Azimuthal anisotropy in the D" layer beneath the Caribbean

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[1] The lowermost mantle beneath Central America has anisotropic seismic velocity structure manifested in shear wave splitting of signals from South American earthquakes recorded at North American broadband recording stations. Prior studies of deep mantle anisotropy in this region have characterized the structure as having vertical transverse isotropy (VTI), which is sufficient to explain a general trend of early tangential (SH) component arrivals. However, VTI models cannot quantitatively match systematic waveform complexities in the onset of many of the shear waves that graze this region. After accounting for splitting effects of upper mantle anisotropy beneath the recording stations, we model the corrected waveform data using full wave theory for mantle velocity models with an anisotropic D" layer. This is the first attempt to quantitatively model a large data set including azimuthal anisotropy in D". The models include transverse isotropy with either a vertical or tilted symmetry axis, the latter resulting in azimuthal anisotropy. For some initial shear wave polarizations, tilted transverse isotropy (TTI) produces small, reversed polarity arrivals on the SV components at the arrival time of SH, consistent with the data. Geographical variations in the azimuth of the TTI symmetry axis are indicated by the data. The lack of azimuthal coverage prevents unique resolution of the TTI orientation and also precludes distinguishing between TTI and other azimuthal anisotropy structures such as that predicted for lattice preferred orientation of minerals. Nonetheless, our modeling demonstrates the need for laterally varying anisotropic structure of more complex form than VTI for this region.


1. Introduction

[2] Determination of the detailed seismological structure of the lowermost mantle is of central importance in efforts to quantify the deep mantle dynamical system [e.g., Gurnis et al., 1998; Lay et al., 1998a]. The ~250 km thick D" region which overlies the core-mantle boundary (CMB) appears to have particularly acute seismic velocity heterogeneity on a wide range of scale lengths [e.g., Garnero, 2000; Lay and Garnero, 2004]. Seismological attributes of D" that have received extensive attention include P and S velocity discontinuities intermittently detected near the top of D" [e.g., Wysession et al., 1998], thin ultralow-velocity zones just above the CMB in many regions [e.g., Garnero et al., 1998], large low shear velocity provinces [e.g., Wen et al., 2001; Ni and Helmerger, 2003], and widespread anisotropic velocity structure within the boundary layer [e.g., Lay et al., 1998b; Kendall, 2000]. The overall seismological complexity of D" has motivated efforts to interpret the region as a major thermochemical boundary layer affecting both mantle and core evolution [e.g., Lay et al., 2004].

[3] While evidence for complex shear wave polarization of phases that sample D" extends back several decades [e.g., Mitchell and Helmlberger, 1973], the recent availability of high-quality broadband data sets has spurred numerous seismological investigations of D" anisotropy, and these, in turn, have motivated research in mineral physics and geodynamics. This broad interest stems from the potential constraints on mineralogy, structural fabric, and shear stress provided by knowledge of anisotropic properties of the medium [e.g., Karki et al., 1997; Yamazaki and Karato, 2002; McNamara et al., 2002, 2003; Moore et al., 2004; Panning and Romanowicz, 2004]. There are formidable challenges to resolving deep mantle anisotropy, with most regions of D" being sampled by very restricted azimuthal ray coverage, and accurate corrections for the significant effects of upper mantle anisotropy being required. It is commonly assumed that midmantle anisotropy is negligible, but this may not be the case for some regions [e.g., Wookey et al., 2002].

