Lithospheric Radial Anisotropy beneath the Gulf of Mexico

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Abstract

The Lithosphere Asthenosphere Boundary (LAB), where a layer of low viscosity asthenosphere decouples with the upper plate motion, plays an essential role in plate tectonics. Most dynamic modeling assumes that the shear velocity can be used as a surrogate for viscosity which provides key information about mantle flow. Here, we derive a shear velocity model for the LAB structure beneath the Gulf of Mexico allowing a detailed comparison with that beneath the Pacific (PAC) and Atlantic (ATL). Our study takes advantage of the USArray data from the March 25th, 2013 Guatemala earthquake at a depth of 200 km. Such data is unique in that we can observe a direct upward traveling lid arrival which remains the first arrival ahead of the triplications beyond 18°. This extra feature in conjunction with upper-mantle triplication sampling allows good depth control of the LAB and a new upper-mantle seismic model ATM, a modification of ATL, to be developed. ATM has a prominent low velocity zone similar to the structure beneath the western Atlantic. The model contains strong radial anisotropy in the lid where V_{SH} is about 6% faster than V_{SV} . This anisotropic feature ends at the bottom of the lithosphere at about the depth of 175 km in contrast to the Pacific where it extends to over 300 km. Another important feature of ATM is the weaker velocity gradient from the depth of 175 to 350 km compared to Pacific models, which may be related to differences in mantle flow.

1. Introduction

A fundamental issue in gaining a better understanding of plate tectonics is a clearer picture of how the Lithosphere Asthenosphere Boundary (LAB) varies with geologic settings. That is, how deep does differential flow between lithosphere and asthenosphere take place and how does it vary with plate history. Viscous shearing accommodates this relative decoupling motion by mantle flow creating seismic anisotropy produced by lattice-preferred orientation along with dynamic recrystallization (*Zhang and Karato, 1995*). *Conrad et al. (2007)* discussed how observations of seismic anisotropy can be used to help constrain geodynamic modeling which appears to work best for oceanic upper mantle beneath Pacific Basin. However, seismic anisotropy of oceanic lithosphere is difficult to image because of lack of station coverage and the difficulties of resolving the seismic low-velocity zone (LVZ). *Tan and Helmberger (2007)* conducted detailed modeling of observed waveforms that traveled along corridors across the Pacific Basin. By modeling the triplications of S-multiples out to SSSSS they were able to fit complete SH and SV records of 50 events containing earthquakes at various depths. Their PAC06 model for SH and SV contains a fast lithosphere (Lid) with a nearly uniform thickness of 70 km (Fig. 1). It compares closely with Pacific Basin models derived for several other corridors (*Gaherty et al., 1999*). We include four other models for comparison, ATL, CR3, PREM, and ATM, and the latter is a preview of results from this study. Fig. 1 also includes a hybrid isotropic model CR3 tested in this study, which is modified from the craton model CR (*Chu et al., 2012a*).

Oceanic upper-mantle structure has been extensively studied in terms of mineral physics. For example, *Stixrude and Lithgow-Bertelloni (2005)* attempted to explain the LVZ and oceanic upper mantle as a solid-state consequence of a thermal boundary layer and the effects of temperature on seismic velocities of sub-solidus mineral assemblages. They were able to explain

the velocity structure beneath the western Atlantic namely ATLS (*Grand and Helmberger, 1984*) and ATLP (*Zhao and Helmberger, 1993*) for S and P velocities, respectively (Fig. 1), but not the Pacific models PAC06. They suggest that adding some melt to the deeper structure (depths from 170 to 300 km) could explain the high velocity gradient of PAC06. Moreover, such melt could also cause anisotropic structures (*Kawakatsu et al., 2016*), which can help explain observed SKS splitting. This feature proves useful in explaining the complexity of SKS splitting near subduction zones as discussed by *Chevrot and van der Hilst (2003)*.

