Mapping the geometry and geographic distribution of a very low velocity province at the base of the Earth's mantle

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[1] We determine the geometry and geographic distribution of a very low velocity province (VLVP) at the base of the Earth's mantle beneath the South Atlantic and Indian Oceans, based on waveform modeling and travel time analysis of seismic data recorded in Africa. Our data sets consist of SH, ScS, SHdiff, SKS, and SKKS phases recorded at two temporary broadband PASSCAL seismic arrays in Africa, the Tanzania array (1994-1995) and the Kaapvaal array (1997–1999), and differential travel time residuals of the ScS-SH phases recorded at the Global Seismographic Network (GSN). These seismic data constitute reasonably good sampling coverage for determining the geographic boundary of a very low velocity province in the South Atlantic and Indian Oceans. The boundary is well determined in the eastern, western, and southern portions but relatively poorly constrained in the northern portion because of the nature of the seismic data. The VLVP exhibits an "L-shaped" form, changing from a north-south orientation in the South Atlantic Ocean to an east-west orientation in the Indian Ocean. It occupies an area of about 1.8×10^7 km² at the core-mantle boundary and a volume of about 4.9×10^9 km³. At least a 4% uncertainty exists in the area estimate, and a 20% uncertainty exists in the volume estimate. Waveform modeling and travel time analysis suggest that the VLVP has rapidly varying thicknesses from 300 to 0 km, steeply dipping edges, and a linear gradient of shear velocity reduction from -2% (top) to -9% to -12% (bottom) relative to the preliminary reference Earth model, consistent with previous results. These structural and velocity features unambiguously indicate that the VLVP is compositionally distinct. INDEX TERMS: 7207 Seismology: Core and mantle; 7260 Seismology: Theory and modeling; 7203 Seismology: Body wave propagation; KEYWORDS: very low velocity province, VLVP, compositional

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1. Introduction

anomaly, core-mantle boundary

[2] The Earth's core-mantle boundary (CMB) region plays a significant role in mantle convection, core dynamics and Earth evolution [e.g., *Lay et al.*, 1998; *Garnero*, 2000]. For example, the bottom thermal boundary layer was proposed to be the birthplace of mantle plumes [e.g., *Olson et al.*, 1987]. Because of the large density contrast across the CMB, the lowermost mantle may be a repository of compositional anomalies whose densities are intermediate between normal mantle and core materials. Indeed, it has long been hypothesized and debated that the lowermost mantle may be the graveyard of subducted slabs [e.g., *Ruff and Anderson*, 1980; *Kendall and Silver*, 1996; *Wysession*, 1996] or composed of light elements ejected during the formation of the inner core [e.g., *Buffett et al.*, 2000; *Rost and Revenaugh*, 2003].

[3] Recent seismic studies also revealed the presence of a very low velocity province (VLVP), possibly indicating an

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ancient compositional anomaly, in the lowermost mantle beneath the South Atlantic and Indian Oceans [Wen et al., 2001; Wen, 2001, 2002]. The VLVP resides in a broad, lower than average velocity region in the lowermost mantle in seismic tomographic models [Su et al., 1994; Li and Romanowicz, 1996; Masters et al., 1996; Grand et al., 1997] (see an example in Figure 1b). Above, there is a significant low-velocity anomaly in the lower mantle extending at least 1500 km above the CMB beneath southern Africa (see an example in Figure 1c). The VLVP has a maximum thickness of 300 km, steeply dipping edges, anomalously low shear wave velocities linearly decreasing from -2% (top) to -9% to -12% (bottom) and a maximum P velocity reduction of -3% relative to the preliminary reference Earth model (PREM) [Dziewonski and Anderson, 1981]. It has been suggested that these seismic characteristics can best be explained by partial melting driven by a compositional change produced early in the Earth's history and a vertical thermal gradient within the anomaly [Wen et al., 2001]. The presence of such a possibly ancient compositional anomaly provides useful information in understanding the nature of the CMB, the early differentiation of the

earth and the origin of geochemical anomalies. In order to further understand its dynamic consequences and implications, it is important to map out the geographic extent of this very low velocity province for several reasons: (1) It has been noted that the geographic location of the anomaly appears to coincide with the geochemical DUPAL anomaly [Hart, 1984, 1988; Castillo, 1988] observed on some parts of the ocean floor in the South Atlantic and Indian Oceans [Wen, 2001; Wen et al., 2001]. A detailed mapping of the geographic extent of this anomaly would further establish the correlation. (2) An estimate of its volume would be important for geochemical mass balance calculation, as this anomaly may also represent a distinct enriched geochemical reservoir [Wen et al., 2001; Wen, 2001]. (3) A determination of the geographic extent and seismic structure of the anomaly would also help to infer more accurately seismic structures elsewhere (such as the lower mantle), as the seismic effects due to this bottom boundary layer could be predicted and subtracted in future seismic studies.

[4] In this study, we construct the geometry and geographic boundary of the VLVP through waveform modeling and travel time analysis of seismic data. We present the seismic method, data and results, and discuss two competing models in the following sections.

2. Seismic Method

[5] We employ a two-dimensional (2-D) SH hybrid method developed by Wen [2002] for synthetic calculations. We briefly discuss the method here. Readers are referred to Wen [2002] for the details of the method. The 2-D SH hybrid method is a combination of numerical and analytic methods with the numerical method (finite difference) applied in the heterogeneous region only. As a result, the hybrid method can accurately deal with high-frequency 2-D synthetic simulations of seismic wave propagation at large distances in a heterogeneous medium. This method has been adopted to study many seismological problems, such as the diffracted SH phases beneath the South Atlantic Ocean, the central Pacific Ocean and the South Indian Ocean, and resolved many detailed structural and seismic velocity features important for understanding the origin of a seismic anomaly [Wen, 2001; Wen et al., 2001; Wen, 2002].

