and Peltier, 1980; Jarvis and Peltier, 1984). It is not clear, though, to what extent observed D″ structure can be explained by thermal models and whether D″ is also a compositionally distinct region of the mantle (see Lay (1989) for a discussion of this). It has also been proposed that the discontinuity at the top of D″ may be due to a phase change (Nataf and Houard, 1993). Testing such hypotheses requires a detailed analysis of lower-mantle velocity structure. The region beneath the southwest Pacific is interesting to explore because the observed large lateral velocity differences suggest strong lateral variations in lower-mantle flow.

Previous work in this region has employed a wide variety of techniques for studying the lower-mantle. Wright et al. (1985) analysed lowermost-mantle turning P-waves using array data. Based on sharp changes in slowness variations, they found evidence for a discontinuity 200 km above the core–mantle boundary (CMB). In a much broader region of study, Young and Lay (1987) proposed a shear-velocity discontinuity 280 km above the CMB beneath the Indian Ocean and India. Their model was obtained through careful analysis of SH waveforms that indicated a reflected phase from the top of D″. In a systematic search of the Bulletins of the International Seismological Centre, Weber and Körnig (1992) found many regions where secondary arrivals between P and PcP were identified and many where they were not. The intermediate arrivals were interpreted as reflections from the top of a highly variable D″ discontinuity. Using stacking techniques, a P-wave discontinuity 320 km above the CMB was suggested by Neuberg and Wahr (1991). Garnero et al. (1993) analysed SH-wave data and identified reflections from a 280 km thick D″ layer in one region and from a 180 km thick layer in another. In an earlier study, Garnero et al. (1988) studied S–SKS travel-times and suggested that a D″ discontinuity on the edge of the Pacific faded moving westward into the Pacific. Although this hypothesis cannot be ruled out, it was later found that the differential travel-times could be fitted with a thin-layered D″ model. Revenaugh and Jordan (1991) analysed ScS reverberations, looking for secondary reflections within the reverberation interval. They found some corridors where a D″ reflection was evident, but others where it was not. Their estimates of D″ thickness varied from 270 km to 340 km. A study of P and S diffracted waves by Wysession et al. (1992) indicated variations of up to $\pm 1.5\%$ in the lowermost 190 km of the mantle. These results were corroborated in a recent study using diffracted P waves by Souriau and Poupinet (1994); they also found a slow region beneath northern Indonesia. In a subsequent study, Wysession et al. (1994) found dramatic variations in D″ velocities of -3% to 5% using ScS–S and sScS–sS travel-time residuals. The region analysed in this paper is roughly the same as that analysed by Wysession et al. (1994).

In a previous study (Kendall and Shearer (1994), hereafter referred to as KS94), we searched through 10 years of long-period SH-wave data from the Global Digital Seismograph Network (GDSN) looking for reflections from the top of D″. To do this we developed a phase-stripping technique for isolating D″ reflections from the stronger neighbouring SH and ScSH phases. In KS94 we observed a general thickening in D″ moving eastward beneath the Australasia region, with D″ thicknesses varying from 100 km to more than 350 km. Unlike KS94, this paper examines the results that did not suggest a D″ reflection as well as those that did, and also considers the amplitude variations of the reflected D″ phases. In an effort to explain the results we study waveform variations for a variety of models using WKBJ and ray-theory synthetics. We also explore the correlation between D″ thickness and lower-mantle velocity and flow proposed in KS94.

2. Estimates of D″ structure

The phase-stripping technique presented in KS94 is summarized in Fig. 1. Experience with synthetic seismograms shows that at epicentral distances between 63° and 74° a D″ reflection, SdS, will in most cases have the same polarity and character as the neighbouring S and ScS phases. This intermediate phase is much weaker and, for long-period waveforms, easily dominated by the neighbouring S and ScS phases. The phase-stripping technique is a way of isolating the D″ signal. The first step is to identify the SH and ScSH signals and cross-correlate them. If the cross-correlation is good, a reference pulse is constructed from the front part of the S phase and the back part of the ScS phase. The presumption is that if the S and ScS phases correlate well, the source and attenuation effects are the same for each phase and any intermediate D″ reflection. A reference trace is formed by superimposing scaled reference pulses aligned on S and ScS and then subtracted from the original trace. The remaining signal is cross correlated with the reference pulse, and, if the correlation is good, a D″ reflection is interpreted. As the epicentral distance approaches the critical angle for reflection, the waveform will distort owing to the ensuing phase shift. By restricting our search to distances less than 74° we mitigate this distortion. Nevertheless, it should be noted that the phase-stripping technique can still do a reasonable job of identifying the SdS phase even when it has experienced some phase distortion. Tests with synthetics show that these picks will generally give less conclusive results, as the final cross-correlation will not be as good. The technique is a localized estimate of the thickness of D″ and assumes a locally horizontal structure. If 3D structure distorts the waveform substantially, the technique will give an inconclu-

