Topography of the western Pacific LLSVP constrained by S wave multipathing

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SUMMARY

We found that SH$_{\text{diff}}$ phases generated by earthquakes in the Fiji-Tonga, recorded in India, are accompanied by secondary pulses. We interpreted them as a consequence of multipathing of S waves caused by the Pacific Large Low-Shear-Velocity Province (LLSVP). We analyzed the differential travel times between SH$_{\text{diff}}$ and the secondary pulse, together with the absolute SH$_{\text{diff}}$ arrival times, to constrain the thickness and velocity perturbations in the western end of the Pacific LLSVP. Our preferred model shows a lateral variation in the thickness of the LLSVP; the southern part reveals a thicker (300 km) low velocity region compared to the northern part (200 km). However, the velocity perturbations of the LLSVP appear to be comparable (-1.5%). The results are consistent with a scenario that the LLSVP is a chemically distinct pile with significant surface topography.

Key words: Diffracted S wave, mantle plume, core-mantle boundary, Pacific Ocean

1 INTRODUCTION

It is now established that two Large Low-Shear-Velocity Provinces (LLSVPs) exist in the lowermost mantle beneath Pacific and Africa. Their overall geometry is constrained both by global (e.g.,

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Dziewonski 1984; Tanimoto 1990; Woodward & Masters 1991; Su & Dziewonski 1991; Ritsema et al. 1999; Becker & Boschi 2002; Takeuchi 2007, 2012; Lekic et al. 2012) and regional array studies (e.g., Ritsema et al. 1998; Ni & Helmberger 2003; He et al. 2006; Wang & Wen 2007; He & Wen 2012). Owing to the higher bulk sound velocities (e.g., Su & Dziewonski 1997; Masters et al. 2000; Koelemeijer et al. 2016), higher densities (e.g., Ishii & Tromp 1999; Trampert et al. 2004), sharp edges (e.g., Ni et al. 2002; To et al. 2005; Takeuchi et al. 2008; Sun et al. 2009; Frost & Rost 2014) and possible presence of sharp top surfaces (e.g., Sun & Miller 2013; Zhao et al. 2015), the LLSVPs have often been postulated to be chemically distinct from the ambient mantle (e.g., Tackley 2002, 2013).

It is now widely accepted that the height of the African LLSVP exceeds 1000 km in a cross sectional view through its southern and eastern ends (e.g., Ritsema et al. 1998; Ni & Helmberger 2003; Wang & Wen 2007). Although the stability of such large scale anomalies was debated in the context of mantle dynamics (Duvallie 1999; Tan & Gurnis 2005), several studies have succeeded in generating synthetic 3D models that explain the major features of seismological observations (e.g., McNamara & Zhong 2005; Bull et al. 2009). However, after the discovery of the post-perovskite phase (Murakami et al. 2004), some studies interpret these features by invoking thermal anomalies alone, considering the topography of the D” discontinuity and the contrast in material properties between perovskite and post-perovskite (e.g., Bull et al. 2009; Davies et al. 2012). In these studies, the LLSVPs are assumed to be plume-clusters blurred by limited resolution of the tomographic images (Schubert et al. 2004) rather than thermo-chemical piles. It appears that the origin of LLSVPs continues to be a matter of debate.

To constrain the origin, deciphering the detailed geometry of the LLSVPs would be useful. The height of the African LLSVP is suggested to be shorter (∼600 km) at its western edge (Sun & Miller 2013). However, the other details are not well constrained. Also, the height of the Pacific LLSVP is less constrained due to the inadequate coverage of the relevant seismic rays. Existence of small scale variations inside the Pacific LLSVP was suggested based on an abrupt change in the observed travel times (e.g., Takeuchi & Obara 2010; He & Wen 2012; Tanaka et al. 2015). However, it is not straight forward to identify whether such variations are caused by the geometry of the LLSVP or volumetric heterogeneities within. Although results from global tomography suggested that the height of the Pacific LLSVP is generally shorter (∼300 km) than that of the African LLSVP (e.g., Takeuchi 2007, 2012), detailed analyses of regional array data suggested significant regional variations (e.g., Takeuchi et al. 2008; He & Wen 2009, 2012).