3. Seismic Data and Coverage

[6] We collect broadband tangential displacements between 70° and 117° recorded at the Tanzania array in eastern Africa and the Kaapvaal array in southern Africa. We select 18 earthquakes (Table 1) occurring in the South American subduction zone and the Fiji subduction zone, based on waveform quality and sampling coverage (Figure 1a). We choose recordings for earthquakes with simple pulse-like source time functions for waveform modeling and travel time analysis. As a result, most of our selected data for waveform modeling are from deep and moderate earthquakes. The use of SH and ScS waveforms and differential travel times minimizes the effect of source mislocation and seismic heterogeneities in the upper and middle mantle, as ScS and SH phases propagate along similar ray paths in the upper or middle mantle (Figure 2). For example, for an earthquake with a source depth of 50 km

and a receiver at 90°, the separation of SH and ScS ray paths is less than 131 km in the upper 1000 km of the Earth and the maximum error of predicting ScS-SH differential travel time will not exceed 0.06 s if the source depth is changed by 20 km. We use the observations from the following earthquakes for waveform modeling by the SH hybrid method: B (98/12/14), C (97/07/20), D (97/11/28), E (97/09/02), F (95/02/08), L (98/09/28), M (97/12/22), N (98/07/16) and O (97/09/04) (bold events in Table 1; see also Figure 1a). The last four deep Fiji events (labeled L, M, N, and O in Figure 1a) provide good sampling coverage in the Indian Ocean. Because they have been modeled in detail in previous work [Wen, 2001], we directly use the results here. Events A to J provide good sampling coverage for the lower mantle in the South Atlantic Ocean and will be studied in detail in this paper. For some directions not covered by these seismic waves, we connect two nearby boundaries.

[7] We use travel time to constrain further the seismic anomaly. We search the seismic events recorded at the two African arrays during their lifetimes and the GSN from 1997 to 1999. SKS, SKKS, and ScS-SH travel times significantly improve the seismic coverage in constraining the geographic boundary, especially in the regions beneath central Africa, southeastern Africa, and the southern edge of the anomaly beneath the South Atlantic and Indian Oceans (Figure 1b). SKS and SKKS travel time residuals (and waveforms) provide no resolution in constraining the vertical extent of a seismic structure as these phases propagate steeply in the mantle. However, they are useful to constrain the lateral boundary of the anomaly when seismic phases (ScS, SKS or SKKS) sample across the edge and show a rapid travel time residual variation across a relatively small distance. With additional constraints from ScS, SKS and SKKS travel time residuals, the boundary is well determined in its southern, eastern and western portions (Figure 1a). The northern portion, however, is still relatively poorly sampled. We only find six events: G (95/04/17), H (98/ 03/21), I (97/10/13), P (97/05/13), Q (97/12/05) and R (99/ 04/08), that can be used to constrain the northern boundary. The first three events provide ScS and SKS travel time residual information, which is useful for estimating the boundary of the anomaly beneath northern Africa. Events Q and R place a tight constraint on the boundary beneath southeastern Africa, as their SKS and SKKS phases sample across the boundary. Event P exhibits no ScS travel time residuals, placing bounds on the northern extent of the boundary. This estimation will cause a 4% uncertainty in determining the surface area. We discuss in detail the seismic data and modeling results in the next section.

4. Detailed Modeling Results

[8] We present seismic observations, detailed waveform modeling results and travel time analysis of the data for each event. Our goal is to determine the geographic boundary and, when possible, the geometry and velocity structure of the VLVP along the sampling paths of each event. We also make significant efforts to estimate uncertainties of boundaries and study the trade-offs of various model parameters. For each event, our approach is to find a best or preferred model that can predict most



Figure 1. (a) The geographic distribution of the VLVP at the core-mantle boundary (black and light blue contour, with light blue contour adopted from a previous study [Wen, 2001], dashed portion being less certain because of the nature of the seismic data), the ScS-SH travel time residuals plotted at the ScS reflected points at the CMB (green triangles) for earthquakes (red stars) recorded by the stations in the Global Seismographic Network (GSN) (red triangles), along with the great circle paths (gray lines) for the seismic data used in constraining the boundary of the VLVP. The boundary (black and light blue contour) and its transitions to the normal mantle (blue lines) are determined from extensive waveform modeling and travel time analysis of the observed SHdiff, SH, ScS, SKS, and SKKS phases for those events represented by black stars. The seismic data used consist of those recorded at the Tanzania and the Kaapvaal arrays (large black solid triangles) for events labeled from A to Q and at the GSN for events labeled 1-9. (b) Blowup of the region with the boundary of the VLVP (black and light blue contour) and travel time residuals of SKS (black open squares), ScS-SH (green open triangles), and SKKS (green open circles) used in constraining the geographic boundary (symbols defined in the top of Figure 1c). The background is a seismic tomographic model in the lowermost mantle [Grand et al., 1997]. (c) A 2-D cross section of Grand's model from event A (black star) to the Kaapvaal seismic array with black solid triangles representing the span of the Kaapvaal seismic stations.

of the waveform characteristics, ScS-SH travel time residuals (or SHdiff travel time residuals) and ScS/SH amplitude ratios observed in the seismic data. We then study the trade-offs of various model parameters and estimate the uncertainties by perturbing the best or

preferred model to the extent that synthetics cannot explain the data. As no noticeable travel time difference is found between the SH and SV direct arrivals for the seismic data sampling the VLVP [*Fouch et al.*, 1999; *Wen*, 2002], we only consider seismic modeling with

Table 1. Event List^a

Event	Origin	Latitude, °N	Longitude, °E	Depth, km
А	97/09/05 03:29:05	-56.26	-27.82	33
В	98/12/14 16:34:32	-38.21	-71.03	138
С	97/07/20 10:23:35	-22.98	-66.30	256
D	97/11/28 23:02:53	-13.74	-68.79	586
Е	97/09/02 12:24:19	3.85	-75.75	199
F	95/02/08 18:40:25	4.16	-76.64	69
G	95/04/17 07:14:35	33.78	-38.60	10
Н	98/03/21 16:45:05	79.89	1.86	10
Ι	97/10/13 13:43:23	36.38	22.07	24
J	98/08/07 01:50:54	-59.59	-150.72	10
Κ	98/01/26 23:16:53	-47.51	165.19	33
L	98/09/28 13:44:31	-8.19	112.41	152
Μ	97/12/22 02:17:50	-5.50	147.87	179
Ν	98/07/16 12:09:30	-11.04	166.16	110
0	97/09/04 04:35:11	-26.57	178.34	625
Р	97/05/13 14:13:46	36.41	70.94	196
Q	97/12/05 19:02:52	53.75	161.75	33
R	99/04/08 13:10:34	42.61	130.35	566
1	98/05/14 13:31:06	-55.37	-28.29	33
2	98/06/07 16:19:20	-31.52	-67.83	113
3	98/06/18 04:25:02	-11.57	-12.89	10
4	98/07/09 14:24:44	38.72	48.51	26
5	98/10/03 01:21:44	-56.63	-25.50	33
6	98/10/03 15:21:05	-56.71	-25.57	33
7	99/03/04 05:47:34	28.34	57.19	33
8	99/03/05 00:43:13	-20.42	-68.90	111
9	99/04/03 06:24:27	-16.66	-72.66	87

^aBold indicates waveform modeling by the SH hybrid method.

isotropic velocity variations. For each event, we present and discuss the seismic observations first and then the modeling results, uncertainties, and trade-offs.

