Evidence for shear velocity anisotropy in the lowermost mantle beneath the Indian Ocean

Jeroen Ritsema
Seismological Laboratory, California Institute of Technology, Pasadena

Abstract. Teleseismic recordings (Δ > 87°) of a deep earthquake beneath the Banda Sea at stations in Tanzania show a difference in the arrival time of the radial (Ssv) and transverse component (SSH) S wave ranging from 1–3 s. Shear velocity anisotropy in the lowermost mantle beneath the Indian Ocean is the likely cause of this signal because recordings at the same stations of closer-in events (Δ < 80°) in the same source region do not present a comparable differential travel time. For the Banda Sea event, the SSH signals are broader than Ssv signals, suggesting that a discontinuity (or strong vertical gradient) in primarily Vsv marks the sudden onset of transverse isotropy in D" (with a magnitude of 1.4–2.7%) about 350 km above the core-mantle boundary. SKKS coda, S-to-p converted phases at the Moho, and upper mantle heterogeneity beneath the stations obscure the onset of Ssv and complicate wave shapes. It is therefore difficult to evaluate whether general anisotropy needs to be invoked into a model of shear velocity anisotropy.

1. Introduction

Structure characterized by shear velocity anisotropy in the lowermost several hundred kilometers of the mantle (D") provides constraints on convective flow patterns in the deep mantle [e.g., Kendall and Silver, 1998; Lay et al., 1998a]. Shear velocity anisotropy in D" is typically examined using models of transverse isotropy and quantified by the difference between the travel times of radial (Ssv) and transverse (SSH) components S or ScS waves, denoted as Tssv-SSH.

While the global compilation of Dziewonski et al. [1996] shows a symmetric distribution of Tssv-SSH about zero, regional investigations present a systematic geographic pattern. Positive values of Tssv-SSH (up to 5 s) are obtained for sampling regions of D" beneath Alaska [Lay and Young, 1991; Matzel et al., 1996; Garnero and Lay, 1997], the Caribbean [Kendall and Silver, 1996], and the northeastern Pacific [e.g., Vinnik et al., 1995; Ritsema et al., 1998]. In these regions, the shear velocity is relatively high and strong vertical shear velocity gradients have been proposed [e.g., Lay et al., 1998b]. Smaller values of Tssv-SSH, ranging from -2 to +2 s, are observed for the D" region beneath the central Pacific [Pulliam and Sen, 1998; Ritsema et al., 1998; Russel et al., 1998] where, on average, the shear velocity is low.

Here, I present recordings of the August 30, 1994 Banda Sea earthquake at seismic stations in Tanzania which provide evidence for the presence of shear velocity anisotropy in D" beneath the Indian Ocean. These data corroborate previous suggestions that anisotropic structure in relatively high shear velocity D" regions yields positive values of Tssv-SSH and that a strong vertical shear velocity gradient marks its upper boundary [e.g., Lay and Young, 1991; Matzel et al., 1996].

2. Indonesia earthquake recordings

Recordings of deep-focus earthquakes allow for precise measurements of Tssv-SSH. S waveforms from deep earthquakes are not complicated by the interference with surface reflections or affected by upper mantle anisotropy in the earthquake source region.

The 1994–1995 broadband deployment in Tanzania [Nyblade et al., 1996] recorded three deep Indonesian earthquakes with body wave magnitudes larger than 5.7 (Table 1). These events are at identical easterly azimuth from the Tanzania Network and generated seismic shear waves that propagate to stations in Tanzania through the same mantle corridor (Figure 1). S waves generated by events A and B (Δ=71–80°) turn at least 700 km above the CMB. S waves generated by event C (Δ > 87°), on the other hand, propagate through the uppermost regions of D" (∼150–350 km above the CMB) where shear velocities are, on average, 1.5% higher than in the Preliminary Reference Earth Model (PREM) [Dziewonski and Anderson, 1981].

Figure 2 compares the highest quality S wave recordings for events A and C. The SSH and Ssv signals of event A have similar onset times and wave shapes. The S waveform complexity, similar for SSH and Ssv, may be caused by the strong lateral variation of seismic shear

Table 1. Event parameters.

<table>
<thead>
<tr>
<th>Event</th>
<th>Lat.(°S)</th>
<th>Lon.(°E)</th>
<th>Depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>5.8</td>
<td>110.3</td>
<td>643</td>
</tr>
<tr>
<td>B</td>
<td>5.6</td>
<td>110.2</td>
<td>559</td>
</tr>
<tr>
<td>C</td>
<td>7.0</td>
<td>124.2</td>
<td>618</td>
</tr>
</tbody>
</table>

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Paper number 1999GL011037. 0094-8276/00/1999GL011037$05.00
velocity in the crust and upper mantle beneath the East African Rift and Tanzania Craton [Ritsema et al., 1998; Ritsema and Van Heijst, 2000].

The pronounced differences between $S^{SH}$ and $S^{SV}$ waveshapes for event C indicate that $S$ wave propagation is strongly affected by structure in the lowermost mantle. At stations URAM, PUGE, and MITU, $S^{SH}$ is broader than $S^{SV}$ and SKS and the $S^{SH}$ pulse at station MTOR exhibits two distinct peaks. The presence of SKS signal on the transverse component recordings of event C at URAM, PUGE, and MTOR suggest that upper mantle anisotropy may, to some extend, be the cause of $S$ waveform complexity. Models of upper mantle anisotropy described by the fast $S$ wave polarization angle ($\Phi_f$) and the differential travel time between the fast and slow $S$ wave ($\Delta t$) [e.g., Silver, 1996] can account for the SKS splitting at URAM and PUGE (but not MTOR). However, corrections for upper mantle anisotropy using such models do not remove the $S^{SH}$ waveform complexity despite the fact the incidence angles of $S$ and SKS differ by less than $10^\circ$. Low values ($< 0.5$ s) of $\Delta t$ were also obtained by Hill et al. [1996] using SKS data for a worldwide distribution of earthquakes.