In order to further constrain the geometry of the LLSVPs, new techniques and observations would be critical. To decipher the detailed structure of the lower mantle, analysis of the discrepancies in splitting of shear waves like S-ScS and SKS-SKKS (e.g., Kendall & Silver 1996; Lynner & Long
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2014; Long & Lynner 2015; Roy et al. 2014), waveforms of turning (e.g., PKP and Pdiff) and reflected ray phases (e.g., ScS, PeP) (e.g., Garnero 2000), is extensively conducted. In this study, we focus on multipathing of SH waves recorded by an array in India. We analyze the SH$_{diff}$ phase and its post-cursor to investigate the topography at the western edge of the Pacific LLSVP. This region is ideally suited for applying this technique because it is well sampled by ScS waves (see, for example, Fig. 2 of He & Wen (2009)) and the boundary location of the LLSVP, one of the critical parameters in this analysis, has been tightly constrained (see, for example, Fig. 8 of French & Romanowicz (2014)). For models with a low velocity zone (LVZ), we expect the arrivals of prograde and retrograde branches of direct waves to be registered at a single station. In the distance range of waveforms used in this study (102°-115°), we usually expect SH$_{diff}$ to be the direct phase. However, in the presence of an LVZ in the lowermost mantle, with appropriate velocity reduction and thickness, we also expect bottoming of the S wave, inside it (Supplementary Fig. S1). In this study, we have used the differential travel time between SH$_{diff}$ and SH, together with the differential apparent velocity, to constrain the thickness of the LVZ (i.e., LLSVP). We apply this approach to data recorded by the broadband seismograph array operated by the CSIR-National Geophysical Research Institute (NGRI), Hyderabad, India. These data have been previously used for studies on the lowermost mantle anisotropy and core-mantle boundary (CMB) structures, through analysis of SK(K)S and ScS phases respectively (Roy et al. 2014; Rao & Kumar 2014). However, the SH$_{diff}$ waveforms have not been analyzed yet, and the maiden use of this dataset therefore provides new information.

2 DATA AND OBSERVATIONS

In total, data from 49 stations in the Indian sub-continent are used in this study (Table S1, Fig. 1). The description of the station configuration and deployment is provided in earlier articles dealing with shear wave anisotropy of the Indian continent (Roy et al. 2012, 2014, 2016). Many of these stations are aligned in the north-south direction, around 80°E, and, for later discussions, we classify the array into the northern and southern arrays (the magenta and dark green triangles, respectively, in Fig. 1). The SH$_{diff}$ phase generated by earthquakes in the Fiji-Tonga subduction zone samples a portion of the western edge of the Pacific LLSVP (solid lines in Fig. 1). A systematic observation of the SH$_{diff}$ phase enabled us to examine the signature of the western edge of the Pacific LLSVP. To test the existence of a secondary pulse in each trace, we applied a two-pole Butterworth band-pass filter in the range of 3.2 - 32 s, to the displacement waveforms of earthquakes having depths > 100 km and M > 5.5, from the Fiji-Tonga slab. We did not observe any secondary pulses in the waveforms corresponding to shallow earthquakes. However, we chose to ignore the data of shallow earthquakes, since they are more complex and may suffer from phase overlapping due to interference with depth phases. While picking
the arrival time of a pulse, we applied a Butterworth band-pass filter of 3.2 - 12.5 s to the velocity
waveforms, because displacement seismograms often suffer from phase overlapping that blurs the
peak location of the secondary pulse. The corner frequencies are empirically chosen to maximize the
peak amplitudes of the secondary pulses.

A record section of the SH$_{diff}$ phase from the Fiji-Tonga region shows the arrival of a second
phase adjacent to it (Fig. 2AB). In order to further authenticate the secondary pulses, we also present
the displacement waveforms (Supplementary Fig. S2). The sections are made with respect to azimuth
(Fig. 2A) and distance (Fig. 2B). The secondary pulses are more coherent in Fig. 2A than in Fig. 2B,
which suggests the existence of azimuthally dependent structures. The second pulse has larger ampli-
tudes on the northern array (with larger azimuths) than those on the southern array. On an average,
the arrival time of the second pulse relative to the initial pulse is more advanced on the northern array.
There are 28 deep (>100 km) events from the Fiji-Tonga slab for which SH$_{diff}$ could be confirmed.
Qualitative inspection of the waveforms from these events resulted in the selection of five events,
which are assigned alphabetic names (Table 1), in which the second pulse is well identified as an
isolated packet (Fig. 2 & Supplementary Fig. S2). The remaining 23 events are not selected because,
either the waveforms are too noisy or the SH$_{diff}$ pulse is so broad that it overlaps with the second
pulse.