[9] The ScS phases recorded in event A (97/09/05) (Figure 1a) show an increasing travel time delay with respect to model IASP 91 [Kennett and Engdahl, 1991] predictions from 44° to 57°, while the direct SH phases exhibit little travel time delays (Figure 3a). The ScS-SH travel time residuals predicted by model IASP 91 show a linear trend with respect to epicentral distance (Figure 3b). We assume this linear trend would hold for smaller distances and determine the edge of the anomaly by simply linearly extrapolating the trend to a smaller distance (Figure 3b). The onset of the seismic anomaly is assumed to be at the ScS phase bouncing point at the CMB at the epicentral distance (37°, Figure 1b) where extrapolated ScS-SH travel time residual is zero. Note that we do not have enough constraints on the thickness and velocity reduction of the low-velocity anomaly. Simmons and Grand [2002] studied the same event and suggested that the probable thickness for this portion of the low-velocity anomaly was in the range of 150-300 km based on travel time modeling. Also, they ruled out an internal ultralow-velocity zone (ULVZ) with thickness less than 10 km beneath the South Atlantic Ocean because no significant precursors were observed in the data. The corresponding average shear velocity reductions were -9% for a thickness of 150 km and -5% for a thickness of 300 km. Their 300-km-thick model is remarkably consistent with our modeling results for thickness and velocity reductions from other portions of the anomaly. We adopt a thickness of 300 km in this study. The uncertainty in estimating the lateral boundary

is about 400 km, depending on how one extrapolates the observed travel time trend to closer distances.

[10] Event B (98/12/14) samples the anomaly slightly north of the area sampled by event A (Figure 1a). Clear direct SH and ScS phases are observed in the tangential displacement records (Figure 4a). It is evident that the direct SH phases are not delayed until 87.5° in the records, indicating that the seismic anomaly should be confined in the bottom 420 km or less in the lowermost mantle. The direct SH waves sampling the lower mantle are dramatically different from those propagating beneath southern Africa from the South Sandwich region to the Tanzania array [Ritsema et al., 1998]. This indicates that the seismic anomaly in the middle-lower mantle is geographically localized beneath southern Africa. The ScS phases are largely delayed relative to the PREM predictions (see synthetics for PREM in Figure 4c) and the ScS delays increase from about -0.7 s at 70.5° to about 7.7 s at 87.5° . There is about a 1.0 s error in the hand-picked travel times, so we assume that the ScS phases start to be significantly delayed after 70.5°. Note that there exist additional pulses between the observed ScS and SH phases. One possible explanation is that they represent a triplicated phase from the D" structure. However, similar pulses also appear after the observed ScS phases (Figure 4a). Therefore we only model the first-order differential features of ScS and S phases and make no further effort to explain the pulses appearing between the SH and ScS phases as a triplicated phase from the lowermost mantle. The observed ScS-SH differential travel times are unlikely to be caused by the heterogeneities near the earthquake source or in the upper mantle and the crust, as the ScS and SH phases propagate along similar ray paths in the shallow earth. Instead, they are caused by a localized low-velocity structure at the base of the mantle. Figure 4b represents seismic synthetics for our preferred model: A basal layer located 4600 km away from the source, with a thickness of 300 km, a steeply dipping edge, and a linear gradient of shear velocity reduction from 0% (top) to -10% (bottom) (see model in the bottom of Figure 4b). The preferred model can predict both the observed ScS-SH travel time residuals and ScS/SH amplitude ratios. We choose a thickness of 300 km in our preferred model to be consistent with the modeling results from events D and E discussed later. The interpretation of the observations, however, is nonunique. There exist



Figure 2. Ray paths of direct SH, ScS, SKS, SKKS, and Sdiff phases based on PREM at epicentral distances from 80° to 110° .



Figure 3. (a) Observed tangential displacements for event A (97/09/05). Seismic waveforms are aligned along the predicted SH arrivals based on IASP 91. Note that while the direct SH arrivals are not delayed in the whole distance range, ScS phases are largely delayed from 3.2 s at 44° to 12 s at 57° (see also Figure 3b). (b) ScS-SH travel time residuals with respect to IASP 91 for event A (97/09/05) as a function of epicenter distance. Solid line represents the least squares fit to the travel time residuals.

significant trade-offs between model geometry, location, thickness and detailed velocity structure in explaining the ScS-SH observations before the diffracted distances [*Wen*, 2002]. We test many models in order to estimate the trade-offs between model location, thickness, dipping edge, and velocity reduction in explaining the same observations. With all other parameters kept the same, models with locations ranging from 4400 km for a thickness of 250 km to 5080 km for a thickness of 420 km can explain the observations (see Figure A1 for some examples).

[11] In event C, the observed ScS tangential displacements are increasingly delayed from about 4.9 s at 77.5° to about 12.0 s at 86.5° while the direct SH phases are not (Figure 5a compare to PREM synthetics in Figure 5c). Similar to the observations for event B, there are additional pulses between the observed ScS and SH phases and after the observed ScS phases (Figure 5a). For the same reason, we only model the first-order differential features of ScS and S phases and make no further effort to explain the pulses between the direct SH and ScS phases. The turning depth for the direct SH wave observed at 86.5° is about 420 km above the CMB. We adopt a thickness of 300 km, a value obtained from waveform modeling of the observations of events D (97/11/28) and E (97/09/02). The observed SH-ScS waveforms are well-explained by our preferred model in Figure 5b, which is located 4400 km from the corresponding source and has a steeply dipping edge and a linear gradient of shear velocity reduction from 0% (top) to -10% (bottom). Here, as in the case of event B, the quality of observations and the epicentral distance range from this event prevent us from resolving the boundary geometry and velocity gradient in good confidence (see Figure A2 for some examples). Models with locations ranging from 4000 km for a thickness of 200 km to 4600 km for a thickness of 350 km can explain the observations, with other parameters kept the same.