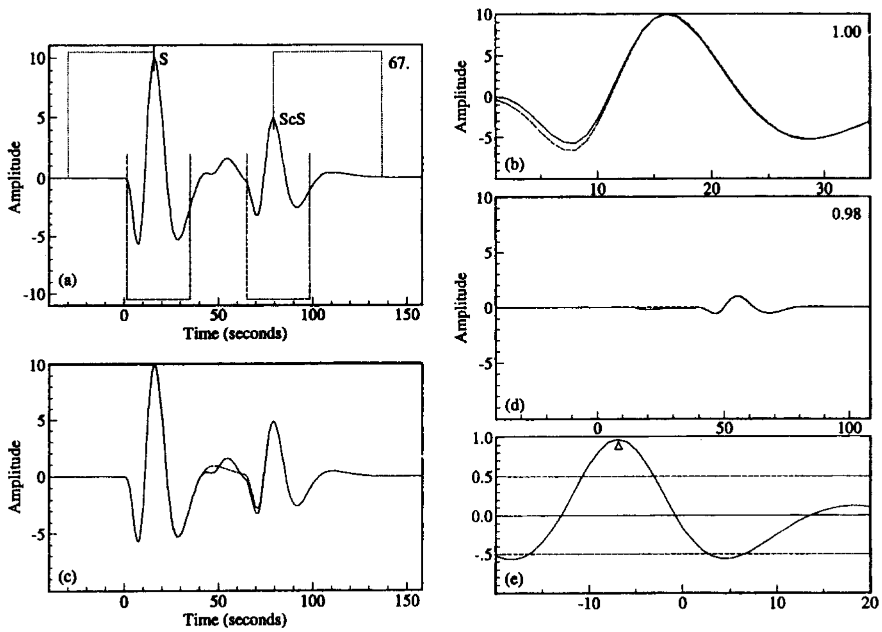

Fig. 1. An illustration of the phase-stripping technique applied to the synthetic SH waveform for the model SGLE (Gaherty and Lay, 1992) at an epicentral distance of 67°. This model has a D″ discontinuity 290 km above the CMB. The S and the ScS pulses (a) are windowed, correlated and displayed in (b). The cross-correlation coefficient is shown in the upper right of (b). The front portion of the S pulse and the rear portion of the ScS pulse (shown in the upper half of (a)) are scaled and joined to form the reference pulse; (c) shows the original waveform and the reference trace. The reference trace is formed by the superposition of reference pulses centred and scaled on the picked S and ScS arrivals. The reference trace is subtracted from the original waveform and the result is displayed in (d). The resultant SdS waveform is cross-correlated with the reference pulse, and the dashed line is the reference pulse positioned where the cross-correlation is best (coefficient is shown in the upper right of (d)). (e) shows the cross-correlation function for the reference pulse and the resultant waveform. The time window is 20 s either side of midway between the S and ScS picks. The triangle indicates the time shift which optimizes the cross-correlation.

sive result. For a more detailed explanation of the method and some examples of its execution, the reader is referred to KS94.

The phase-stripping technique has been applied to long-period GDSN SH-wave data from 1977 to 1987. This data set is composed of nearly 4000 seismograms that have an epicentral range between 63° and 74°. We used a presorted subset of this data that was compiled by Woodward and Masters (1991) for a lower-mantle tomography study. In this subset, 176 seismograms have bounce points in the region of study, and of those, 43 showed evidence of a D″ reflection. D″ thicknesses are then estimated from the SdS–S and ScS–SdS residuals. Results for the global data set are presented in KS94; in this work we will concentrate on the southwest Pacific, southeastern Asia and Australasian regions.

As a whole, the results of previous studies and those using the phase-stripping method suggest that the lowermost mantle beneath this region is highly variable in lateral velocity structure. Fig. 2 is a summary of the results, showing regions where the top of D″ appears to be a discontinuity and reflects seismic energy, regions where there is little evidence for a D″ reflection and regions where the presence of a D″ reflection is inconclusive. In general, the results suggest that D″ is thin