Perhaps the most definitive proof of the existence of anisotropy beneath the Pacific is long-period surface waves (e. g. *Beghein and Trampert, 2004; Kustowski et al., 2008*). However, to resolve the depth dependence of anisotropy requires samples of other phases such as those provided by triplication data as addressed in this report. In particular, we take advantage of USArray data from an event occurring beneath Guatemala to develop the ATM model. The upper mantle sampled along these paths is beneath the Gulf of Mexico, which is thought to be mostly old oceanic structure (*Bird et al., 2005*). Thus, our model will allow a comparison against PAC06 with distinctly different spreading rates and age.

In the next section, we introduce predictions from existing tomographic models and our modeling strategy. Next, we discuss the waveform fits in the results section and lastly, a discussion.

2. Data and Modeling Strategy

The waveform data addressed in this study were obtained from the USArray deployment in eastern US. In particular, we take advantage of the deep Mw6.2 earthquake at a depth of ~200 km which occurred beneath Guatemala on March 25th, 2013, see Fig. S1 for earthquake location and station distribution. All the recordings were first deconvolved from the instrument responses,

and rotated into the transverse and radial components to create a P-SV and SH set of record sections. A sample along a constant azimuth through Western Florida (Fig. S1) is displayed in Fig. 2.

2.1 Tomographic Predictions

Oceanic upper mantle structures are difficult to model because global seismic stations are mostly on land with distinct differences in crustal thickness. Reference 1D models assume basically an average of oceanic and continental crustal thickness, e.g. 24 km in PREM (Fig. 1), or they sometimes remove the crust assuming corrections from Crust 2.0, e.g. S362ANI (*Kustowski et al.*, 2008), see Fig. 2 and Fig. S2. The model S362ANI is special in that it was formed from a huge dataset and removes the velocity jump near 220 km in PREM which is not supported in triplication studies (e. g. *Tan and Helmberger*, 2007). It is, also, the 3D reference model used in the “Shake Movies” (<http://global.shakemovie.princeton.edu>). Thus, there are routine synthetics generated by the Spectral-Element Method (SEM, *Komatitsch and Tromp*, 2002a, 2002b) for special events such as the Guatemala event. Because these synthetics are filtered to 17s and longer, we regenerated these down to 10s and compared them against 2D displacement synthetics, see Fig. S2 for images in 3D and 2D corrected for the reference velocities. In general, 2-D synthetics based on cross-sections through smooth 3D models agree well with 3-D SEM synthetics as discussed in *Li et al.* (2014) for example record sections. However, the predictions given in Fig. 2 are not as good at the shorter distances since triplication data was not used in the making of these models. We have included a 2D image through the LLNL model (*Simmons et al.*, 2015), which contains a 3D crust, Fig. S2. The LLNL model is also special for several reasons; (1) it contains horizontal paths (P_n); (2) refined locations and origin times by the use of the multiple-event locator; and (3) it uses 3D ray-tracing. Note that

their image produces evidence of the subducted slab. Fortunately, global models based on adjoint full-waveform inversions are beginning to display evidence of slabs (*Bozdag et al.*, 2016). Slabs have a significant impact on waveforms at the shorter periods as demonstrated in *Zhan et al.* (2014). However, slabs can also introduce errors in location and origin times even at long periods, >17 s, which can easily introduce errors in tomographic images, $\pm 0.5\%$ (*Lu and Grand*, 2016). Thus, it is not surprising to see the obvious differences displayed in Fig. 2A which frames the difficulties in resolving velocity models from tomography alone. Fortunately, differential timing between the triplication branches isolated from depth phases (see ray paths in Fig. 2B) proves helpful. Note that direct paths out to about 20° sense a relatively uniform structure in both models. Larger distances begin to sample the craton structure.

In short, direct arrivals from such deep events provide nearly a pure-path sample of the upper mantle structure. Thus the structure can be idealized to a 1D model with less parameters than involved in 3D models. However, depth phases obviously sample a different 1D model which can affect the true source depths as in global body wave modeling. If we use the same 1D model for the source as for the receivers, the true depth and origin time will be incorrect. In particular, allowing different 1D velocity structures in the source and station regions can be easily handled for body waves with generalized ray theory (e.g. *Zhao and Helmberger*, 1994). Thus, oceanic structures can be studied using triplications from the upper-mantle for S, SS, etc. for the Northwest Atlantic (ATL) when recorded on land (e.g. *Helmberger et al.*, 1985; *Grand and Helmberger*, 1984). Fortunately, the Cut-And-Paste (CAP) methodology (*Zhao and Helmberger*, 1994) allows for such flexibility as reviewed next.