[12] Event D (97/11/28) has been modeled in detail in previous work [*Wen et al.*, 2001]. The seismic waveforms observed for this event have also been used extensively by *Ni et al.* [2002] and *Ni and Helmberger* [2003b]. They invoked a model with a thickness of 300 km and a uniform velocity reduction of -3%. However, *Wen et al.* [2001] and *Wen* [2002] modeled the same event and pointed out that the best fitting model had a strong negative velocity gradient and there existed significant trade-offs between velocity reduction and geometry. In this paper, we review the seismic data and test more models. It is evident that the direct SH phases begin to arrive late only at distances larger than 88°, when they start to sample the lowermost 300 km of the Earth's mantle, while the ScS phases are delayed throughout



Figure 4. (a) Observed tangential displacements for event B (98/12/14). Note that the direct SH phases are not delayed in the whole distance range, while ScS phases are. (b) Seismic synthetics for our preferred model: a 300 km thick layer with a steeply dipping edge and a linear negative shear velocity gradient from 0% (top) to -10% (bottom), located 4600 km away from the source (bottom of the panel). (c) Seismic synthetics predicted by PREM. The theoretical ScS arrival times predicted by IASP 91 are indicated by the dashed lines labeled ScS. The phases labeled 0 are truncation phases of the hybrid method calculation.

the whole distance range (compare Figure 6a to Figure 6f for PREM synthetics). A model, consisting of a basal layer located 4900 km away from the source, with a thickness of 300 km, a steeply dipping edge and a linear gradient of velocity reduction from -2% (top) to -12% (bottom), can explain the observations well (Figure 6b). The synthetics produced by this model match the observed SH arrivals, ScS arrivals, and the distance dependence of the ScS/SH amplitude ratios and general waveform characteristics. Even in this prediffracted distance range, we find trade-offs between velocity reduction, location and geometry in predicting the ScS-SH travel times (Figures 6c and 6d). A basal layer located 4800 km from the source with a linear gradient of shear velocity reduction from -2% (top) to -8%(bottom) in Figure 6c, and a basal layer located 4300 km away from the source with a linear gradient of velocity reduction from 0% (top) to -5% (bottom) in Figure 6d, make no difference in predicting the ScS-SH travel time residuals. Note that the observed rapid increase of ScS/SH ratios from 83° to 95° favors models with large negative

shear velocity gradients (Figure 6b). Note also that we still need a steeply dipping edge to explain the observed seismic waveforms. Synthetics for models with a uniform -3%shear velocity reduction show small time separation between ScS and S phases at large distances and strong ScS phases at close distances regardless of the geometry and location, different from the observations (see Figure 6e for an example). The predicted small ScS-SH time separations at large distances may be compromised by introducing some low-velocity anomaly in the middle-lower mantle [Ni et al., 2002; Ni and Helmberger, 2003b]. However, even if we assume some low-velocity anomaly in the middle-lower mantle, the observed small ScS/SH amplitude ratios in the close distances and the rapid increase of ScS/SH amplitude ratios for this event cannot be reconciled with models with a uniform -3% shear velocity reduction (see an example in Figure 6e). Note that the waveform complexities observed between 92° and 95° are not accounted for by any of the models in Figures 6b-6e. These observed complexities are well above the noise level of the data and they are likely



Figure 5. (a) Observed tangential displacements for event C (97/07/20). Note that the waveform features are similar to those in event B (Figure 4a): direct SH phases are not delayed between 77.5° and 87.5°, while ScS phases are. (b) Seismic synthetics calculated using our preferred model: a 300 km thick layer with a steeply dipping edge and a linear negative shear velocity gradient from 0% (top) to -10% (bottom), located 4400 km away from the source (bottom of the panel). (c) Seismic synthetics predicted by PREM. The theoretical ScS arrival times predicted by IASP 91 are indicated by the dashed lines labeled ScS. The phases labeled 0 are truncation phases of the hybrid method calculation.

related to the seismic structures in the lowermost mantle. One possible explanation is that the edge of the structure is not a simple sloping transition as in our models and involves multiple steps of transition. As significant trade-offs (see Figures 6b–6e) already exist for explaining the ScS-SH data, we do not attempt to try more complex models. Rather, we emphasize the first-order features, such as ScS-SH differential travel times, ScS/SH amplitude ratios and distance dependence of ScS/SH amplitude ratios, and trade-offs.

[13] Wen et al. [2001] and Wen [2002] pointed out that unlike those recorded in the prediffracted distances, the broadband SH phases observed in the diffracted distance range placed extremely tight constraints on model parameters. SHdiff phases propagate along the CMB and they are very sensitive to the seismic structure near the base of the mantle [Wen, 2002]. Figure 7a shows the tangential displacements recorded in the northern part of the Kaapvaal array for event E (97/09/02) (compare PREM synthetics in Figure 7f). The observed tangential displacements for event E exhibit a rapid change of waveform characteristics and

can be divided into two groups based on their waveform complexities [Wen et al., 2001]. Wen et al. [2001] and Wen [2002] have discussed in detail these waveform characteristics, possible interpretations of these waveform features and the simplicity of the earthquake source, and tested hundreds of models in explaining the observed travel time delays, waveform features and the rapid change of the waveform complexities. The observed travel times and waveform features for this seismic event place extremely tight constraints on model thickness, geometry and detailed seismic velocity structure. A 300 km basal layer, located 5800 km away from the source with a steeply dipping edge and a linear gradient of shear velocity reduction from -2%(top) to -12% (bottom), explains the observed linearly increasing travel time delays and the triplicated phases (phases labeled as 2 and 3 in Figure 7a). Readers are referred to these two papers for a more detailed discussion. Here, we present the seismic observations recorded in the northern part of the array only and extensively test models with a uniform shear velocity reduction of -3% with various model geometries and locations. These observations