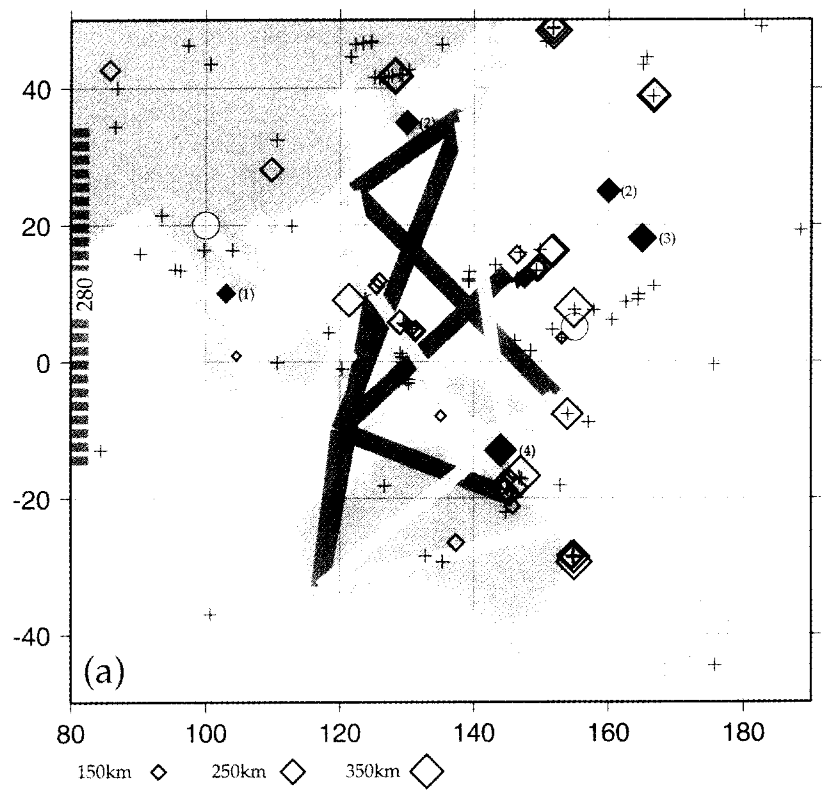

Fig. 2. Interpreted D″ thicknesses plotted at mid-point projections on the surface of the Earth. ♦, Locations where previous studies have identified D″ reflections: (1) Wright et al. (1985); (2) Weber and Körnig (1992); (3) Garnero et al. (1993); (4) Neuberg and Wahr (1991). ◇, Positive results from the phase-stripping results. The size of the diamonds is proportional to D″ thickness. +, Cases where it is not clear from the phase-stripping method if a D″ discontinuity exists. The shaded circles are regions where Weber and Körnig (1992) found no evidence of D″ reflection. The stippled region to the west is the edge of the region where Young and Lay (1987) found evidence of a D″ discontinuity. The shaded corridors are regions studied by Revenaugh and Jordan (1991). The darker corridors are those where they identified a D″ reflection (thicknesses are shown) and the lighter ones are regions where there was little evidence for a D″ reflection.

through a central region and thickens moving westward and eastward. There is a cluster of thin predictions around 5° and 130° (with one exception near 120°E), whereas there are thicker predictions beneath the eastern part of Australia (here the results of Neuberg and Wahr (1991) agree well with the phase-stripping estimates). Although isolated, the Wright et al. (1985) estimate (10°N and 110°E) is not inconsistent with this. In the vicinity of 15°N and 155°E the phase stripping results and those of Weber and Körnig (1992) and Garnero et al. (1993) are in agreement and the thickening trend eastward is fairly strong. There are few data moving to the west of the area, except for those of Young and Lay (1987) which, for a fairly large number of seismograms, predicted a 280 km thick D″ region. Comparing results from P- and S-wave studies could be a little misleading, as P- and S-wave anomalies do not always coincide (Wysession et al., 1992; Weber, 1993). It is conceivable that some of the observed mismatch could be due to comparing P-wave study results with S-wave study results. Furthermore, Bolton and Masters (1992) have shown that Poisson's ratio changes in the lowermost mantle, suggesting a chemical change within the D″ region.

Upon closer inspection, there are some regions where there seems to be consistently an absence of a D″ discontinuity or at least no conclusive evidence of one. For example, the phase-stripping results and those of Weber and Körnig (1992) for the region around 20°N and 100°E show little evidence of a reflector. Weber and Körnig (1992) showed that the region beneath the Caroline and Marshall Islands (roughly 5°N and 155°E) did not indicate a D″ reflector. There is a strong cluster of phase-stripping results which could not iden-

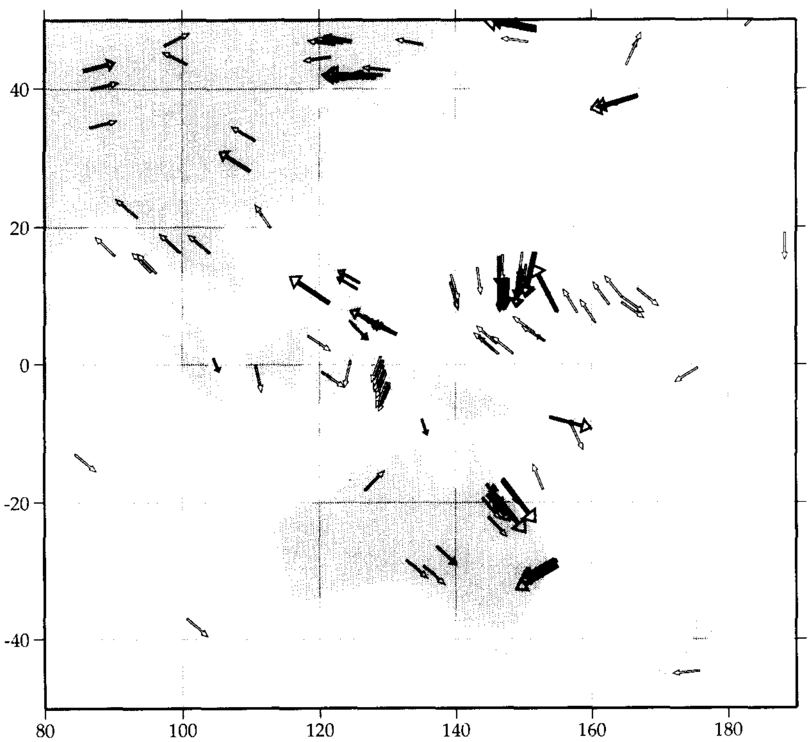

Fig. 3. Source–receiver azimuths plotted at bounce point projections for the phase-stripping results. Points of interpreted D″ reflections are denoted with heavy arrows, with the length of the arrows proportional to the interpreted thicknesses. The light arrows mark spots where a D″ reflection could not be identified or dismissed. Arrows point towards the receivers, with the base of the arrow at the midpoint projection.

tify a SdS phase in the same area, with one notable exception immediately to the north. The region around the Philippines and Indonesia show either inconclusive evidence of a D″ reflection or very thin thickness estimates. It is possible that the SdS signal is so close to the ScS phase that it is interfering with the tail of the ScS pulse and the phase-stripping technique is breaking down. It is important to note that there are many regions where the phase-stripping results identified a D″ reflection for some seismograms but not others. Seismograms for which we could not see an SdS signal do not mean that there is no D″ discontinuity, only that we cannot identify one.

Ideally, regions would be sampled with good azimuthal coverage in the direction of ray propagation. Fig. 3 shows the source–receiver azimuth plotted at the midpoints for the entire data set. Owing to the restricted source and receiver locations with midpoints in this region there is limited azimuthal coverage in many areas and often a D″ thickness estimate comes from repeated earthquakes recorded at the same station (e.g. SE Australia). There are, though, some regions where the data coverage has a few different azimuthal orientations.