[4] The primary seismic phases that have been utilized to study D" anisotropy include (1) ScS, which reflects from the CMB, (2) Sdiff which is the combination of grazing S and ScS energy that diffracts along the CMB, and (3) SmKS phases, which traverse the core as P waves prior to their
upward bound legs through the mantle as S waves (Moore et al. [2004] review past studies using these various phases). Given the long path lengths of relevant seismic waves in the lower mantle, isolating the contribution to shear wave splitting from D'' structure is difficult. However, many observations of waves that traverse D'' have relatively simple shear wave splitting with the horizontally polarized S wave component (SH) arriving earlier than the vertically polarized S wave component (SV) [e.g., Lay et al., 1998b; Kendall, 2000; Thomas and Kendall, 2002; Moore et al., 2004]. This tendency is observed both for waves that graze nearly horizontally through the D'' region (S\text{diff}) and for waves with steeper angles of incidence through D'' (ScS). This type of splitting can be attributed to vertical transverse isotropy (VTI) at the base of the mantle, because the SH and SV systems are decoupled in a VTI medium. Comparisons of shear wave splitting in S and ScS arrivals on the same seismogram further demonstrate that the waves traversing D'' split in a manner consistent with VTI anisotropy, whereas those turning in the midmantle (and traveling through the same upper mantle region) do not [e.g., Kendall and Silver, 1998; Garnero and Lay, 1997; Lay and Young, 1991]. In a VTI medium purely SV-polarized phases such as SKS should not have any shear wave splitting (no SH component). Similarly, purely SH-polarized waves do not acquire any SV component.

[5] VTI could arise from lattice-preferred orientation (LPO) of crystals with their fast axes oriented randomly in the horizontal plane, or from hexagonal crystals with their symmetry axis oriented in the radial direction. Alternatively, VTI could also arise from fine-scale lamellae of strongly contrasting material properties in layers above the CMB [e.g., Kendall and Silver, 1996, 1998; Moore et al., 2004].

[6] There are some observations of ScS and S\text{diff} shear wave splitting that have SV components arriving earlier than SH components [e.g., Russell et al., 1999, 1998; Pulliam and Sen, 1998; Panning and Romanowicz, 2004], with little apparent coupling between the SH and SV components. There are also a few possible observations of coupling between SV\text{diff} and SH\text{diff} signals [e.g., Vinnik et al., 1989, 1995]. These indicate more complexity in D'' anisotropy than a uniform layer of VTI. The existence of azimuthal anisotropy in D'', with azimuthal variations of velocities and polarizations, has important implications for procedures used to determine upper mantle anisotropy. For example, SKS and SKKS splitting would occur on the upward bound leg through D'' [e.g., Hall et al., 2004], not just in the upper mantle as usually assumed. Niu and Perez [2004] found very limited evidence for variability between SKS and SKKS recorded at the same station, suggesting that the influence of lower mantle anisotropy is limited at most. If lower mantle anisotropy goes unrecognized it could lead to erroneous characterization of upper mantle anisotropy. It is also possible that apparent shifts of SV and SH arrivals may not involve complete separation of fast and slow polarized arrivals; weak energy on both components may easily be overlooked and the nature of the anisotropy misinterpreted. It is thus important to critically assess splitting observations that appear to be consistent with VTI, since many configurations of azimuthal anisotropy can be misinterpreted as VTI when limited azimuthal coverage is available (as is almost always the case for deep mantle studies).

[7] In this paper we conduct a detailed analysis of broadband waveforms sampling the D'' region beneath the Caribbean and Central America (Figure 1). The shear velocity structure of this region has been extensively investigated, with several studies addressing the shear wave splitting of ScS and S\text{diff} phases [e.g., Mitchell and Helmberger, 1973; Lay and Helmberger, 1983b; Ding and Helmberger, 1997; Kendall and Silver, 1996; Garnero and Lay, 2003; Rokosky et al., 2004; Garnero et al., 2004a, 2004b]. The general characterization of this region from these studies is that the D'' layer has VTI over a rather extensive region, with weak correlation between the strength of anisotropy and isotropic shear velocity heterogeneity. Shear wave splitting of as much as 9 s has been suggested for S\text{diff} observations beyond 105 degrees [Kendall and Silver, 1996], but most splits are from 0 to 5 s in the range 90 to 110 degrees, and most ScS splits are less than 3 s [e.g., Garnero and Lay, 2003]. If the anisotropy is uniformly distributed over a 250 km thick D'' layer, it involves about 0.5% anisotropy. Rokosky et al. [2004] performed a detailed analysis of ScS splitting recorded by dense broadband networks in California, finding that paths through D'' under the Cocos plate (the western part of our study area) are very well characterized by VTI-type splitting, with very little or no coupling between SH and SV arrivals for some of the cleanest data. However, Garnero and Lay [2003] noted other cases where a total decoupling of SH and SV is not observed. We reevaluate the broadband observations, finding a systematic presence of subtle waveform characteristics that require azimuthal anisotropy over the broad region below the Caribbean Ocean and Central America. In what follows, we perform a quantitative record-by-record waveform modeling study to characterize D'' azimuthal anisotropy that explains observations not fit by either VTI or isotropic models.