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Figure 6. (a) Observed tangential displacements for event D (97/11/28). (b) Seismic synthetics for the best fitting model: a 300 km thick layer with a steeply dipping edge and a linear negative shear velocity gradient from -2% (top) to -12% (bottom), located 4900 km away from the source. Seismic synthetics calculated using various velocity gradients and different dipping edges, (c) with a location at 4800 km and a linear velocity gradient from -2% (top) to -8% (bottom), (d) with a location at 4300 km and a linear velocity gradient from 0% (top) to -5% (bottom), and (e) with a location at 4200 km and a uniform velocity reduction of -3%. Note all synthetics (Figures 6b–6e) can explain the differential ScS-SH travel times as well, indicating the trade-offs between location, velocity reduction, and dipping edge in predicting the ScS-SH travel times. The rapid increase of ScS amplitudes from 83° to 95° observed in the data (Figure 6a) favors models with a large negative velocity gradient (e.g., Figure 6b). (f) Seismic synthetics predicted by PREM shown for reference. All the seismic waveforms from Figures 6a–6f are aligned by the SH arrivals predicted by IASP 91. The theoretical ScS arrival times predicted by IASP 91 are indicated by the dashed lines labeled ScS. The phases labeled 0 are truncation phases of the hybrid method calculation.



Figure 7. (a) Observed tangential displacements for event E (97/09/02). (b) Seismic synthetics for the best fitting model: a 300 km thick layer with a steeply dipping edge and a linear negative shear velocity gradient from -2% (top) to -12% (bottom), located 5800 km away from the source. Synthetic examples calculated using models with a -3% uniform shear velocity reduction and various dipping edges, (c) with a location at 6500 km and a steeply dipping edge, (d) with a location at 5700 km and a shallowly dipping edge, and (e) with a location at 7300 km and an inward dipping geometry. Models are shown in the bottom of the panels. Note that the synthetics in Figures 7c-7e cannot explain the seismic data. The travel time and the triplicated phases (labeled as 2 and 3) place very tight constraints on the geometry, velocity gradients, and thickness of the best fitting model in Figure 7b. (f) PREM synthetics are shown for reference. All the seismic waveforms from Figures 7a-7f are aligned by the SH arrivals predicted by IASP 91. The phases labeled 0 are truncation phases of the hybrid method calculation.



Figure 8. (a) Observed tangential displacements for event F (95/02/08). (b) Seismic synthetics calculated by our preferred model: a 300 km thick layer with a steeply dipping edge and a linear negative shear velocity gradient from 0% (top) to -10% (bottom), located 6250 km away from the source. The linear line shows the observed SHdiff travel time in the data (see bottom of the panel for the model). All seismic waveforms are aligned along the predicted SHdiff arrivals based on IASP 91. The phase labeled 0 is a truncation phase of the hybrid method calculation.

cannot be explained by models with a uniform shear velocity reduction of -3%, regardless of the geometry and location. Such models always fail to generate the triplicated phases (phases labeled as 2 and 3 in Figure 7a) observed in the data (see Figures 7c–7e for examples).

[14] We select the observed SHdiff waveforms recorded at the Tanzania array from event F (95/02/08) to constrain the boundary further north. The waveforms do not appear triplicated in the epicentral distance range. However, large travel time delays of SHdiff phase are observed (Figure 8a). Synthetics for a 300 km thick model, located 6250 km away from the source with a linear gradient of shear velocity decrease from 0% (top) to -10% (bottom) and a steeply dipping edge, explain the observations well. In the absence of triplicated phases, which depend on the relative position of the anomaly to the earthquake source, trade-offs exist among the model location, dipping edge and thickness in explaining the data (see Figure A3 for examples). Models with locations ranging from 5900 km for a thickness of 275 km to 6400 km for a thickness of 328 km can explain the observations, with other parameters kept the same.

[15] The constraints on the northern boundary of the VLVP become problematic, as no data at large distances are available in this direction, and we have to rely on the seismic ScS-SH observations in close distance ranges. Three events, G, I, and H, provide some constraints on the boundary locations in the north (Figure 1a). The ScS or Sdiff phases for these events sample either outside the anomaly or inside it. So we model these events only with 1-D models to place some bounds on the approximate boundary location and velocity reduction. We assume these observations are affected only by the seismic structures in the lowermost 300 km of the mantle, as we have no additional constraints on model thickness. The observations for event G (95/04/17) show no significant travel time delays (Figure 9a) and can be explained with a linear negative shear velocity gradient of -3%



Figure 9. (a) The observed tangential displacements for event G (95/04/17) and (b) seismic synthetics calculated using a 300 km thick basal layer with a linear shear velocity gradient of -3% with respect to PREM. Note that synthetics explain the observations well. Because a linear negative shear velocity gradient of -3% may represent normal mantle in a hot region, we interpret that the ScS phases sample outside the anomalous layer. All the seismic waveforms are aligned by the SH arrivals predicted by IASP 91. The theoretical ScS arrival times predicted by IASP 91 are indicated by the dashed lines labeled ScS.

(Figure 9b). The ScS phases for event I (97/10/13) exhibit large travel time delays (Figure 10a) and can be explained by a linear negative velocity gradient of -6%(Figure 10b). The Sdiff phases for event H (98/03/21), which propagate along the same great circle paths as the ScS phases for event I, show no travel time delays until 105°. A linear negative shear velocity gradient of -3%found in event G may reflect either a reduced thickness of the VLVP at the ScS bouncing points, or normal mantle in a hot region such as what has been found in the central Pacific [Ritsema et al., 1997; Wysession et al., 1999; Wen, 2002]. As the VLVP ends rapidly in other directions and the data for event G do not show any variations across the array as would be expected for the seismic waves sampling the VLVP edges, we consider a linear negative velocity gradient of -3% a more likely representation of normal mantle in a hot region. Under this assumption, the VLVP ends south of the ScS bouncing points for event G, north of the ScS bouncing points for event I, and south of the Sdiff paths for event H (Figures 1a and 1b).