In KS94, we ruled out many alternative explanations for the SdS signal, such as noise, contamination from other phases and slab effects. One phase we never considered was SKS. Lay and Young (1986) have explored the possibility of SKS contamination between the transverse-component S and ScS arrivals. Based on assumptions of isotropy, they argued that leakage of energy from the sagittal components into the transverse component, either owing to 3D effects or errors in rotations, would give rise to a small effect over a limited epicentral range. The presence of upper-mantle anisotropy could produce large SKS signals on the transverse components, but again over only a small epicentral range. It is possible that a limited number of the phase-stripping picks could be identifying a SKS signal owing to anisotropy. It is not clear, however, that the polarity of such a phase would be the same as that of the SH and ScSH phases. Furthermore, the region of upper-mantle anisotropy is thought to be on the order of less than 200 km (Silver and Chan, 1991). As the delay times are generally less than 1–2 s, the effect on long-period waves (approximately 25 s) will be small. Anisotropy within D″ would not affect our SdS travel-times, as the phase reflects off the top of the the D″ layer. Obviously, the presence of anisotropy would not explain PcP precursors which have been interpreted as D″ reflections (Wright et al., 1985; Neuberg and Wahr, 1991; Weber and Körnig, 1992).

In general, the phase-stripping results agree reasonably well with the results of other studies. An exception are the results of Revenaugh and Jordan (1991), where they estimated D″ thicknesses in corridors connecting various source–receiver combinations. Corridors which show a reflector suggest significant variability in D″ thickness, and, in some cases, overlapping corridors show evidence for and against a D″ reflection. This is perhaps not too surprising considering the seemingly large variations in D″ structure predicted from the other results. The most easterly corridors do not show a D″ reflection (Fig. 2).

Wysession et al. (1994) used ScS–S travel-time residuals to estimate lateral variations in D″ velocities based on ScS–S travel-time residuals and an assumption of a 300 km thick layer. These results indicate low average velocities (up to −3% relative to the Preliminary Reference Earth Model (PREM; Dziewonski and Anderson, 1981)) in a broad region roughly 20°S–20°N and 120–170°E. Where there is coverage, this region seems to be surrounded by areas of higher than average velocity (up to a 5% anomaly in some areas). Many of the predictions of D″ reflections, both via phase-stripping and other observations, lie in this region (Fig. 2). The implication is that, even though there is a sharp increase in velocity at the top of D″, there must be, relative to PREM, an overall decrease in velocity through D″. Perhaps such a velocity reduction could be due to the presence of thin lamellae within D″ (Weber, 1994). Another possibility is that a low-velocity anomaly lies above D″ contributing to the strength of the reflection. Regions of higher than normal velocities must have a more gradual velocity gradient moving into D″ (synthetic waveforms for such cases are shown in the following section). A decrease in velocity at the top of D″ in the slow regions is not likely because synthetics show that a D″ reflection would be of opposite polarity to those of S and ScS. In such cases, the phase-stripping technique would give a negative correlation coefficient, but this effect is not observed. Furthermore, relative to the case of a positive discontinuity, the amplitudes are weaker as the reflection is always precritical.

3. Variations in amplitudes

The strong lateral variations in D″ structure suggested in the previous section should have significant effects on the amplitudes of waveforms. Fig. 4 shows the amplitude ratios, as a function of epicentral distance, for ScS/S and SdS/S. Both sets of ratios show considerable scatter, the implication being that there is significant heterogeneity in the region. The strong lateral variations in D″ topography coupled with strong variations in D″ velocities (as shown by Wysession et al. (1994)) suggest considerable focusing and defocusing effects throughout the region. Some of the amplitude variation could also be due to near-receiver structure.

The amplitudes of the D″ reflection are sensitive not only to the velocity contrast across the discontinuity, but also to the velocity gradient below the discontinuity. Fig. 5 shows variations in the waveforms resulting from changes in the strength of the discontinuity, the underlying velocity gradient and the thickness of the D″ region. The waveforms are modelled using the WKBJ method (Chapman, 1978). The amplitudes of the D″ reflection and ScS are predictably not very sensitive to variations in D″ thickness; travel-times are most diagnostic of such variations (Fig. 5(a)). On the other hand, changes in the magnitude of the D″ velocity discontinuity affect the amplitudes of the SdS reflections rather obviously (Fig. 5(b)). The ScS signal shows little change in amplitude, only changes in arrival time. Fig. 5(c) shows the variations in waveforms for negative, zero and positive velocity gradients below the D″ discontinuity. The amplitudes of both the SdS and ScS signals are strongly affected by these variations. Although in the tested cases the travel-times are most sensitive to the thickness of the D″ layer, they are also somewhat sensitive to changes in the velocity structure within the layer. This poses a problem when trying to invert travel-time residuals for D″ thicknesses. As pointed out by Schlittenhardt et al. (1985), it is important to examine waveforms from a wide range of epicentral distances, even in the diffracted phases past the triplication (greater than 80°). The phase-stripping provides a good first estimate of D″ thickness, but more detailed regional surveys are required to constrain further the velocity gradients within the layer.