2. Presentation of the Data
2.1. Data Selection

[5] We analyze shear waves from 16 intermediate and deep focus earthquakes in the subducting slab beneath South America recorded by various broadband stations in North America. The data are a subset of those analyzed by Garnero and Lay [2003], selected for suitable signal-to-noise ratio attributes for analysis of shear wave splitting. Three component data have been deconvolved by the instrument response to obtain ground displacements, then rotated to the great circle path/angle of S wave incidence reference frame to obtain SH and SV displacements in the plane of the wave front. The rotation to obtain true SV helps to minimize any S-to-P conversions at velocity discontinuities below the receiver [e.g., Jordan and Frazer, 1975], which arrive as precursors on the vertical and horizontal SV component of the S or S\text{diff} arrival. Visual inspection of all three components is used to ensure that no S-to-P energy of significance is present on the true SV component that we model.

[5] We concentrate on S waves that turn deep in the mantle, either grazing the D'' region or diffracting along
the CMB, recorded at epicentral distances greater than 86 degrees. This constrains our data to observation from the Canadian National Seismic Network. Splitting of \( ScS \) phases from closer distance observations is not considered, because these arrivals are in the coda of direct \( S \), and the onset of \( ScS \) is often either ambiguous or not isolated enough to detect any weak coupling between \( SV \) and \( SH \) components.

2.2. Reversed Polarity \( SV \) Onsets

While the \( SH \) component is commonly observed to arrive ahead of the \( SV \) component in this data set, we do observe numerous waveforms with energy present on both the \( SH \) and \( SV \) components at the time of the \( SH \) onset. An example of such data is shown in Figure 2. The broadband velocity and displacement traces are shown for station FCC. On the velocity seismograms, the \( S \) wave appears to split fairly clearly, apparently with a simple delay of the \( SV \) component relative to \( SH \). However, on the displacement recording, there is energy on the \( SV \) component at the time of the \( SH \) onset. The onset of the \( SV \) component has opposite polarity to that predicted by the focal mechanism (\( SV \) is expected to have opposite polarity to \( SKS \) if both phases radiate from the same quadrant of the radiation pattern). This precursor does not appear on the other phases in the seismogram such as \( SKS \) or the \( SH \) component, ruling out explanations involving source complexity or overshoot effects from the instrument deconvolution process. A reversed polarity \( SV \) onset is evident in many waveforms of our data set, and mandates a fast wave polarization that has a significant \( SV \) component. This arises in structures which have anisotropy more general than VTI, involving \( SV-SH \) coupling \cite{Maupin, 1994}.

Figure 3 shows a collection of additional displacement recordings, illustrating that a variety of \( S \) and \( S_{\text{diff}} \) behaviors are observed for this region, including: rather simple \( SV \) and \( SH \) components with little apparent splitting.
records that are best explained by simple SV delays relative to SH (thus well explained by VTI), and data with reversed polarity SV onsets, suggesting SV-SH coupling.

[12] The problems with measuring splitting of Sdiff waves are compounded by the propagation complexities associated with structure in D[sub 00] and the difference in diffraction behavior for SV and SH components. The primary D'' structural affect is the possible presence of a discontinuous shear velocity increase at the top of D'', which causes a triplication of the wave field [e.g., Lay and Helmberger, 1983a]. If lowest mantle VTI anisotropy is distributed in D' (below the D'' discontinuity), the SV and SH components will differ in the strength of the triplication effects: for a VTI medium, the effective size of the SV velocity increase may be less than that for the SH velocity, causing polarization dependence of the triplication [see Matzel et al., 1996; Garnero and Lay, 1997]. Furthermore, as the wave approaches diffraction, destructive interference occurs between the opposite polarity SV and ScSV arrivals, while constructive interference occurs between the same polarity SH and ScSH arrivals. These effects result in (1) the SV signal rarely having the same wave shape as the SH signal, (2) rapid decay of SV amplitudes after the onset of core diffraction, and (3) polarization of the overall S arrival that is not linear, even for isotropic structures [e.g., Lay and Young, 1991; Maupin, 1994; Ritsema et al., 1998].