[16] We use a large number of ScS, SKS and SKKS travel times to further constrain the VLVP boundary. These travel times significantly improve the boundary sampling beneath southeastern Africa, central Africa, the South Atlantic Ocean and the Indian Ocean (Figure 1b). The SKS travel time residuals from event J (98/08/07) vary dramatically

from 0 to 4.8 s over 460 km across the seismic array, placing tight constraints on the boundary beneath the South Atlantic Ocean. The SKKS travel time residuals gathered from event K (98/01/26) change from 0 s for the southern stations to 3.2 s for the northern stations across the Kaapvaal array, providing tight constraints on the boundary edge beneath the South Atlantic Ocean. The SKS travel time residuals for events L, M, N, O, show rapid changes from 0 s to 4.5 s from north of $(32^{\circ}S, 35^{\circ}E)$ to the south (Figure É 1b). These SKS travel time delays are consistent with each other in their SKS exit points at the CMB from event to event. A portion of the boundary beneath eastern Africa is well determined from the ScS travel time residuals from event P (97/05/13) and the SKS and SKKS travel time residuals from events Q (97/12/05) and R (99/04/08). The ScS travel times observed from event P show no delays, limiting the extent of the boundary west and south to the ScS bouncing points. The SKS and SKKS phases from event Q (97/12/05) in the Japan region sample across the boundary and show systematic travel time delays of about 5.5 s and 4.5 s, respectively, placing tight constraints on the boundary in this direction (Figure 1b). The SKS travel time residuals observed for event R (99/04/08) exhibit a rapid variation from 0 s to 3.3 s (Figure 1b), placing a tight constraint on the lateral boundary in the sampling region. The SKKS phases for event R (99/04/08), whose exit points at the CMB are outside of the geographic boundary, are not



Figure 10. (a) The observed tangential displacements for event I (97/10/13) and (b) seismic synthetics calculated using a 300 km thick layer at the bottom of the mantle and a linear decrease of shear velocity from 0 (top) to -6% (bottom) with respect to PREM. The theoretical ScS arrival times predicted by IASP 91 are indicated by the dashed lines labeled ScS. The phase labeled 0 is a truncation phase of 1-D generalized ray theory calculation.

delayed (Figure 1b), well corroborating our boundary determination.

[17] The ScS-SH travel time residuals observed at the GSN stations for events occurring from 1997 to 1999 (events labeled as numeric numbers, Figure 1a) corroborate remarkably well with the VLVP boundary determined from the above waveform modeling and travel time analysis (Figure 1a). The ScS-SH travel time residuals show little travel time delays when the ScS phases sample outside the boundary, and large travel time delays up to 10 s, an amount exactly predicted by our velocity structure inside the VLVP when the ScS phases sample the seismic anomaly (Figures 1a and 11).

[18] We construct the geometry and geographic boundary of the VLVP based on the results obtained from the above waveform modeling and travel time analysis. Note that we can only make an approximate estimation of the boundary beneath northern Africa. The VLVP exhibits an L-shaped form, changing from a north-south orientation in the South Atlantic Ocean to an east-west orientation in the Indian Ocean (Figure 1a). It occupies an area of about 1.8×10^7 km² and a volume of about 4.9×10^9 km³ excluding the portion in the middle-lower mantle beneath southern Africa. The error in the area estimation is about 4% due to the uncertainties in determining the boundary in the northern portion of the anomaly. The thickness has an uncertainty of 50 km, therefore an error of at least 20% exists in the volume estimation. A 3-D view of the anomaly is presented in Figure 12.

5. Discussions

5.1. Magnitude of Shear Velocity Reduction Within the VLVP

[19] There still exists debate concerning the magnitude of shear velocity reduction within the VLVP near the CMB. Most notably, *Ni et al.* [2002] and *Ni and Helmberger* [2003a, 2003b, 2003c] invoked a model with a uniform shear velocity reduction of -3% in the lowermost mantle. *Ni and Helmberger* [2003b, 2003c] even claimed that our models with a large, reduced velocity were not compatible with the data sets of *Ritsema et al.* [1998] and the data sets sampling a 2-D corridor from the east Pacific rise to the South Sandwich islands to Tanzania. We discuss here how these two competing models explain the seismic data along different sampling paths.

[20] We start with the paths from the South American subduction zone to the Kaapvaal seismic array. The primary data set used by *Ni et al.* [2002] and *Ni and Helmberger* [2003b] in deriving their uniform -3% model for these paths was the ScS-SH data recorded for event 97/11/28 (event D above). As noted above, the interpretations of the seismic data of that event are nonunique; there exist significant trade-offs between model geometry and assumed



Figure 11. The tangential displacements selected from the seismic data recorded at the GSN stations (see Figure 1a for sampling coverage). Each trace is aligned according to its direct S phase. Distance corrections are made so that each trace is plotted to a distance corresponding to a common source depth of 27 km. For each trace, distance correction is made by plotting the seismogram at a distance that would predict same ScS-SH differential travel time for a source depth of 27 km as the recorded distance would do for the actual event depth. The triangles represent hand-picked arrival times of SH and ScS phases, and dots indicate predicted ScS arrival time based on IASP 91. The ScS ray paths reflected inside the VLVP are significantly delayed with respect to IASP 91, while those reflected outside are not (see Figure 1a). The dashed line is the theoretical ScS arrival times by IASP 91 for a source depth of 27 km. Event numbers are labeled on the top of each trace in the left.

seismic velocity reductions within the anomaly. The ScS-SH differential travel times in this data set can be explained with or without contributions from an anomaly in the middle-lower mantle, although a model with a strong negative velocity gradient can better explain the small ScS amplitudes in close distance and the rapid increase of ScS amplitudes to larger distance (see previous section for discussion). The tightest constraints on the geometry and velocity structure of the lowermost mantle come from the

broadband SH data in the diffracted distance range (event E above; also see detailed discussions by *Wen et al.* [2001] and *Wen* [2002]). As we show above, models of a uniform -3% shear velocity reduction cannot account for the observed waveform features. Introducing some seismic heterogeneity in the middle-lower mantle would account for some fraction of the observed travel time delays, but it would not explain the triplicated phases and their move outs observed in the seismic data. The observed waveform