The large-scale structure of the lower mantle will have an appreciable effect on waveform amplitudes. In Kendall et al. (1992), it was shown that for a number of P- and S-velocity models of the 3D structure of the Earth, amplitudes will be much more sensitive to the laterally varying structure than the travel-times. Fig. 6 shows the varia-

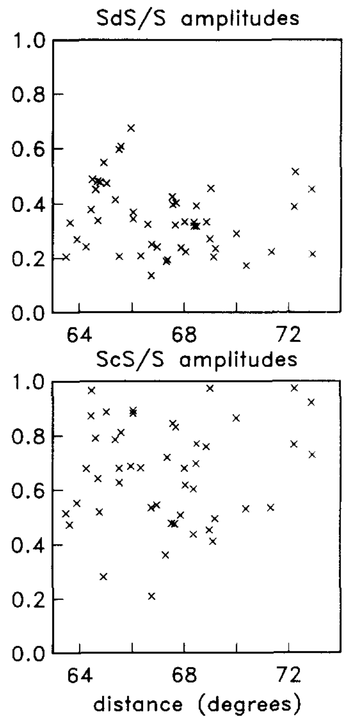

Fig. 4. SdS/S (top) and ScS/S (bottom) amplitude ratios as a function of epicentral distance for the phase-stripping results.

tions in amplitudes for the shear-velocity model WM13 (Woodward et al., 1993). The results are for lower-mantle turning rays from a surface source in the Kuril subduction zone (45°N and 150°E). The variations are most dramatic for the rays turning in the lowermost 300 km of the mantle. For example, amplitudes vary by almost ±30% across Australia. Travel-time corrections for large-scale velocity structure in the lower-mantle above D″ are seldom done in D″ studies,

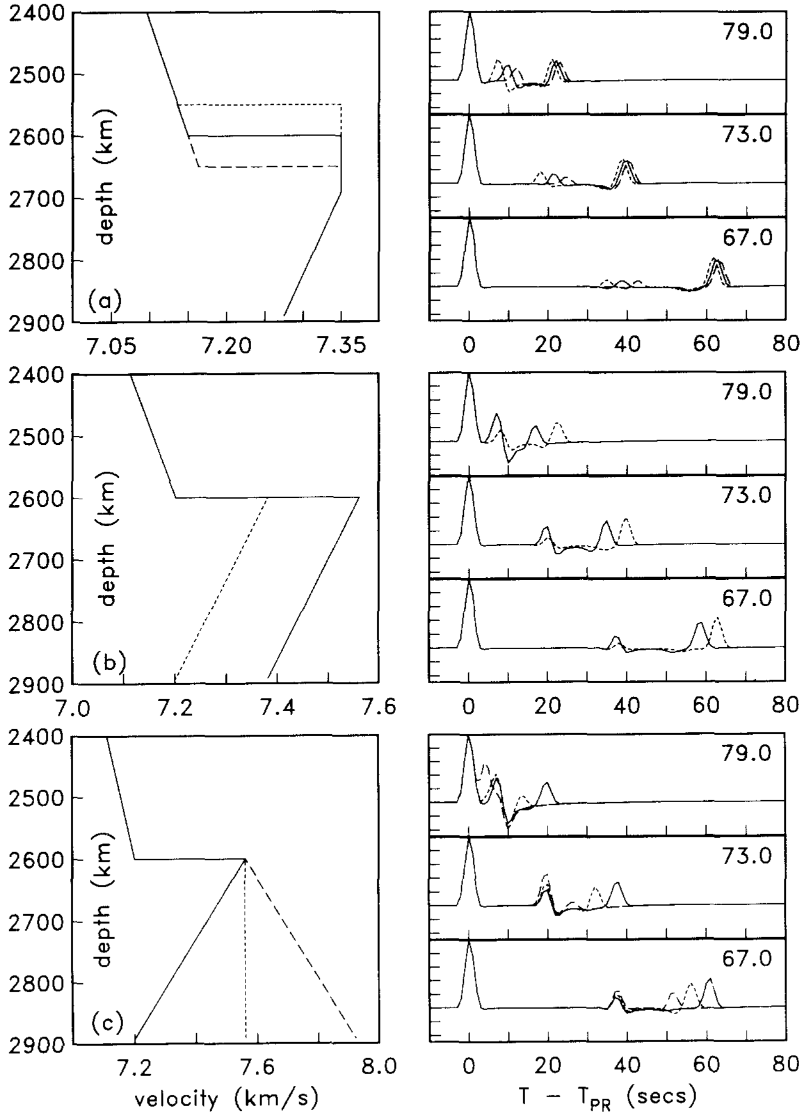

Fig. 5. WKBJ S, SdS and ScS synthetics, given a δ-function at a point source, for a variety of D″ models. Travel-times are reduced by the S-wave travel-times for PREM (Dziewonski and Anderson, 1981). (a) Waveforms for models with a D″ discontinuity 240 km above the CMB (dashed line), 290 km above the CMB (continuous line) and 340 km above the CMB (dotted line). These models are based on the model SGLE (Gaherty and Lay, 1992). (b) Waveforms for models with a D″ discontinuity 290 km above the CMB, a 2.5% decrease in velocity through D″ and a D″ velocity discontinuity of 2.5% (dotted line) and 5% (continuous line). (c) Waveforms for models with a D″ discontinuity 290 km above the CMB, a D″ velocity discontinuity of 5% and a velocity gradient through D″ of −5% (continuous line), 0% (dotted line) and 5% (dashed line).

but Wysession et al. (1994) showed that such corrections reduced the scatter in their data and improved their analysis. Our results suggest that corrections for amplitude variations will be even more crucial. In the context of the phase-stripping approach, this forward modelling should be extended to examine variations in waveform amplitudes for the ScS and SdS signals. More importantly, waveform sensitivity to variations in the topography of the D″ discontinuity should be explored. These are daunting tasks, but the theory and software required to undertake such a project are rapidly emerging (see recent work by Weber (1993) and Neuberg and Pointer (1995)).