[13] Waveform complexities like those in Figures 2 and 3 are typical of our data, with contributions from the D'' discontinuity, D'' anisotropy and differential interference of S and ScS all playing a role. Given the complexity of the waveforms, the most direct indicator of the existence of non-VTI anisotropy is an SV arrival with onset polarity opposite to that expected based on the source focal mechanism. In an analysis based on a simple assessment of the initial polarization, it would be important to discard data for which either SV or SH is close to a radiation node. However, an approach based on full waveform analysis allows use of data with strong SH and weak SV because the “leakage” of the fast polarization is then easy to detect. In the present study, we retain all data for which the focal mechanism is known precisely enough for variations within the focal mechanisms error bars not to alter our conclusions.

2.3. Upper Mantle Anisotropy Corrections

[14] Before performing shear wave splitting analysis for teleseismic S and Sdiff, it is important to remove the signature of upper mantle anisotropy (UMA) by applying appropriate waveform corrections [e.g., Kendall and Silver, 1996; Garnero and Lay, 1997; Vinnik et al., 1998; Fouch et al., 2001; Garnero et al., 2004b]. Generally speaking, upper mantle anisotropy likely persists to depths of 200 to 400 km as suggested by mineral physics results [Karato and Wu, 1993] and corroborated by a host of seismic studies (i.e., see reviews by Silver [1996] and Savage [1999, and references therein]), but may exist to greater depths, particularly near subduction zones [i.e., Sharp et al., 1994; Fouch and Fischer, 1996; Montagner, 1998; Trampert and van Heijst, 2002; Wookey et al., 2002; Kavner, 2003; Cordier et al., 2004].

2.3.1. Receiver-side Anisotropy

[15] Receiver-side upper mantle anisotropy effects can be corrected for by using splitting parameters obtained from
studies of $SKS$, $SKKS$, and other shear waves for each station [Bostock and Cassidy, 1995; Barruol et al., 1997; Bank et al., 2000; Currie et al., 2004]. Using the fast polarization direction and splitting time values from these studies for the stations in our data set, we rotated the $SV$ and $SH$ components of each waveform to the fast and slow polarization directions, time-advanced the slow component by the splitting time, and rotated the components back to the original $SV$ and $SH$ orientations. When multiple published upper mantle corrections exist, we examined the range of corrections and found that the differences generally did not significantly modify our teleseismic $S$ and $S_{\text{diff}}$ splitting time observations.

In some cases, however, application of an UMA correction using published splitting parameters actually worsened the overall waveform quality, degrading the linearity of $SKS$ particle motion of our waveforms. For these stations (FRB, INK, PGC, RES, and WHY), we approached corrections in two ways that we believe best reduce potential contamination from upper mantle anisotropy. For stations FRB, PGC, and WHY, we evaluated $SKS$ splitting for individual events and utilized these single-event parameters to correct each corresponding waveform for upper mantle effects. Our derived splitting parameters were sometimes not significantly different from published results, but tended to improve the overall waveform quality, as discussed below. In other cases, individual splitting parameters were significantly different from published results, but again improved overall waveform quality. This result suggests that upper mantle anisotropy beneath several of these stations may be more complicated than originally interpreted (i.e., dipping or multiple layers); however, data sets for these stations were not sufficient to fully assess alternative models.