Ruling out upper mantle structure, the $S^{SH}$ waveform complexity suggests that $S^{SH}$ is interacting with a lower mantle shear velocity discontinuity or strong vertical gradient [Young and Lay, 1987] while $S^{SV}$ is not. Lay and Young [1991], Mутzel et al. [1996], and Garnier and Lay [1997] showed similar waveform characteristics for $S$ waves propagating through the lowermost mantle beneath Alaska.

3. Modeling of $T_{S^{SV}-S^{SH}}$ times

Figure 3 shows $T_{S^{SV}-S^{SH}}$ measurements obtained by estimating time shifts necessary to match the upswing of the $S^{SH}$ and $S^{SV}$ wave signals. $T_{S^{SV}-S^{SH}}$ values scatter about zero between $71^\circ$ and $80^\circ$ (data from events A and B) while $T_{S^{SV}-S^{SH}}$ is larger than $+2$ s at epicentral distances greater than $87^\circ$ (data from event C). $T_{S^{SV}-S^{SH}}$ predictions for three sets of shear velocity models are also shown. These predictions are determined in a similar manner as the data using $S^{SV}$ and $S^{SH}$ waveform synthetics computed separately for different isotropic 1-D $V_{SV}$ and $V_{SH}$ profiles (Figure 4). These velocity profiles are simple deviations from model SYL1 [Young and Lay, 1987]. $V_{SH}$ in model class I increases discontinuously at a depth of 2010 km (280 km above the CMB). Following Garnier and Lay [1997], shear velocity anisotropy is incorporated by in-
voking smaller (0%, 0.7%, 1.4% and 2.1%) discontinuous jumps of $V_{SV}$. Below this depth, $V_{SH}$ and $V_{SV}$ decrease monotonously with depth. In models of classes II and III, $V_{SH}$ and $V_{SV}$ discontinuities are at depths of 2540 km (350 km above the CMB) and 2491 km (400 km above the CMB), respectively.

$T_{SVV-SSH}$ computed for these models increases systematically with epicentral distance as S waves propagate an increasingly larger distance through anisotropic structure in the lowermost mantle. The increase is strongest for models with the smallest $V_{SV}$ increase at the lower mantle discontinuity (i.e., stronger shear velocity anisotropy). The minimum distance at which $T_{SVV-SSH}$ is larger than zero value decreases when the onset of shear velocity anisotropy is placed shallower in the mantle.

Model class I predictions underestimate the observed values of $T_{SVV-SSH}$ for either magnitude of the $V_{SV}$ jump. This indicates that shear velocity anisotropy must be present at depths shallower than 2610 km in order to match the large values of $T_{SVV-SSH}$ at 88°. Model classes II and III, which invoke $V_{SV}$ and $V_{SH}$ jumps 50 to 100 km higher in the mantle, provide a better match to the data. In order to reproduce the increase of $T_{SVV-SSH}$ with epicentral distance, the jump in $V_{SV}$ is constrained to be about two to three times smaller than the $V_{SH}$ jump. A smaller jump in $V_{SV}$ results in an increase of $T_{SVV-SSH}$ faster than observed. The significant scatter in $T_{SVV-SSH}$ and the short epicentral distance range spanned by the data preclude placing more stringent constraints.

4. Discussion and Conclusions

Various mechanisms underlying the cause of shear velocity anisotropy in D" have been proposed. Kendall and Silver [1998], Lay et al. [1998a], and Karato [1998] provide extensive reviews of these mechanisms. Shear velocity anisotropy can be produced by horizontally laminated structures such as melted former-oceanic crust [Kendall and Silver, 1996], core–mantle boundary reaction products [Pulliam and Sen, 1998], and partial melt associated with the ultra–low velocity zone in the low-
Seismic shear velocity anisotropy beneath the Indian Ocean is characterised by positive values of $T_{3SV-SSH}$, consistent with transverse isotropy with a vertical symmetry axis. Values of $T_{3SV-SSH}$ of 2 s at distances as short as 88° indicate that anisotropic structure is present at least 350 km above the core–mantle boundary. Because of the significant scatter, the magnitude, $\eta = 1 - V_{sv}/V_{sh}$, is uncertain but 1.4% and 2.7% represent lower and upper bounds. The variation of shear velocity anisotropy is also poorly constrained because of the limited epicentral distance span provided by the Tanzania network.

SKKS coda, S-to-p converted phases at the Moho, and upper mantle heterogeneity beneath the stations obscure the onset of S$^{SV}$ and complicate wave shapes. Hence, it is difficult to infer whether azimuthal anisotropy needs to be invoked in the model.

Acknowledgments. All figures were generated with the GMT software of P. Wessel and W. H. F. Smith. The IRIS/DMC provided the Tanzania data. This research is supported by NSF grant EAR9896210. Contribution 8674 of the Division of Geological and Planetary Sciences, California Institute of Technology.

References


J. Ritsema, Seismological Laboratory 252–21, California Institute of Technology, Pasadena, CA 91125, USA.

(Received September 9, 1999; revised January 21, 2000; accepted February 20, 2000.)