2.2 Structure modeling with CAP

While broadband modeling of regional seismograms proves useful in validating models, it becomes increasingly difficult at the shorter periods. There are numerous possible reasons for this, such as (1) source complexity involving finiteness, (2) site effects involving local geology near the stations (3) crustal structure with basins involving edge effects, (4) unknown depth dependent anisotropy. Inversion methods such as the one discussed at length in *Lee et al. (2014)* of local events (distances less than 500 km) demonstrate that 3D crustal models (assumed to be isotropic) are becoming resolvable down to 5s and can be used to predict accurate synthetics matching observations from assumed focal mechanism solution in California. Still shorter periods become increasingly difficult to model, because of source complexity mixing with the other propagation features in terms of a convolution operator.

An earlier and simpler approach was introduced by *Zhao and Helmberger (1994)* that grid searches for the source parameters (strike, dip and rake) by matching the data with 1D synthetics. Here in Fig. 3, decomposed 1D synthetic segments of P_{nl} (Extended P-wavetrain) and surface waves are allowed to shift to align with portions of the observed records to obtain the best fits, (*Zhu and Helmberger, 1996*).

Although CAP was originally developed for local source inversion, it is also useful in modeling structure with array data, essentially allowing individual station corrections to the velocity model. For example, sensitivity tests demonstrate that each regional record can be matched quite well by a particular simple layered model (*Song et al., 1996*). A best fitting model can be obtained by varying the layer velocities to obtain the best waveform Cross-Correlations (CC's) along with a constraint of allowed timing shifts. Essentially, one conducts a grid search of model parameters searching for the highest CC's and the smallest shifts. This is the approach we take in this study. While low velocity zones are relatively hidden in dispersion analysis, they

become particularly important in broadband modeling. For example, a similar recent study, applied to a much larger USArray data set for the Midwest, developed a crustal model with a prominent low velocity zone in the lower crust with the same low shear velocity of 3.6 km/s but higher P velocity of 6.7 km/s (*Chu and Helmberger, 2014*). Fig. 3A displays two examples of the procedure at stations U55A and N59. Note that the P-wave arrives 5.6 s and 3.8 s late with respect to velocity model CR3 at stations U55A and N59A (see Fig. S1 for locations), while the SV-segment is shifted by 6.8 s for both stations even though their waveform fits are quite good. Unfortunately, the SH radiation pattern is relatively weak compared to SV but many stations allow reasonable fits, i.e. N59 where the timing shifts are over 8s with SH being early. These timing shifts can be considered local corrections to the assumed 1-D velocity model.

A preliminary study of applying the CAP procedure to a large data set is presented in Fig. 3B. The 1D test model called CR3 (Fig. 1) fits this dataset quite well, in particular SV waves. Here, the SV model was produced by assuming the P-wave model CR3 with $V_s = 1.73V_p$ (*Chu et al., 2014*). Note the rapid change in the P_{nl} time shifts starting near the latitude of 40°. This feature is not so obvious in SV because the Rayleigh wave dominates the CC's which is insensitive to the deeper structure. This experiment established that $V_{SV} \neq V_{SH}$ and distinct mapping of seismic parameters is necessary, which is well known in long period dispersion studies, see *Lebedev et al. (2009)* for a modern treatment of this feature. However, the trade-off between anisotropy and source mechanism can be significant. Moreover, the estimates of origin time and depth depend on the model as well. Separating all these complications is difficult and running CAP on large datasets exposes these issues rather directly. For example, one advantage of this approach is that the direct trade off of source parameters involving depth phases, pP, sP etc., with the assumed model can be assessed and changes in the ratio of P to S velocity as well

as allowing V_{SV} to V_{SH} to vary independently (i.e. radial anisotropy). Essentially the CAP code is useful for self-correcting on non-1D features to assist the development of more complex models.