For stations INK and RES, we reevaluated upper mantle anisotropy effects by examining the full range of available events for shear wave splitting analysis of $SKS$ phases. These analyses yield a clear pattern of backazimuthally dependent fast polarization directions and splitting times. To first order, our results are consistent with the presence of two layers of anisotropy beneath these stations, contrary to the original interpretation of receiver anisotropy [Bostock and Cassidy, 1995]. Our analyses involve over twice the data numbers and the necessary improved backazimuthal coverage to assess the possibility of multilayer anisotropy, which was not viable in the original analysis of these stations. We are currently evaluating the range of anisotropic models that may explain these observations in order to provide improved predictive models for upper mantle anisotropy corrections based on backazimuth and incidence angle. For the purposes of this current work, however, we use uncorrected data, since the noncorrected data gave $SKS$ waves with least splitting in our data set.

The waveforms corrected for receiver-side upper mantle anisotropy were individually inspected to ensure that data quality was not compromised and, in fact, was improved. In most cases, we found that the corrected $SKS$ waveforms were accompanied by significantly less transverse energy, and that $S_{\text{diff}}$ waveforms were generally simpler in character. While it is likely that upper mantle anisotropy varies with angle of incidence between $S$ and $SKS$, the incidence angles for these phases typically differ by less than 10 degrees in our distance range and the effects are most likely negligible. Certainly, the receiver anisotropy models are not so robustly determined to warrant making specific angle of incidence corrections. While some uncertainty is associated with our corrected waveforms, we believe that at least the first-order effects of receiver anisotropy have been accounted for.

2.3.2. Source-side Anisotropy

To correct for source-side anisotropy is more difficult than to correct for receiver-side anisotropy. Shear wave anisotropy has been reported for stations in the vicinity of the epicenters of the events used here [e.g., Polet et al., 2000; Anderson et al., 2004]. By analyzing local $S$ waves and teleseismic phases, Polet et al. [2000] show that a significant part of the seismic anisotropy is located below the slab, but that there is no anisotropy below 500 km depth. Our deeper events should therefore be free from source-side splitting, whereas our most shallow events, at around 100 km depth, may have waveforms contaminated by source-side splitting. Shallow events are necessary to get some variation in the events focal mechanisms, and thereby in the $SV/SH$ ratio of the data, which in turn helps resolve elements in the model. We have therefore chosen to retain them in the analysis and made a thorough evaluation of possible biases due to source-side anisotropy in section 4.

3. Synthetic Seismograms in an Azimuthally Anisotropic $D''$

3.1. Modeling Method and Anisotropy Models

Considering the complexity of the propagation of $S$ waves close to the CMB diffraction onset, analysis of the data can only be done by comparing them with synthetic seismograms. The synthetics are made using the full wave theory method described by Maupin [1994]. This method is an extension of the Langer method [Richards, 1976]. It can account for full anisotropy in the $D''$ layer but does not allow for lateral variations of the elastic parameters. As opposed to ray-based methods, it can model diffracted waves and takes into account the multiple interactions of the waves with the CMB and the top of the $D''$ layer. This semianalytical method is very efficient from a numerical point of view and has been tested to model waves propagating in a large range of structures. An alternative to this method would be to use a reflectivity algorithm.

We calculate displacement seismograms in the frequency range 0 to 0.5 Hz, with an inverse quality factor of 0.003 throughout the mantle. We use the Harvard CMT solutions for the sources. The reference mantle shear velocity structure down to $D''$ is model SKNA1, derived for the Caribbean region by Kendall and Nangini [1996]. The presence of a widespread $D''$ shear velocity discontinuity in this region, first proposed by Lay and Helmberger [1983a], is supported by the work of Zhang and Lay [1984], Kendall and Shearer [1994], Kendall and Nangini [1996], Ding and Helmberger [1997], Garnero and Lay [2003], and Thomas et al. [2004]. Model SKNA1 has a 2.8% shear velocity discontinuity 250 km above the CMB, quite similar to the SLHA model of Lay and Helmberger [1983a]. The observation that regions with a strong $S$ velocity discontinuity at the top of $D''$ are often coincident with regions showing VTI [e.g., Matzel et al., 1996] leads...
us to consider models with a discontinuity in \(SH\) wave velocity but a smaller discontinuity in \(SV\) wave velocity. We primarily use models of \(D^0\) with a thickness of 250 km, an \(SH\) velocity discontinuity varying from 1 to 3.5% at the top, and different gradients down to the CMB (as shown in Figure 4). \(D^0\) layer thicknesses up to 400 km were also explored. Prior work on \(D^0\) anisotropy in the Caribbean region by Kendall and Silver [1996] and Garnero and Lay [2003] propose VTI models with 0.5–1% anisotropy in shear velocity distributed over the full thickness of \(D^0\). We consider here models with a somewhat stronger mean anisotropy of either 2.2% distributed over 250 km or 1.5% over 400 km. In all of the models, the cumulative amount of anisotropy (in depth) is the same, and only the depth distribution varies. If the depth distribution is not otherwise specified, the synthetics are made using model TOP250, a model having a 250 km thick \(D^0\) layer. We also produce synthetics for isotropic models, where the \(SV\) wave velocity is set identical to the \(SH\) wave velocity of the VTI models (see Figure 4).