For the selected events, we pick the peak time of the first and second pulses within the SH$_{diff}$
window in the velocity seismogram. The second pulse is consistently well observed in the northern
array, whereas it is sometimes difficult to identify it in case of the southern array. We obtained 54
observations for the first and second pulses, in the distance range of 102.0° to 115.1°. The observed
relative arrival time of the second pulse with respect to the first pulse varies from 3.85 to 11.17 s. To
illustrate the sampling region, the vertical cross section including event B and station KLR (shown
in Fig. 1), superimposed on the tomography model of Takeuchi (2012), is shown in Fig. 3A. The
differential travel times show systematic variations when plotted at the entry point of the SH$_{diff}$ phase
at the CMB (Fig. 4). The differential times get larger when the ray samples the southern region. These
times also get larger when the entry point is located to the east and the ray travels a longer distance
in the lower velocity region. However, such a tendency gets weaker for the data sampling the northern
region. The observed amplitude of the secondary pulse is 32% to 95% of the SH$_{diff}$ pulse.

Fig. 5 shows the observed relative travel times as a function of distance. In many studies, the
distance is normalized to the source at the surface. However, in this study, we prefer not to follow that
approach because the ray is influenced more strongly by the distance between the source and the side
boundary, compared to the event depth. Indeed, even if we normalize the distance in a conventional
manner, the distance is shifted by less than 1°, which will not essentially affect the discussions below.
We modeled the observed travel times (shown by thick green bars) by the regression line in the form of $\delta t = a(\Delta - 100^\circ) + b$, where $\Delta$ is the distance and $\delta t$ is the differential travel time. The estimated values are $a = 0.10 \pm 0.05$ and $b = 4.84 \pm 0.47$ for the northern array and $a = 0.31 \pm 0.08$ and $b = 4.78 \pm 0.71$ for the southern array. We observe significant differences in the slope, which suggest structural variations in the azimuthal direction.

3 INTERPRETATION

The cross section of Fig. 3A shows that the SH$_{diff}$ phase samples the region with strong low-velocity anomalies associated with the Pacific LLSVP, traverses the western edge and pierces through the slightly higher velocity region. We interpret that the second pulse is generated from the lowermost mantle structure because of the following four reasons. (1) The possibility of a source side structure is small, since we did not observe a second pulse in the direct SH phase, for the same event recorded at closer distances in southeast Asia (Supplementary Fig. S2), (2) in a similar fashion, the possibility of receiver side structure is small, since we did not observe the pulse for events at closer distances, (3) the observed amplitude of the secondary pulse is $32 - 95\%$ of that of the direct wave, and such larger amplitude is not likely except for the mulitpathing or post critical reflection and (4) the Pacific LLSVP is the most pronounced feature in the cross section of Fig. 3A, and the LVZ above the CMB can generate multipathing (Supplementary Fig. S1); also, the differential traveltimes between the first and second pulses are correlated with the LLSVP geometry (Fig. 4). We interpret the second phase to be the direct SH wave that samples the top edge of the LLSVP and then gets refracted (or diffracted) before reaching the receiver (Fig. 3B).

We think it is unlikely that the secondary pulses are caused by random scattering or refraction traversing far outside of the great-circle plane. A plot of the displacement seismograms confirms that the secondary pulses gradually change with azimuth (Fig. 2 & S2). It intuitively implies that random scattering is unlikely to be the cause. Typically, the differential traveltime between the first and second pulses is $\sim 5$ sec. Considering the possible locations of point scatterers for a differential traveltime of 5 sec on the CMB, the seismic ray of the secondary pulse will not traverse more than $\sim 500$ km from the great circle plane.

Besides the volumetric heterogeneities, anisotropy can be another candidate to produce multipathing. Indeed, several studies suggested the existence of azimuthal anisotropy in the vicinity of LLSVPs (e.g., Wang & Wen 2007; Cottaar & Romanowicz 2013; Lynner & Long 2014; Ford et al. 2015; Creasy et al. 2017) and another lower velocity province on the CMB (Perm Anomaly; e.g., Long & Lynner 2015). However, these studies suggested anisotropy outside of the low velocity provinces rather than the inside. Considering the differential traveltimes are likely to be related to the LLSVP.
structure (Fig. 4), anisotropy is unlikely to be the major cause of the multipathing. We therefore ignore these effects in the following modeling.