4. D″ topography, lower-mantle velocities and flow

In KS94 it was suggested that there is a correlation between D″ thickness and lower-mantle velocities. Fig. 7(a) shows the variation in D″ thickness moving east to west between 5°S and 40°N, and Figs. 7(b) and 7(c) show lower-mantle velocities 260 km above the CMB along an east–west path 15°N for the models WM13 (Woodward et al., 1993) and SH10C (Masters et al., 1992). In general, areas of a thin D″ layer are sites of higher than average velocity and thick D″ estimates are in low-velocity regions. Revenaugh

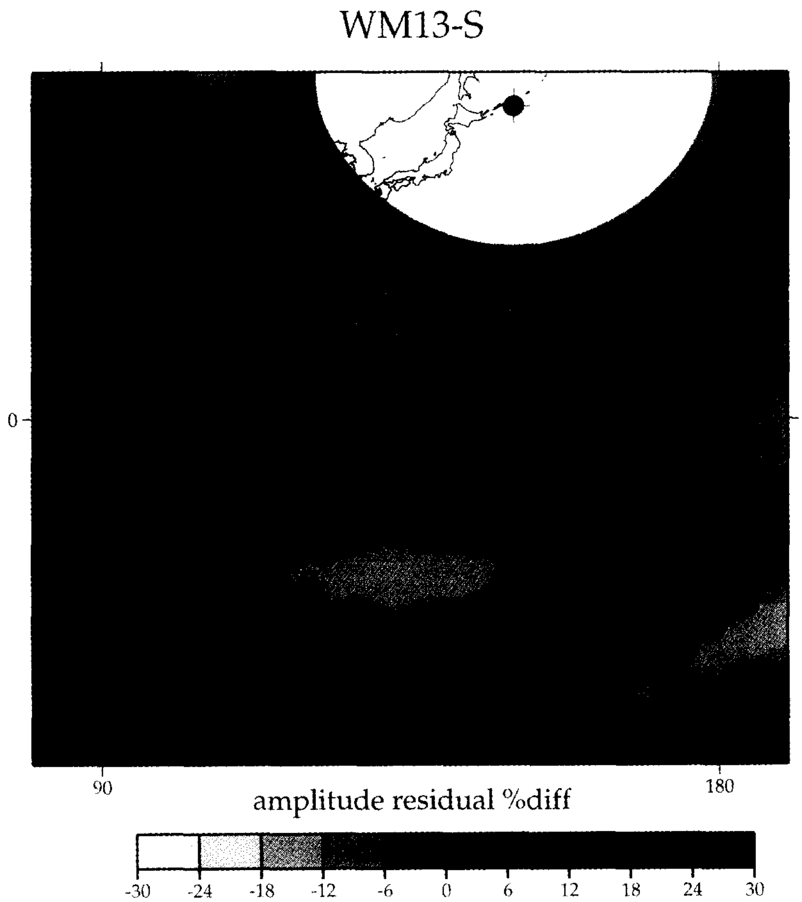

Fig. 6. S-Wave amplitude variations for lower-mantle turning rays from a source in the Kuril subduction region (45°N and 150°E) in the model WM13 (Woodward et al., 1993). Rays which turn in the lowermost mantle are those emerging farthest from the source.