Figure 12. Three-dimensional views of the VLVP beneath the South Atlantic Ocean and the Indian Ocean (vertical exaggeration: 5.55), viewed from (a) 340° N, (b) 20° N, (c) 220° N, and (d) 150° N. This 3-D structure is constructed on the basis of the results from waveform modeling and travel time analysis. It has a linear gradient of shear velocity reduction from 0% (top) to -9% to -12% (bottom). The approximate location of the seismic anomaly in the middle-lower mantle is illustrated by red traces [*Wang and Wen*, 2003].

features of event E would require a model with a strong negative velocity gradient (> -9%) and a thickness of at least 250 km [Wen et al., 2001; Wen, 2002]. This set of observations cannot be explained by a small velocity reduction embedded with ultralow-velocity zones, as such models produce strong multiples inside the ultralow-velocity zones, different from the observations [Wen et al., 2001; Wen, 2002]. Wen [2002] also discussed in great detail how the observations in events D and E can be consistently explained by a 300-km-thick basal layer with a strong negative velocity gradient. Numerical tests also indicate that the seismic structures with horizontal length scales of a few hundred kilometers and vertical scales of tens of kilometers, such as those of Wen and Helmberger [1998], are unable to explain the data in events D and E, regardless of the position of the structure and velocity reduction used. The rationale for introducing some anomaly in the middlelower mantle was that the seismic observations sampling the middle-lower mantle show large travel time delays from some South Sandwich events to the Tanzania array [Ritsema et al., 1998]. However, the sampling regions in the middlelower mantle from those South Sandwich events are about

3400 km east of the sampling paths from these South American events (such as event B in this paper). Along the path from South America to the Kaapvaal array, as evident from the direct SH wave recorded in events B, C and D, no direct S waves show travel time delays until they sample the bottom 300 km of the mantle (see Figures 4a, 5a, and 6a). This suggests that there is no evidence that the seismic waves encounter a low-velocity anomaly in the middle-lower mantle along the path from South America to the Kaapvaal array. The seismic observations from other event-station pairs from the South Atlantic Ocean and the South Sandwich islands to the European-Mediterranean area, which sample slightly west of the South Sandwich-Tanzania path, also show no travel time delays for the S waves sampling the middle-lower mantle until those S waves sample the bottom 300 km of the mantle [Wen et al., 2001]. All these observations suggest that the lowvelocity anomaly in the middle-lower mantle is located further east of the sampling points of the S waves from the South American events in the middle-lower mantle.

[21] Seismic velocity reductions in the lowermost mantle beneath the Indian Ocean are remarkably similar to those



Figure A1. Seismic synthetics calculated by various models perturbed from the preferred model (Figure 4b), (a) with a location at 4400 km and a thickness of 250 km, (b) with a location at 4800 km and a thickness of 350 km, (c) with a shallowly dipping edge, and (d) with a uniform velocity decrease of -3% with a location of 4200 km from the source. Models are illustrated at the bottom of each panel. Note both Figures A1a and A1b can also explain the observations. All the seismic waveforms from Figures A1a to A1d are aligned by the SH arrivals predicted by IASP 91. The theoretical ScS arrival times predicted by IASP 91 are indicated by the dashed lines labeled ScS. The phases labeled 0 are truncation phases of the hybrid method calculation.

obtained beneath the South Atlantic Ocean. Large shear velocity reductions up to -10% at the base of the mantle are also clearly evident from the observed Sdiff phases sampling the Indian Ocean [Wen, 2001]. Some observed Sdiff phases exhibit multiple phase interferences with small travel time delays, indicating the seismic structures are confined in thin zones near the CMB with large velocity reductions (see examples in Figures 6c, 6d, 3a, and 3b [Wen, 2001]). Small thicknesses of the seismic structures generate small travel time delays, while large velocity reductions produce strong reverberations within the seismic structures that interfere with each other [Wen, 2001]. In that case, those Sdiff phases propagate along the edges of the VLVP [Wen, 2001], which have small thicknesses and large velocity reductions and act like ULVZs. Some observed SHdiff phases show large travel times while maintaining pulse-like arrivals with triplicated phases, suggesting steeply dipping edges and large negative velocity gradients (see examples in Figures 3b, 3d, 6a, and 6b of Wen [2001]). Readers are referred to Wen [2001] for detailed discussions.

[22] We should also mention that our models with large shear velocity reductions at the bottom 300 km of the

mantle are also compatible with the data sets along the 2-D corridor from the East-Pacific rise to the Sandwich Islands to Tanzania, with an additional anomaly extending higher into the middle-lower mantle just beneath southern Africa [*Wang and Wen*, 2003]. The basal 300 km of our model along this cross section is consistent with the results by *Simmons and Grand* [2002] (see also discussion above for event A). *Ni and Helmberger* [2003b, 2003c] argued that the seismic structure in the lower mantle extends at least 1200 km above the CMB along this cross section, we agree [*Wang and Wen*, 2003].

5.2. Interpretations of the VLVP

[23] It is now generally agreed that the seismic structure at the base of the mantle in this region represents a compositional anomaly. The seismic structure in this region was initially thought to represent a large young thermal plume which is erupting off the CMB [e.g., *Ni et al.*, 2002; *Ni and Helmberger*, 2001]. *Wen et al.* [2001] however, suggested that the steeply dipping edges, large negative shear velocity gradient, and uniqueness of the anomaly indicate that it is compositionally distinct. A subsequent



Figure A2. Seismic synthetics calculated using models perturbed from the preferred model (Figure 5b), (a) with a location at 4000 km and a thickness of 250 km, (b) with a location at 4600 km and a thickness of 350 km, (c) with a shallowly dipping edge, and (d) with a uniform velocity decrease of -3%. Note synthetics in Figures A2a-A2c can also explain the observations. All the seismic waveforms from Figures A2a to A2d are aligned by the SH arrivals predicted by IASP 91. The theoretical ScS arrival times predicted by IASP 91 are indicated by the dashed lines labeled ScS. The phases labeled 0 are truncation phases of the hybrid method calculation.

study revealed that the VLVP also has rapidly varying thicknesses and geometry beneath the Indian Ocean, unambiguously indicating that the VLVP is a compositional anomaly [*Wen*, 2001]. The interpretations of the seismic structure in this region now seem to converge, at least for the portion at the base of the mantle [e.g., *Wen et al.*, 2001; *Wen*, 2001, 2002; *Ni et al.*, 2002; *Ni and Helmberger*, 2003a]. *Wen* [2001] and *Wen* [2002] showed that the large linear negative velocity gradient within the VLVP can be explained by partial melting of a compositional anomaly situated in a thermal boundary layer at the base of the mantle.