4 MODELING

We search for a model that best explains the observed travel times. Because the observed waveforms are highly azimuthally dependent, we need to search for an azimuthally dependent structural model. Also, since the observed waveforms in Fig. 2A show that the amplitude and travel time of the secondary pulse gradually change from the southern to the northern array, we probably need a 3-D model with gradual azimuthal variations. However, since the computation of synthetic seismograms incorporating three dimensional models requires huge computational resources, we search for two dimensional models to explain the data from northern and southern arrays separately, to discuss the azimuthal dependence. We approximated the heterogeneous structures shown in Fig. 3A by a two dimensional model illustrated in Fig. 3B. The modeling is performed using the spectral-element code AxiSEM that simulates a 3-D wave field for an axisymmetric spherical model (Nissen-Meyer et al. 2014). Models for both the northern and southern arrays are azimuthally homogeneous and defined in terms of the cross section including Event B. The location of the other events and stations is appropriately projected on the defined cross sections. The heterogeneities are approximated by lower velocity and higher velocity blocks (illustrated by the red and blue boxes in Fig. 3B), and the rest of the regions are assumed to be laterally homogeneous. Readers might think that the model configuration assumed in Fig. 3B appears to be slightly ad-hoc, however, we think it is sufficiently justified. The boundary between the lower and higher velocity anomalies is assumed to be sharp, but it is consistent with the findings of Idehara et al. (2013), who suggested an abrupt change in ScS traveltimes with less than 500 km of transition zone. The top surface of the LLSVP is assumed to be horizontal, but we confirmed by numerical simulation that the presence of a large amplitude topography (>~ 200 km) with scale length comparable to the whole of our study region is unlikely because the direct S ray-path quasi parallel to the CMB and the top of LLSVP results in its refraction (or diffraction), and arrival as a secondary pulse adjacent to the SH_{diff} phase.

We define the following five model parameters (Fig 3); the height of the lower and higher velocity blocks (denoted by \( h_1 \) and \( h_2 \), respectively), their volumetric velocity perturbations (denoted by \( \delta v_1/v \) and \( \delta v_2/v \), respectively, where \( v \) is the velocity of the spherically symmetric reference model, PREM (Dziewonski & Anderson 1981)), and the boundary location (denoted by \( b^\circ \), which is the angular distance from event B). We determine the optimal parameters to explain the phases recorded by the northern and the southern arrays.
4.1 Model Inference

Synthetic travel times are evaluated using synthetic seismograms computed for the source-receiver geometries in our data set. We assumed a point source with a Gaussian source time function having a dominant period of 3.2 s. However, because the effects of anelasticity prevail, the resultant synthetic seismograms are essentially identical to those computed for a delta function. The temporary grid interval used in the computations is 0.025 sec, which is sufficiently small compared to the dominant period. After comparing synthetic seismograms computed using PREM with the observed waveforms corresponding to the selected events, we found that the pulse width in the synthetic seismograms is systematically broader than the observations. We therefore tuned the anelasticity factor $Q_s$ in the PREM; by tripling the $Q_s$ value in the lower mantle (i.e., we used $Q_s = 936$). We also found that the synthetic seismograms are more severely contaminated by crustal reverberations than the observed waveforms. Therefore, we replaced the Moho discontinuity with a smooth one. Using synthetic seismograms, the synthetic relative travel time between the first and the second phase is measured in the same way as done for the observed data. The above modifications of the model will not significantly affect the measurements of differential travel times. In view of the trade-off among model parameters $(h_1, h_2, \delta v_1/v, \delta v_2/v$ and $b^\circ$), we fixed three parameters $(h_2, \delta v_2/v$ and $b^\circ$) to appropriate values and searched for the optimal values of $h_1$ and $\delta v_1/v$ for the northern and southern array, respectively. We fixed $b^\circ=49^\circ$ at the location shown by the dark blue lines in Fig. 4, referring to the tomography model of Takeuchi (2012). Based on this tomography model, we also assumed $h_2 = 200 km$ and $\delta v_2/v = 1\%$.