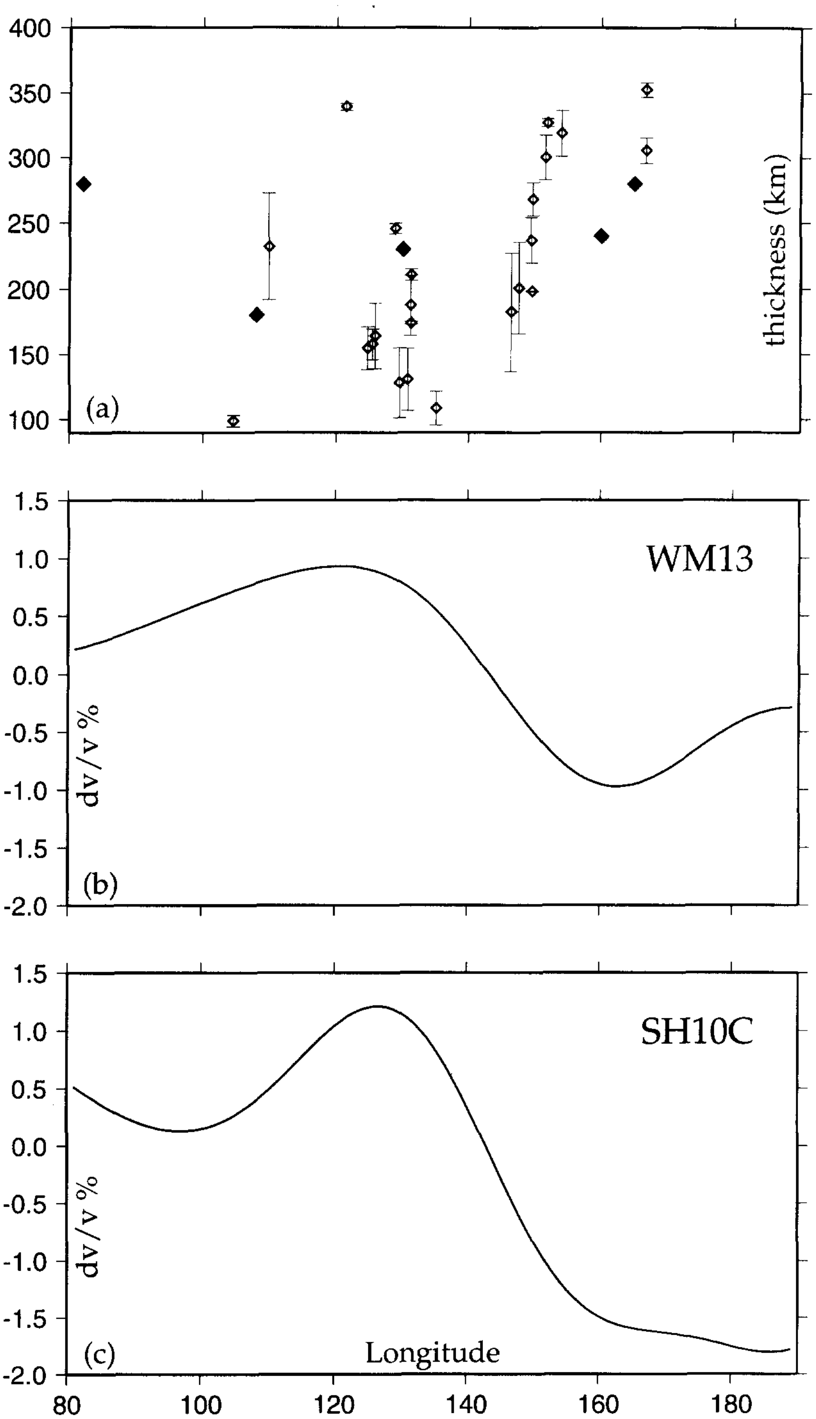

Fig. 7. Comparison of estimated D″ thicknesses and D″ velocity anomalies. (a) Variation in thickness between 5°S and 40°N. ◇ with error bars, estimates from the phase-stripping results; ◆, results of previous studies (see Fig. 2). (b) and (c), Velocity-anomaly profiles along 15°N for the models WM13 (Woodward et al., 1993) and SH10C (Masters et al., 1992), respectively.

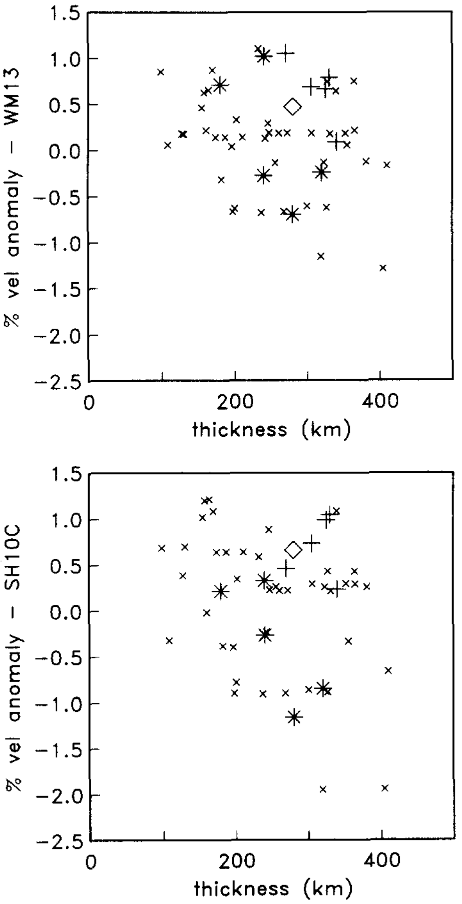

Fig. 8. Correlation between lower-mantle velocity at a depth of 2630 km and estimated D″ thickness for the models WM13 (Woodward et al., 1993) and SH10C (Masters et al., 1992). *, Results from previous studies; +, corridor averaged values from Revenaugh and Jordan (1991); ◇, averaged value for Young and Lay (1987); ×, phase-stripping results.

and Jordan (1991) observed a similar correlation in comparing their observed D″ thicknesses with lowermost mantle velocities for the P-wave model L0256 (Dziewonski, 1984). They suggested that D″ thickens as material is warmed and swept toward sites of mantle upwellings. Fig. 8 shows the correlation between D″ thickness for the entire region and mantle velocities 260 km above the CMB for the models SH10C and WM13. Previous results, including those of Revenaugh and Jordan (1991), are included. Estimates of average velocity anomalies are used for those studies which sampled broad regions (i.e. Young and Lay, 1987; Revenaugh and Jordan, 1991). For example, for Revenaugh and Jordan (1991) we obtain the average velocity using lower-mantle values at five evenly spaced sites along each corridor of investigation. Although the pattern shows more scatter than that obtained by Revenaugh and Jordan (1991) for L0256, there is a general, but weak, trend of thickening as velocities decrease. Furthermore, this trend is not dependent on the choice of lower-mantle velocity model. A second interpretation of the thick but low-velocity D″ region is that these are regions of melt inclusions or lamellae (Weber, 1994). Perhaps the lamellae form as the material warms and migrates horizontally across the CMB.

Seismic tomography models can be used to predict mantle flow fields by assuming a velocity-to-density scaling relationship and a mantle viscosity model (e.g. Hager and Richards, 1989; King and Masters, 1992; Forte et al., 1993). Phipps Morgan and Shearer (1993) recently computed large-scale 3D mantle flow fields which take into account the observed topography on the upper-mantle discontinuities. Fig. 9 shows the predicted flow in the lowermost 300 km of the mantle based on the laterally varying Earth model WM13 (Woodward et al., 1993). Central Australia and southeast Asia are sites of downwelling mantle material. Upon reaching the bottom of the mantle, this material flows horizontally east and west, warming as it approaches sites of mantle upwelling.