5.3. Future Directions

[24] Several important issues remain to be resolved in mapping the seismic structure in this region. (1) The geographical location and the nature of the boundary in the northern part of the VLVP are still uncertain. More new seismic data would be needed for studying that portion of the boundary. (2) The velocity structure, structural characteristics, and geographic extent of the seismic anomaly in the lower mantle are not clear, neither is the relationship between the VLVP in the lowermost mantle and the seismic anomaly in the middle-lower mantle. Future seismic studies and full waveform calculations are needed to address those issues. (3) High-frequency 3-D waveform modeling is needed. All our modeling efforts have been based on 2-D synthetic calculations. Although the determination of the geographic boundary based on such 2-D approaches would not change much, the 3-D effects could potentially be very important when the structure varies rapidly in the direction perpendicular to the ray sampling paths and those inferred structural features are required to be tested and refined using high-frequency 3-D synthetic modeling in the future. To resolve the detailed velocity gradient and geometry using the data such as those for event E, synthetic calculations up to a frequency of at least 0.5 Hz are required.

6. Conclusions

[25] We determine the geometry and geographic boundary of a VLVP at the base of the mantle extending from the South Atlantic Ocean to the Indian Ocean, based on extensive waveform modeling of SH, ScS and SHdiff phases and travel time analyses of ScS-SH, SKS, and SKKS phases. The VLVP has an "L-shaped" form, changing from



Figure A3. Seismic synthetics calculated using models perturbed from the preferred model (Figure 8b), (a) with a location at 5900 km and a thickness of 275 km, (b) with a location at 6400 km and a thickness of 325 km, (c) with a shallowly dipping edge, and (d) with a uniform velocity decrease of -3%. Note synthetics in Figures A3a–A3c can explain the observations, while Figure A3d underpredicts the travel times for Sdiff phases. All seismic waveforms are aligned along the predicted Sdiff arrivals based on IASP 91. The phases labeled 0 are truncation phases of the hybrid method calculation.

a north-south orientation in the South Atlantic Ocean to an east-west orientation in the Indian Ocean. It occupies an area of about 1.8×10^7 km² with an estimate error of 4% at the core-mantle boundary and a volume of about 4.9×10^9 km³ with an estimate error of at least 20%. The geographic boundary is well constrained in the South Atlantic and Indian Oceans and is relatively poorly determined beneath northern Africa. Waveform modeling suggests that the VLVP has rapidly varying thicknesses from 300 km to 0 km, steeply dipping edges, and a linear gradient of shear velocity reduction from -2% (top) to -9% to -12% (bottom), indicating that it is compositionally distinct.

Appendix A

[26] We discuss trade-offs between model parameters in order to explain the observations in events B (98/12/14), C (97/07/20) and F (95/02/08). We test many models in order to estimate variations of model location and thickness in explaining the same observations. In order to investigate the influence of dipping edge on synthetics, we also test many models with various dipping edges and velocity reductions. Different trade-offs exist depending on epicentral distance and data quality.

[27] For event B (98/12/14), trade-offs exist between model location and model thickness. For example, if we move our preferred model -200 km (closer) toward the earthquake source and decrease the model thickness to 250 km (Figure A1a), or 200 km (farther) away from the source and increase the model thickness to 350 km (Figure A1b), the synthetics can also explain the observations. The quality of observations and their epicentral distance range of this event also prevent us from resolving the boundary geometry and velocity gradient in good confidence. For example, models with a shallowly dipping edge or a smaller velocity reduction seem to fit the observations as well (see Figure A1c for an example). There also is a trade-off between the thickness and the linear gradient of velocity reduction [Wen, 2001; Wen et al., 2001; Wen, 2002]. For example, synthetics for a model with a uniform -3% velocity reduction can match the observations for this event (Figure A1d).

[28] For event C (97/07/20), similar trade-offs exist between model location and model thickness. For example, if we shift the preferred model -400 km (Figure A2a) and 200 km (Figure A2b) away from the source, but change the thickness to 250 km (Figure A2a) and 350 km (Figure A2b), respectively, the synthetics of the two models can explain

the observed waveforms. We also test many models with different dipping edges. Synthetics (Figure A2c) for a model with a shallowly dipping edge (transition length is 800 km) can generally match the ScS-SH travel time residuals in Figure 5a. However, the ScS-SH time separations for this event, unlike those in event B (98/12/14), are so large that they allow us to resolve the velocity reductions of the model to some extent. We test a variety of models with a uniform velocity reduction of -3% and we are not able to find an appropriate model that can consistently explain the observations in the whole distance range, either by increasing the thickness, or varying the dipping edge or changing the location. We present such an example of synthetics for a model that is located 4400 km away from the source and has a steeply dipping edge, a thickness of 300 km and a uniform -3% velocity reduction (Figure A2d). The synthetics from this model match the ScS-SH travel time residuals and ScS/SH amplitude ratios well before 80°. However, they underpredict the ScS-SH travel time residuals after 80° (Figure A2d).

[29] For event F (95/02/08), in the absence of triplicated phases, significant trade-offs exist between model parameters. We present synthetics for some such models that can equally well explain the observations (Figures A3a-A3c) as the preferred model in Figure 8b. All these models have a linear gradient of velocity reduction from 0% (top) to -10%(bottom), but differ in thickness, location and dipping edge. In Figure A3a the model is moved 350 km closer toward the source relative to the preferred model in Figure 8b with a decreased model thickness of 275 km; in Figure A3b the model is moved 150 km farther away from the source relative to the preferred model in Figure 8b with an increased model thickness of 328 km; In Figure A3c the model has a shallowly dipping edge. Models with a uniform -3% shear velocity reduction underestimate the SHdiff travel time delays after 107° (see an example in Figure A3d). This problem cannot be solved by either increasing model thickness or moving the model closer toward the source, as both will result in larger coda phases and predict the same slowness of the SH diffracted phases as in Figure A3d. One needs to introduce a seismic anomaly in the shallower mantle for the models with a uniform 3% shear velocity reduction to be able to explain the observations in Figure 8a.

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