The choice of these values is justified below. The boundary location ($b^\circ = 49^\circ$) is almost consistent (within an accuracy of $\sim 3^\circ$) among various tomography models (e.g., Houser et al. 2008; Simmons et al. 2010; Koelemeijer et al. 2016). All these tomography models show slightly higher velocity anomalies ($\sim 1\%$) to the west of this boundary, which is also consistent with our assumed model ($\delta v_2/v = 1\%$). Travel times of ScS reflections from this region also suggest slightly higher velocity anomalies (e.g., Houser et al. 2008; Idehara et al. 2013), and the absolute values of residuals are generally moderate/small ($\sim 4 s$), around a half of that for the phases reflected in the lower velocity region of our study area. We chose the appropriate values of $h_2$ and $\delta v_2/v$ from the tomography model of Takeuchi (2012) but the values ($h_2 = 200 km$ and $\delta v_2/v = 1\%$) are consistent with the ScS observations by Houser et al. (2008) and Idehara et al. (2013).

Figs. 6A,B show a comparison between the observed and synthetic differential traveltimes of the southern array for various pairs of $h_1$ and $\delta v_1/v$. To quantitatively confirm the consistency between the observation and the synthetics, we also modeled the travel times picked from the synthetic seismograms (shown by black pluses), by a regression line in a fashion similar to that in Fig. 5. If the thickness of the LLSVP increases or decreases, then the relative apparent velocity (i.e., slope of the
regression line) increases or decreases, respectively (Fig. 6A). In contrast, this velocity is not very sensitive to the volumetric velocity perturbation of the LLSVP, $\delta v_1/v$ (compare Figs. 6A & 6B). This is probably because, although the model is two-dimensional, the apparent velocity of the first and the second pulses primarily reflects the velocity at the bottoming points: mantle velocities at the bottom and at the top of the LLSVP (note that the bottoming point of the secondary pulse is below the top of the LLSVP). However, considering the grazing features of the ray path, we can assume that the velocity at the bottoming point is essentially equal to that at the top of the LLSVP). From a comparison of Figs. 6A,B, we concluded that $h_1 = 300$ km is optimal for the southern array.

Figs. 6C,D show the residuals between the observed and synthetic absolute travel times of the first pulse of the southern array for various pairs of $h_1$ and $\delta v_1/v$. The average residuals and their standard deviations are also shown in each figure. Because our data set of differential travel times does not constrain the volumetric heterogeneity very well (Figs. 6AB), we also refer to the absolute time. The primary problem related to the use of absolute arrival times is that the data is affected not only by heterogeneity in the lowermost mantle but also by other parts of the mantle. We therefore use the tomography model of Takeuchi (2012) to subtract the travel time residuals resulting from heterogeneities in the mantle, which are greater than 300 km from the CMB. The corrected residuals are plotted as brown pluses in Figs. 6C,D. However, it can be seen that the mantle corrections do not cause any systematic bias. From these comparisons, we concluded that $\delta v_1/v = -1.5\%$ is optimal for the southern array.

Similarly, for the northern array, the model parameters $h_1 = 200$ km and $\delta v_1/v = -1.5\%$ explain the travel time observations (Fig. 5A & Fig. 7). However, for the northern array data, the mantle corrections are rather systematic (Fig. 7B), probably because of higher velocities in the upper mantle beneath it. We therefore need to admit larger uncertainties for the volumetric heterogeneity ($\delta v_1/v$) of the northern region. However, we can well constrain the thickness $h_1$ from the relative apparent velocity that is insensitive to receiver- and source-side structures, and can conclude that azimuthal variation of topography is a robust component of our model.

In summary, our preferred model for the southern array is $h_1 = 300$ km and $\delta v_1/v = -1.5\%$, which produces travel times resembling the observed ones (Fig. 5B & Fig. 6). Our preferred model for the northern array is $h_1 = 200$ km and $\delta v_1/v = -1.5\%$, which explains the observations (Fig. 5A & Fig. 7). Although these preferred parameters are inferred from travel times, the resultant synthetic seismograms show a satisfactory agreement with the observed waveforms (Fig. 2).
4.2 Effects of Uncertainties in the Assumed Models

Our derivation of the optimal models is based on the assumption of the values of three parameters \((h_2, \delta v_2/v \text{ and } b^o)\). Of course, there are uncertainties in the assumed values, and we need to understand how they affect the waveforms and thus bias our inference of \(h_1\) and \(\delta v_1/v\). The boundary location \((b^o)\) is primarily sensitive to the amplitude of the secondary pulse and does not affect the traveltimes. When the boundary is shifted 3° towards east/west, the amplitude of the second pulse decreases/increases.