Starting with model ATLS, we concentrated on the triplication waveforms and timing along profiles while checking CAP runs on the full array for testing as in Fig. 3. We performed several iterations of CAP, adjusting V_P , V_{SV} and V_{SH} each time, to produce our preferred velocity model by a combination of trial-and-error forward modeling and grid search approach (e.g. *Chu et al.*, 2012a).

3. Results

The ray geometry provided by this deep event allows good sampling of the mantle structure beneath the Gulf of Mexico in terms of triplications (Fig. 4B). In particular, the up-going direct arrivals sample the high velocity lid (red) while the diving rays turn back below 200 km to the surface. We indicate the 410 sampling points by blue dots, which provide a clear view of this poorly sampled region by other methods (Fig. 4A). Our preferred models are presented in Fig. 4C which show a strong anisotropic lid structure with V_{SH} roughly 6% faster than V_{SV} .

The paths to Florida stations display some variation separated by the heavy green line in Fig. 4A. SH1 synthetics (in black) fits the southern data the best 14.8° to 16.2° and the SH2 synthetics (in red) fits the data at the larger distances somewhat better. We display the tangential component waveforms for the Florida stations in Fig. 5. Note that the small first arrival provides a strong constraint on the velocity of the lid although this arrival at some stations, especially to the south, are late and appear longer period than the 1D synthetics. We observed these small arrivals at many stations to the north and modeled them quite well in model SH2.

The high velocity lid not only produces the fast up-going first arrival, but also affects the diving ray paths when they return to the surface. If the V_{SH} and V_{SV} lid velocities are different, the SH and SV triplications will be shifted relative to each other as displayed in Fig. 6. Note that the differential times between the two branches CD (black) and AB (blue) is consistently larger and later for the SV phase, i.e. about 2-4s. Moreover, the effect becomes even more noticeable at the shorter distances where the ray bending encountered in crossing the high velocity lid in SH is stronger. The best fitting SV model also fits the Rayleigh wave presented in Fig. 7. Here, the synthetics for the two branches CD and AB match the data particularly well at the smaller distances. The differential timing displays some variation, i.e. at stations 957A to 959A where the CD branch displays the most scatter. This appears to be related to the issues raised in the SH modeling where there is evidences for a LVZ just above the 410. We fixed the depth features to be compatible with those determined in the SH-modeling, i.e. SH2 in Fig. 4. Since this data is broadband, it shows considerable scatter between the body-phases and the Rayleigh wave, but the principal arrivals fit both timing and wave shape for much of the TA stations as addressed in detail later.

The P-wave model is essentially unchanged from the starting model ATLP (Fig. 1). Synthetics for this model produce reasonable fits to the triplication data as presented in Fig. S3. A better fit is probably possible but our efforts here are on the SH and SV structures.

Beyond 20° the surface reflected phases become more prominent as displayed in Fig. 8. Both the tangential (Fig. 8A) and vertical components (Fig. 8B) are plotted with different reduced time scale along an eastern corridor where the station locations are identified in Fig. S1. The comparison between the SH synthetics for model SH2 and the data is given in Fig. 6A where the timing is satisfactory but there is some misfits with considerable more ringing in the data.

This is probably caused by the shallow basin structures near the coastal stations, i.e. 351 and is not in the 1D model. The sS phase is near nodal and, as expected, poorly explained along with the Love wave. In contrast, the S and sS (SV) arrivals are matched quite well including the Rayleigh wave and relative timing. Note that these fits are plotted in velocity which represents higher frequency and fit much better than the predictions given in Fig. 2. However, the P waveforms containing the surface reflected phases, pP and sP triplications are less well matched, probably due to the 3D structure in the source region. However, the fits are much better at the smaller distances as addressed next.