[22] In order to introduce azimuthal anisotropy, we chose the simplest deviation from the VTI end-member: the symmetry axis of a transverse isotropy medium was tilted from the vertical, a configuration we refer to as tilted transverse isotropy (TTI). TTI has the advantage of involving only two free orientation parameters: the tilt of the transverse isotropy (TI) symmetry axis from the vertical, and the azimuth of the tilt direction. As there are observations in our study area that are fully compatible with VTI intermingled with some requiring azimuthal anisotropy, it is convenient to consider a range of models from vertical to tilted TI media.

[23] When there is little azimuthal coverage of an anisotropic medium, as is the case for our data set, Maupin [1994] has shown that the \(S\) waves can resolve only two elements in the anisotropic parameters of the \(D^0\) layer: the direction of polarization of the fast and slow waves in the plane perpendicular to the propagation direction, and the overall amplitude of the anisotropy. Different anisotropic models which yield the same time shift and the same direction of fast polarization cannot be distinguished with the present data set and we cannot hope to uniquely constrain the geometry of the anisotropy. Since TTI is able to represent all possible directions of fast and slow polarizations and degrees of anisotropy, it provides a complete model space for the purpose of assessing whether azimuthal anisotropy can quantitatively explain the waveforms observed under the Caribbean. The inferences based on TTI models have the advantage of being independent of any explicit physical model of the anisotropy in \(D^0\). However, we also calculated synthetics using the elastic coefficients measured by Yamazaki and Karato [2002] for magnesiowustite ((Mg,Fe)O), a candidate for acquiring mineralogical LPO anisotropy in \(D^0\). Comparison of these seismograms with those calculated for TTI models will be explored in section 5.

[24] We illustrate here the influence of different model parameters on \(S\) waveforms by analyzing synthetics calculated for various isotropic and anisotropic models of \(D^0\) in a limited number of source and station configurations. In
section 4, we compare the whole data set with appropriate synthetics. The source characteristics and epicentral distances corresponding to three actual records in our data set are used to demonstrate the dependence of synthetics on the model parameters. The first two cases are recordings of the 29 April 1994 earthquake, at stations PGC and YKW2. The Harvard centroid moment tensor (CMT) solution for this event predicts negative polarizations for $SH$ and $SV$ components at both stations. A reversal of the initial $SV$ polarization (i.e., an upswing) is observed at YKW2, but not at PGC. The third case is for the 29 June 2001 event, at station DLBC, for which the CMT solution predicts opposite polarities of $SV$ and $SH$ components, as observed in the data. This demonstrates the fact that not all data have flipped polarity $SV$ onsets (coincident in time with the $SH$ onset), and in fact, many appear well explained by VTI anisotropy as put forth in previous studies of this region.