For our preferred model with the boundary location \(b^o = 49^o\), the amplitude of the secondary pulse in the synthetic seismogram is \(\sim 2.7 - 15.8\% \) of that of first pulse, whereas for the model with the boundary located at \(b^o = 46^o\), it is \(\sim 2.6 - 9.2\%\) (Supplementary Fig. S3). However, considering that the focusing effects cannot be modeled by our two dimensional simulation, we did not use detailed amplitude information to constrain the model. The volumetric heterogeneity \((\delta v_2/v)\) is primarily sensitive to the arrival times of the first pulse, while the thickness of the higher velocity anomalies \((h_2)\) does not much affect the waveforms, as long as it is confined to the lowermost 200 km (or as long as its height is smaller than that of the LLSVP). Therefore, the uncertainties in \(h_2\) do not affect much, but the uncertainties in \(\delta v_2/v\) should have some trade-off with our inference of \(\delta v_1/v\). However, we do not think that the assumed uncertainty in \(\delta v_2/v\) is large. We can therefore conclude that our inference on the variation in the inferred \(h_1\) is robust.

5 DISCUSSION

In this study, we showed that the observed waveforms can be explained without much alteration of the existing tomography models. Our observations suggest that the western edge of Pacific LLSVP extends in the NE-SW direction from 129.5°E to 136.5°E and has a smaller-scale topography. There is a variation in the topography of the western edge of the Pacific LLSVP from 300 km to 200 km, from south to north. The observations also suggest that the edge of the LLSVP (i.e., our study region) has both significant thickness and surface topography, which favor the LLSVP to be a chemically distinct pile. The existence of topography whose scale length is much smaller than the whole Pacific LLSVP suggests some dynamic process at the western edge such as a regional plume, instability of dome-like structure, or push due to downwelling from the outside (e.g., Tan & Gurnis 2007; Tan et al. 2011).

Sharpness of the top of LLSVP is important to constrain the origin of this large anomaly. We consider the same model as in Fig. 3B, except for the blurred top of the low velocity region, with a transition zone of up to 70 km thickness. The amplitude of the secondary phase decreases if the thickness of the transition zone is greater than \(\sim 60 \text{ km}\) (Supplementary Fig. S4). Although precise quantification is difficult, considering the observed amplitudes of the secondary pulses being more...
than \( \sim 30\% \) of the first pulse for most of the cases (see Fig. 2 & Supplementary Fig. S2), a gradual transition from the ambient mantle to the LLSVP is unlikely and the LLSVP should be a distinct region.

It is debated whether the origin of the LLSVP is thermal or chemical (e.g., McNamara & Zhong 2005; Bull et al. 2009; Davies et al. 2012). Our results favor a chemical origin because of the following reasons. If the thermal anomalies were the origin, we expect predominance of smaller scale anomalies with a complex geometry composed of plume clusters (e.g., Davies et al. 2012). Such smaller scale features (< 100 km) were required to explain the sharp sides inferred from previous seismological studies (e.g., Ni et al. 2002; To et al. 2005; Sun et al. 2009). However, to explain the secondary pulses observed in this study, we need a quasi-horizontal upper boundary of the lower velocity region (such as shown in Fig. 3B), otherwise we hardly expect phases with similar apparent velocity and arrival time as those for the \( \text{SH}_{\text{diff}} \) phase. To further constrain the origin, we probably need to analyze \( \text{P}_{\text{diff}} \) and its post- or pre-cursors, which is a future research topic.