The summary of CAP results (Fig. 9) assuming the SV and SH model (ATM) derived above produce small travel time shifts and excellent waveform fits, many with cross-correlation coefficients (CC's) over 95%. Here we have included the entire dataset for completeness. First, the best fits are for the Rayleigh and SV phases as discussed above. Note that these high values are caused mostly by the timing differential and the longer period nature of the Rayleigh wave. The timing difference derived from the CAP method is displayed in the second column of Fig.9, mostly ranging from -2s to +2s. Note that at larger distances the ray samplings correspond to the transitions where upper-mantle structure changes from oceanic ATM to continental CR3 as suggested in Fig. 2. Note that the CR3 hybrid model used in Fig. S2 actually fits the light colored zone near the latitude of 40° for the extended P-waves better than the new ATM model since the latter has the 300 km structure, which impacts the CD back-branch. These 3D features will be addressed in future efforts where the slab-like feature near 40° in Fig. 2 becomes evident (*Ko et al.*, 2016).

Although the waveform modeling of the tangential profile (Fig. 8A) display some issues with sS, the overall fits become distinctly better at smaller azimuths where sS increases in

strength (Fig. 9C). Moreover, the southeastern Florida prefers some structural changes at depth, SH1, discussed earlier which becomes apparent in Fig. 9C. Note that both the Love (SH) and Rayleigh (SV) have high CC's for stations in the southern states with small time shifts, which is the object of this modeling strategy. The P-wave results are less good since the triplication branches produced by pP and sP became more 3D in nature near the source region.

Discussion

Our preferred model ATM is similar to that beneath the old western Atlantic (ATL) with a prominent low velocity zone centered at a depth of 175 km (Fig. 1). Note that the opening of the Gulf of Mexico occurred early in spreading history (160-140 My ago) and expected to have lid structure similar to that of the Western Atlantic (*Bird et al.*, 2005). The lid thickness is about 90 km with a $V_{SH}=4.63$ km/s, which is slightly lower than in the ATL model and also lower than the Pacific lid ($V_{SH}=4.78$ km/s). However, here this velocity was determined by the apparent relative timing involving 20 stations and is probably more accurate. The thickness control involves defining the differential timing between the AB branch (both V_{SV} and V_{SH}) which trades off with the SV velocity in the lid. In particular, the velocity profile for SH has a strong shadow zone for the AB branch while the SV does not. Thus, the ratio between AB/CD amplitude is distinctly different at the smaller distances (Fig. 6). Not only is AB(SH) earlier than AB(SV) it is also smaller. Some of the AB(SH) have some longer period precursors (i.e. 061Z) which could be tunneled waves (e.g. *Helmberger et al.*, 1985). In short, the SV displays a simple triplication while the SH becomes quite complex. Because the seismic radiation is quite smooth between these ray paths for AB and CD, this detailed data provides an excellent tool for studying the detailed wavefield caused by the structure of the LAB. In short, the timing and amplitude combinations between the AB and CD branches are quite consistent along with the surface wave

timing. In particular, the CAP shifts at distances less than 21° for both Love and Rayleigh waves
 are nearly zero. The lowest velocities in the LVZ are $V_{SV}=4.38\text{km/s}$ and $V_{SH}=4.46\text{ km/s}$
 considerably higher than beneath the Pacific Basin, Fig. 1, which reach to the lows of 4.22 km/s
 and 4.34 km/s . At longer distances beyond 22° , the edge of the craton complicates the picture as
 discussed in *Biryol et al. (2016)* and *Schmandt and Lin (2014)*. The transition to craton structure
 is particularly clear in the 2D images from the tomography models (Fig. 2) and plotted in true
 velocity after including the reference models, i.e. Fig S4. Inversions using long-period data
 capture this craton transition as a gradient, e.g. S362ANI, while shorter-period data image the
 change as a sharp feature (*Biryol et al., 2016; Schmandt and Lin, 2014*). Note that both short-
 period tomography models image fast velocities below 600 km apparently associated with the
 Hess Conjugate Subduction (*Ko et al., 2016*). Moreover, the ATM model has a low velocity
 gradient approaching the 410 km discontinuity. This structure appears to change somewhat for
 observations occurring along the southern Florida corridor. However, this feature has been
 observed further to the north beneath the craton and has been interpreted as caused by the release
 of water and slab-debris from deeper sinking slabs (*Chu et al., 2012a*). Such a feature appears to
 have global support and included in the new 1D earth reference model STW105 (*Kustowski et*
al., 2008). Note that the presence of such a molten layer provides evidence for a water
 component just above the transition zone which could be global (*Bercovici and Karato, 2003*). In
 short, this feature needs more attention from the mineral physics community (e.g. *Stixrude and*
Lithgow-Bertelloni, 2005).