### 3.2. Effect of Tilt

[25] Figure 5 shows predicted $S$ waveform responses for VTI and TTI models. The synthetics are shown at two different epicentral distances (93.6 and 99.4 degrees). Synthetics for the isotropic model are shown on the bottom line and those for transversely isotropic models on the four lines above, with the tilt increasing from zero (VTI) to 10, 20 and 40 degrees. The tilt is orthogonal (tilting to the right) to the propagation path. The $SH$ waveforms show little effect of TTI. The $SV$ component, which has two pulses in the isotropic case due to the presence of the $D''$ discontinuity, has a simpler waveform in the VTI model, because the discontinuity in $SV$ velocity is relatively weak. Our VTI model causes an $SV$ time delay of 3 to 4 s at 93.6 degrees epicentral distance. Within this time shift there is a small positive upswing at the front of the $SV$ pulse when the TI symmetry axis is tilted away from the vertical axis. This upswing is more prominent at larger epicentral distances.

### 3.3. Effect of Depth Distribution

[26] Synthetic seismograms were constructed for different $D''$ thicknesses and depth distributions of $D''$ anisotropy. Figure 6 addresses sensitivity of $S$ waves to vertical distribution of anisotropy in $D''$. All of the synthetics shown were calculated for a tilt of the TI symmetry axis 20 degrees to the right of the propagation path (i.e., to the east for our waves traveling north from the earthquake to receiver). The three bottom traces correspond to a 250 km thick $D''$ layer with three different distributions of anisotropy, as indicated in Figure 4. The three synthetics are similar, indicating that $S$ waves are not able to resolve the detailed depth distribution of $D''$ anisotropy across a 250 km thick layer (at least for these structures). The two top traces correspond to distributions of the same total amount of anisotropy over a 400 km thick layer. The transverse components are affected by this thickening of the $D''$ layer. They display...
pulse broadening at closer epicentral distances and an elongated tail at larger epicentral distances. These characteristics are also present in isotropic models. The \( SH \) component waveforms thus are sensitive to the depth of the discontinuity, but less so to the depth distribution of the anisotropy. The \( SV \) component, which is the component sensitive to anisotropic structure, does not change significantly when the \( D^0 \) layer thickens from 250 to 400 km. We therefore cannot resolve differences between models ranging in thickness between 250 km and 400 km.

3.4. Effect of Azimuth

[27] The influence of the azimuth of the TTI symmetry axis on the wave field is shown in Figure 7. The azimuth increases from zero (bottom trace) to 315 degrees (top trace) in 45 degree increments. The source-receiver geometries are the same as those of Figure 5, which produce \( SH \) and \( SV \) components with different polarities. The \( SH \) component varies little with azimuth and is therefore plotted only for one azimuth in order to better display the \( SV \) components. We observe that for azimuths of 0 and 180 degrees, corresponding to a tilt either toward the receiver or toward the source, the \( SV \) components arrive with a small delay compared to the \( SH \) components; these models cannot be distinguished from a model with a vertical symmetry axis. For tilts at azimuths 45, 90, and 135 degrees (i.e., to the right of the propagation direction), the \( SV \) components in synthetics corresponding to the 29 April 1994 (Figure 7, left) event have a small initial upswing, which does not vary much with azimuth. Synthetics corresponding to parameters of the 29 June 2001 event (Figure 7, right) show no delay or polarity changes for the same azimuths. Conversely, for tilts toward the left of the propagation direction, from 225 to 315 degrees azimuth, the \( SV \) components show a polarity change in synthetics for the event 29 June 2001 but not for 29 April 1994 event. The presence of a reversed polarity onset depends on how the incoming \( S \) wave is polarized compared to the fast and slow velocity directions. If its polarization is closer to that of the slow velocity, we observe energy of the fast wave during the first few seconds preceding the arrival of the slow wave, before the two waves interfere with a polarization close to that of the original \( S \) wave. Conversely, if the incoming \( S \) wave is polarized close to the fast velocity direction, we do not observe a polarity change at the beginning of the waveform.

[28] The \( SV \) waveforms change significantly depending on the azimuth of the tilted TI symmetry axis. Flipped polarity \( SV \) onsets occur only for specific combinations of focal mechanisms and TI symmetry axis azimuths. In section 4 we show that through analysis of a large data set, we can distinguish between best fitting \( D^0 \) anisotropy models having VTI or TTI, for specific TTI orientations.