A qualitative comparison of Figs. 2 and 7 suggests that in general the thickest D″ predictions are in regions of predicted strong horizontal flow, whereas the thinner D″ predictions lie in regions of mantle downwelling. This would suggest that the downwelling cold material either depresses the D″ boundary or is entrained in a thin D″ region. As the material flows horizontally it appears to thicken as it warms. Eventually, the material equilibrates with the surrounding mantle, presumably diminishing the magnitude of the D″ velocity discontinuity. Closer investigation of the correlation of flow velocity and direction with D″ thickness suggests that there is not a simple

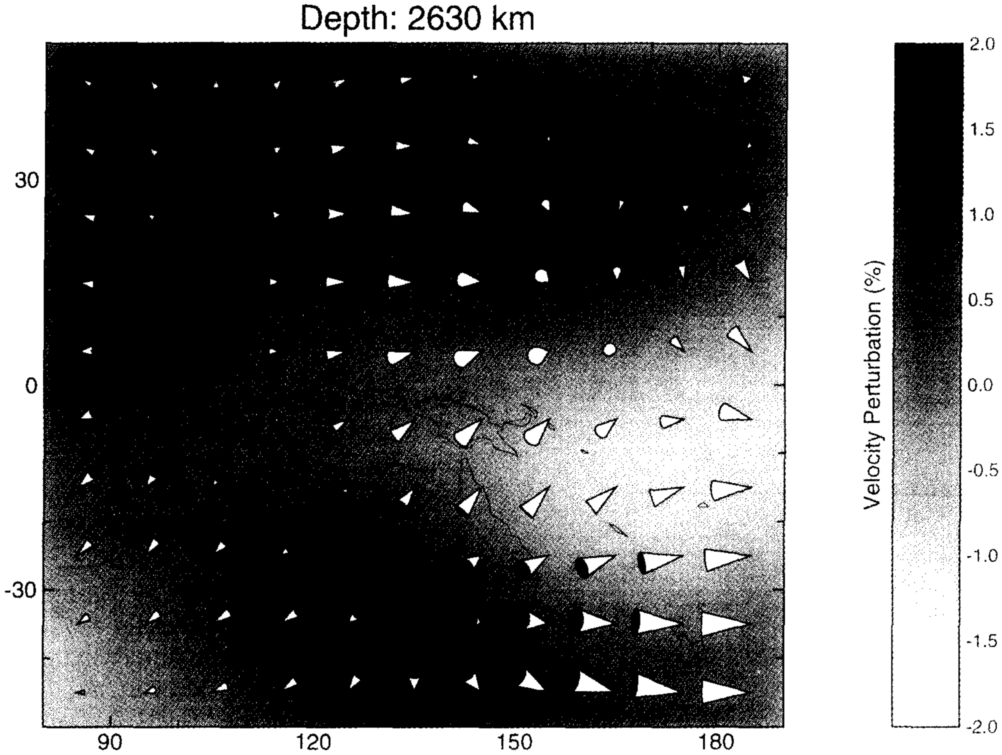

Fig. 9. Map of buoyancy-driven flow patterns in the lowermost mantle based on the model WM13 (Woodward et al., 1993). White cones with black tails display the 3D flow-velocity vectors. The size of the cone is proportional to the magnitude of the flow. Upwards flow appears as a white circle, whereas downwards flow appears as a black circle.

relationship between thickness and flow. A more quantitative interpretation of this correlation is perhaps unwarranted at this point, as the maps of predicted flow are based on the large-scale structure of the lower mantle, whereas the variations in D″ thickness are on much smaller scales.

5. Discussion

A D″ layer which thickens away from sites of convective downwelling is not inconsistent with ideas of D″ being a thermal boundary layer. Jarvis and Peltier (1984) confined temperature variations within the mantle flow to a narrow boundary layer. Their calculations indicate that the temperature profile within D″ could overshoot to colder temperatures at the top of the layer. Beneath this feature is the steady increase in temperature that the boundary layer must accommodate. They also show how a thermal anomaly within D″ evolves and thickens moving away from the centre of a descending plume (see Fig. 3 of Jarvis and Peltier (1984)). Whether or not this temperature overshoot is large enough to lead effectively to a velocity discontinuity and whether D″ thicknesses in excess of 300 km are consistent with this theory are still to be examined. The apparent finer-scale roughness of D″ and the regions where a D″ reflector appears to be absent may be explained by thermal instabilities and convective rolls (Yuen and Peltier, 1980; Stacey and Loper, 1983). Although the possibility of a chemically distinct D″ region cannot be ruled out, the seismic implications of the thermal boundary layer hypothesis should be explored more thoroughly.

The emerging picture of the velocity structure of the lowermost mantle beneath the southwest Pacific, Australasia and southeast Asia is of a region of considerable complexity. Travel-times of D″ reflections imply large variations in the topography of the discontinuity. The observed strong variations in the amplitudes of both D″ and CMB reflections can be due to lateral variations in the strength of the discontinuities and changes in the velocity gradients below the discontinuities. Detailed waveform modelling which allows such variations is required to understand fully and interpret lower-mantle structure in this region.

Acknowledgements

We thank Guy Masters for the presorted S–ScS GDSN data set, and Jason Phipps Morgan for providing the lower-mantle flow calculations. Michael Weber is acknowledged for his constructive review of the paper. This research was undertaken with the support of an NSERC operating grant (J-M.K.) and NSF grant EAR93-15060 (P.M.S.).

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