Given the complexity in models such as displayed in Fig. 2 and the strong radial
 anisotropic introduced above, one might seek additional constraints from other datasets as in
Yuan et al. (2011) such as SKS splitting. Unfortunately, studies of anisotropy provide conflicting

views when comparing surface wave observations with SKS splitting (*Long and Becker, 2010*), especially near subducting slabs. Some recent results by *Song and Kawakatsu (2012; 2013)* on Pacific subduction zones have made significant progress in explaining such complexity by introducing a two-layered Lithosphere. They interpret the Pacific Basin Structure (PAC06 in Fig. 1) including the lid structure or slab and the lower gradient down to 160 km as a sub-slab. The model assumes that the lid has strong radial anisotropy (up to 6%) with $V_{SH} > V_{SV}$ which is frozen-in at birth, with the sub-slab influenced by flow. In particular, the subduction of the lid and sub-slab gradually tilts and allows V to become faster than V_{SH} for steeply dipping slabs and they can explain the SKS observations with a combination of general anisotropy. Such features are becoming apparent in global adjoint tomography (*Bozdag et al., 2016*), but subdued at such long periods. In contrast, the old Atlantic does not have much of a sub-slab (Fig. 1) and agrees with seismic velocities of expected minerals with no melt and a thinner decoupling zone, as discussed earlier (*Stixrude and Lithgow-Bertelloni, 2005*).

Since oceanic basins have relatively simple geologic history, they are probably the easiest to model following the two-layered lithosphere approach (*Song and Kawakatsu, 2013*). The question is how supportive are the global 3D models for such a feature. For example, a detailed comparison of the PAC06 model (SH) was performed along the path from Fiji-Tonga to Southern California for the tomographic model (*Liu et al., 2011*). Synthetics for both an idealized 1D model (see Fig. S4) and the 3D model produce reasonable fits but not as well as the PAC06 model that also fits the upper mantle reverberations equally well along the entire profile including the ScS^4 sample near the California coast. The lid thickness at this location is 55 km accordingly to *Melbourne and Helmberger (2001)*. The most definite definition of lid thickness was made by *Kawakatsu et al. (2009)* using OBS receiver-function analysis. They obtain a

thickness of 70 km with sharp bottoms just off of Japan at the age of 130 My and slightly thinner beneath the Philippine (20 My) plate of 60 km. These estimates are well below those found in the ATM and in ATL values of over 90 km. Note that the later two models have a very thin second layer and reduced velocity gradient down to 350 km. Thus, the model differences between the Pacific and Atlantic remains an issue with respect to lithospheric feature and it appears possible that the difference in spreading rates and tectonic history is the controlling physical feature. In particular, a fast spreading ridge (Pacific) would likely be hotter with a thinner lid and stronger LVZ. Partial melt is likely to further reduce the velocities and allow stronger convection which could promote anisotropic effects in the lower Lithosphere and upper most Asthenosphere (e.g. *Holtzman et al.*, 2003; *Lebedev et al.*, 2009). However, major compositional differences exist between the Atlantic and Pacific pillow basalts which could play a role (*Dick and Zhou*, 2014). More high-resolution sampling is needed to better map out the changes in Lid thickness and their relationship to major hot spots and tectonic history. In particular, 3D global models such as those in Fig. 3 could introduce slabs as proposed recently by *Lu and Grand* (2016) or add the lid structures approaching the subduction zones (e. g. *Chu et al.*, 2012b) involving the Juan de Fuca structure (Fig S5). In this case, the addition of slabs to tomographic models actually fit both regional travel times and waveforms better in support of plate tectonics.

In conclusion, there appears to be many approaches to understanding such complexity of the LAB. One is to reduce uniqueness in tomographic models by applying model cluster analysis (*Lekic et al.*, 2012). Another is to generate models from mineral physics and thermodynamic assumptions and generate artificial seismic models with embedded slabs and associated synthetic seismograms at all distances. Hybrid parameterizations can be injected into seismic waveform

modeling in self-consistent approaches (e.g. *Sun et al.*, 2013) for deep structure. Such an approach allows testing of models against seismological data more directly.