Finally, we show the discrepancies between our preferred model and previous regional models, and discuss possible reasons for them. Idehara et al. (2013) previously studied a region similar to our study region. They used ScS phases propagating in the N-S and E-W directions. The ScS data in the E-W direction has an azimuthal coverage similar to our data from the northern array. Their model has a 250 km thick LLSVP with \(-1.5\%\) velocity reduction, bounded at 137°E with a +1.0% higher velocity region which is 250 km thick, while our model for the northern array has a 200 km thick LLSVP with \(-1.5\%\) velocity reduction, bounded around 136.5°E with a +1.0% higher velocity region. Our model is close to the model of Idehara et al. (2013), but has marginal discrepancies. We think it is, at least, partly due to the assumption in Idehara et al. (2013) that the boundary should be precisely in the N-S direction. Indeed, the boundary locations suggested by the N-S and E-W data of Idehara et al. (2013) are not very consistent with each other (see Fig. 3c of Idehara et al. (2013)). Furthermore, the residuals for the longitude ranges 126°E to 140°E are highly scattered (see their Fig. 2b). It appears that the ScS reflections sampling the northern part of their study region between 126°E – 135°E (which is in the higher velocity region in their model) systematically show delayed arrivals, which may be interpreted as a consequence of inappropriate boundary direction assumed in Idehara et al. (2013). The boundary direction and location of our model are generally consistent with the result of He & Wen (2012), who studied ScS travel times sampling the Pacific LLSVP and its adjacent regions. Although the boundary location of He & Wen (2012) is slightly eastward of our boundary, the difference is within the uncertainty limits. The fact that the higher velocity anomalies adjacent to the LLSVP were not considered in He & Wen (2012) may be partly contributing to the discrepancy.
6 CONCLUSIONS

The waveforms of five deep earthquakes from the Fiji-Tonga subduction zone recorded by a seismic array in India reveal a secondary pulse just after the $S_{\text{diff}}$ phase. The second pulse is suggested to be generated by multipathing of $S$ waves due to the Large-Low-Shear Velocity Province (LLSVP) atop the core-mantle boundary. The travel time difference between the second and first pulse together with their differential apparent velocity can be used to retrieve information on the azimuthal variations of the thickness at the western edge of the Pacific LLSVP. Forward modeling suggests that the western boundary of the Pacific LLSVP is oriented in the NE-SW direction and the thickness of the Pacific LLSVP varies from 300 km to 200 km from south to north. The observation favors the LLSVP to be a chemically distinct pile with significant surface topography whose overall geometry is dome-like rather than sandpile-like.

7 ACKNOWLEDGMENTS

SKR is thankful to the Director, NGRI for his support during the work carried out at the Earthquake Research Institute (ERI), University of Tokyo and for his permission to publish the work. SKR is thankful to Xin Long for his help in running the AxiSEM code. We sincerely thank the Editor and reviewers for their constructive comments that improved the quality of the manuscript. Waveform data used in this study can be obtained by sending a mail to the Director (director@ngri.res.in), CSIR-NGRI, Hyderabad. The Council of Scientific and Industrial Research (CSIR) and the Ministry of Earth Sciences (MoES) are gratefully acknowledged for funding the projects in the Indian seismic array. Plots are generated using the Generic Mapping Tools (GMT). This research is partly supported by JSPS KAKENHI Grant Number 15H05832.

REFERENCES


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Sun, D., Helmberger, D., Ni, S., & Bower, D., 2009. Direct measures of lateral velocity variation in the deep
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Table 1. List of earthquakes which show clear secondary pulses after the \( \text{SH}_{\text{diff}} \) phases. These earthquakes are assigned a alphabetic event name.