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Figure captions

Figure 1. (A) Comparison of velocity model ATM_{SH} from this study with ATLS (Grand and Helmberger, 1984), CR3 (Wei et al., 2013), and PREM (Dziewonski and Anderson, 1981). (B) Comparison of anisotropic models ATM with PREM (Dziewonski and Anderson, 1981) and PAC06 (Tan and Helmberger, 2007).

Figure 2. Waveform comparison of data (black), 2D (green), and 3D (red) synthetics generated from velocity models S362ANI (Kustowski et al., 2008) and LLNL (Simmons et al., 2015). (A)

2D cross-sections of S362ANI (left) and LLNL (right) along AA' shown in Fig. S2. The yellow star represents the event location used in this work. The black lines indicate the S wave raypaths predicted by PREM. (B) Comparison of tangential and vertical data and synthetics. All waveforms are band-pass filtered to 0.01~0.1 Hz.

Figure 3. (A) Representative waveform comparison between data (black) and the 1D synthetics (red) from velocity model CR3. The data have been divided into P_{nl} wave and surface wave segments, which are allowed to shift independently to find the optimal fits. The mechanism of the earthquake is shown as the beach ball. The cross-correlation coefficients of the fits between data and synthetics for P_{nl} waves and surface waves are shown in (B), while travel-time residuals for these two segments are displayed in (C).

Figure 4. (A) Map of the earthquake (beach ball), the southern USArray stations (triangles), and the mid-point of the ray path (blue dots) at distances less than 20.5° . Red triangles denote stations with a small arrival before the AB branch of the triplications, with station names in Fig. S1. The green line marks the boundary between SH1 and SH2 models. (B) Ray paths for a station with an epicentral distance of 15 degrees show three arrivals, labeled as red (traveling in the lid), AB (blue), and CD (black). (C) P, SH, and SV velocity models derived in this study.

Figure 5. Comparison of tangential data (black) and synthetics (red) generated from velocity models SH1 (A) and SH2 (B). Dashed lines denote three arrivals along the three paths in Fig. 2(B). All waveforms are band-pass filtered to 0.01~1 Hz. Note that the Lid arrival (red) is the earliest and branches AB (blue) crosses CD (black) near 18° . The green line marks the boundary of SH1 and SH2 in Fig. 4. Small changes in the fine structure approaching the 410 km discontinuity become important in timing.

Figure 6. SH (A) and SV (B) waves for Florida stations as in Fig. 3. Red dashed lines denote the arrivals traveling in the LID. Note that there is little evidence for the Lid arrival (red line) in the SV observations. The AB and CD branches are delayed on the SV component with respect to SH waves. All waveforms are band-pass filtered to 0.01~1 Hz.

Figure 7. SV and Rayleigh waveform fits for Florida stations applying the ATM_{SV} model. Both data and synthetics are band-pass filtered to 0.01~1 Hz.

Figure 8. Waveform fits for tangential SH (A) and vertical P-SV (B) waves. Synthetic tangential and vertical surface waves are calculated using velocity models ATM_{SH} and ATM_{SV} in Fig. 1, respectively, while synthetic P waves are generated using P model in Fig. 4C. Both data and synthetics are band-pass filtered to 0.01~0.1 Hz. Note the development of the Rayleigh wave with distance.

Figure 9. Cross-correlation coefficients and travel-time residuals between data and synthetics generated by our best model for P_{nl} (A), Rayleigh (B), and Love (C) waves. Synthetic Love and Rayleigh waves are calculated using velocity models ATM_{SH} and ATM_{SV} in Fig. 1, respectively, while synthetic P_{nl} waves are generated using P model in Fig. 4C. In the comparison, data and synthetics are both band-pass filtered to 0.01~0.1 Hz. Note that the radial component (SV) fits of the synthetics are very good out to about 25° along with the only small timing corrections. The stations near the east coast are near nodal for the tangential Love waves where the fits drop-off.

Figure

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