<table>
<thead>
<tr>
<th>E_Date</th>
<th>E_Time</th>
<th>E_Lat.</th>
<th>E_Long.</th>
<th>E_Depth</th>
<th>Mag.</th>
<th>Remark</th>
</tr>
</thead>
<tbody>
<tr>
<td>10/16/2007</td>
<td>21:05:46.800</td>
<td>-25.700</td>
<td>179.720</td>
<td>512.40</td>
<td>6.6</td>
<td>Event A</td>
</tr>
<tr>
<td>11/09/2009</td>
<td>10:45:03.400</td>
<td>-17.110</td>
<td>178.530</td>
<td>603.90</td>
<td>7.3</td>
<td>Event B</td>
</tr>
<tr>
<td>02/21/2011</td>
<td>10:57:57.600</td>
<td>-25.950</td>
<td>178.470</td>
<td>567.50</td>
<td>6.6</td>
<td>Event C</td>
</tr>
<tr>
<td>04/28/2012</td>
<td>10:08:12.700</td>
<td>-18.790</td>
<td>-174.260</td>
<td>140.70</td>
<td>6.7</td>
<td>Event D</td>
</tr>
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Figure 1. Source-receiver geometry of our data set for the five deep earthquakes from Fiji-Tonga slab (filled star) and Indian seismic array (inverse triangle). Color of the stars denotes the source depth, and magenta and dark green triangles indicate the stations belonging to the northern and the southern array, respectively. The magenta and dark green lines demarcate a ray-segment of $\text{SH}_{\text{diff}}$ phase on the CMB for the northern and southern seismic array, respectively. Background color represents the shear wave velocity variation in the lowermost mantle in the model SH18CE (Takeuchi 2012). The locations of station KLR, event B and event D (see the text for details) are also shown. The red triangles denote the locations of reference stations (whose waveforms are shown in Supplementary Fig. S2).
Figure 2. Velocity waveforms recorded by the Indian network for the event D (shown in Fig. 1) from Fiji-Tonga slab. The waveforms are plotted in terms of (A) azimuth and (B) distance. The horizontal axis is the time relative to the synthetic arrival time of $SH_{diff}$ phase computed using the PREM model (Dziewonski & Anderson 1981). The red arrow marks the second pulse. Synthetic seismograms for our preferred model are also plotted in terms of (C) azimuth and (D) distance.
Figure 3. A) Depth section of the shear wave tomographic velocity model (Takeuchi 2012) including the event B and station KLR. SH$_{diff}$ ray path for event B and station KLR (shown in Fig. 1) and direct S raypaths for epicentral distance between 80$^\circ$ and 100$^\circ$ computed using PREM (Dziewonski & Anderson 1981) are shown. (B) The two-dimensional model considered in this study together with the definition of model parameters. $h_1$, $\delta v_1/v$ and $h_2$, $\delta v_2/v$ represent the thickness and shear velocity perturbation for the first and second heterogeneous block, respectively. $b^\circ$ denotes the boundary location which is measured in terms of the distance from the epicenter of event B. Schematic ray paths of the first and the second pulses are also shown. The preferred values of the model parameters can vary between the northern array and southern array; $h_1 = 200$km, $h_2 = 200$km, $\delta v_1/v = 1.5\%$, $\delta v_2/v = 1\%$, $b^\circ = 49^\circ$ for the northern array and $h_1 = 300$km, $h_2 = 200$km, $\delta v_1/v = 1.5\%$, $\delta v_2/v = 1\%$, $b^\circ = 49^\circ$ for the southern array, respectively.
Figure 4. Travel time residuals (black circles) projected at the entry point of the SH$_{diff}$ ray-path to the CMB. The magenta and dark green lines represent the ray-segment of SH$_{diff}$ on the CMB for the northern and southern seismic array respectively. The background color represents the tomographic shear velocity model of Takeuchi (2012).
Figure 5. The observed differential travel times (black pluses) between the second and first pulses as a function of distance for the (A) northern and (B) southern seismic array, respectively. Green color bar represents the regression fit. The numerical expressions of $\delta t$ (in the inset) represent the explicit equations of the regression line along with their uncertainties. $\Delta$ denotes the distance in degrees.
Figure 6. (A) Synthetic differential travel times (black pluses) for the stations in the southern array calculated using PREM (Dziewonski & Anderson 1981) for $\delta v_1/v = -1.5\%$ and $h_1 = 200\ km$ (top), $h_1 = 300\ km$ (centre), $h_1 = 400\ km$ (bottom). Green color bar represents the regression fit of observed data for the southern array. The numerical expressions of $\delta t$ (in the inset) represent the explicit equations of the regression line for the synthetic travel times. $\Delta$ denotes the distance in degrees. (B) Same as (A) for $\delta v_1/v = -2.5\%$. (C) The travel time residuals of the first pulse (pluses) for the models with $\delta v_1/v = -1.5\%$ and $h_1 = 200\ km$ (top), $h_1 = 300\ km$ (centre), $h_1 = 400\ km$ (bottom). The black and brown pluses denote the residuals without and with the corrections for heterogeneities in the mantle, except for the lowermost 300 km region. The numerical values in the top inset represent the average and standard deviation of the residuals uncorrected for heterogeneities. (D) The same as (C) except for cases with $\delta v_1/v = -2.5\%$. In all the plots, we assume $\delta v_2/v = 1.0\%$, $h_2 = 200\ km$ and $b^\circ = 49^\circ$. Pink fill represents the selected model for the southern array.
Figure 7. (A) Synthetic differential travel times (black pluses) for the stations in the northern array calculated using our preferred model. Green color bar represents the regression fit of the observed data for northern array. (B) The same as Fig. 6C (top) for the residuals from the northern array.