4. Analysis of the Data Set

[29] We analyze every pair of \( SH \) and \( SV \) components by computing a suite of synthetic seismograms for the appropriate source parameters and epicentral distances. Synthetics for four different models are computed: (1) an isotropic model (ISO), (2) the VTI model TOP250, (3) a TTI model with the same depth dependence of anisotropy as TOP250, but with the TI symmetry axis tilted from the vertical by 20 degrees toward the right (east) of the propagation path, and (4) the same as above, but the TI axis tilts 20 degrees to the
left (west). These synthetics were compared to observations before and after correction for upper mantle anisotropy. For each record, our first step was to eliminate models that poorly match observed $SH$ and $SV$ behavior, paying particular attention to the first cycle of $SH$ and $SV$ pulses. For many observations, more than one model fits the data equally well. At most stations, the UMA correction does not modify the result of the analysis. We consistently used noncorrected data at stations RES and INK, where it was not possible to find a consistent UMA correction, as discussed in section 2.3.1. At other stations, e.g., DAWY, FRB, PGC and WHY, the UMA correction consistently reduced any transverse component energy of the SKS wave. For such stations, we compare synthetic predictions to the UMA-corrected data.

4.1. Detailed Analysis of Selected Data

[30] To demonstrate the constraints and uncertainties in the modeling procedure, in Figure 8 we show data synthetic comparisons for four example data. Since the $SH$ components match well for all the models, we only show $SV$ components. $SH$ components have been used for the alignment of the traces in time and amplitude scale (by aligning synthetic $SH$ predictions to respective observed $SH$ arrivals). The $SV$ predictions for the 4 different models are typically quite different. For record 6, only the TTI model dipping eastward is able to reproduce the small positive arrival at the beginning of the $S$ phase. Westward tilting models produce a negative pulse at the beginning of the phase and can be ruled out. VTI or isotropic models would fit the data well if the initial pulse were absent or smaller.

[31] Record 7, from the same event, does not show any significant $SV$ pulse at the arrival time of the $S$ phase. This fits better with isotropic or VTI models which are predicted to produce smaller and lower frequency arrivals than the TTI models.

[32] Record 30 is an example of data which do not allow discrimination between the isotropic model and the eastward TTI model. The data show a clear negative $SV$ pulse which is not delayed relative to the $SH$ component. This rules out the westward TTI model and the VTI model. While the amplitude and the arrival time are slightly better fit by the eastward dipping TTI model, in this case it is difficult to confidently distinguish it from the isotropic model predictions, thus both are considered plausible.

[33] Record 53 has a clear positive pulse at the beginning of the $SV$ component of the $S$ wave. Only eastward TTI models are able to produce such a reversed polarity onset. In this case, the data have a higher frequency content than the synthetics, a feature which would be better matched by anisotropy higher up on the wave path, in particular under the station. We note, however, that the $SH$ component at this station is also significantly higher frequency than the synthetics and that the SKS wave, visible 60 s before the $S$ wave on the $SV$ component, also has a high-frequency content. The SKS wave has a clear onset, without any
significant incorrect polarity at its onset, ruling out deconvolution or crustal conversion problems. The SKS arrival also does not have any SH component, indicating an effective UMA correction at this station (Yellowknife, Canada). Despite its somewhat high-frequency content, we consider the flipped polarity onset at the beginning of the SV waveform to indicate eastward dipping TTI in D00 for this record. Yellowknife contains 4 broadband stations, (YKW1-YKW4). In most cases, data from 3 to 4 of the YKW stations were available, showing almost identical waveforms. We model only one of the stations (YKW2) so that the data are not weighted higher than for other isolated stations.

4.2. Analysis of the Whole Data Set

[34] The waveform analysis detailed above was conducted for the entire data set, and the results are summarized in Table 1. Events are ordered by latitude (starting from the north) with records ordered by increasing epicentral distance. For each record, best fitting model(s) are indicated. Figure 9 shows the fits between all the data and synthetics, ordered as in Table 1. UMA-corrected data are shown except for stations RES and INK, where the correction is not applied. One synthetic is shown in each case: the synthetics for the isotropic model is shown when it fits. Otherwise, whenever one of the TTI models fits and the isotropic model does not, we show the synthetics for the corresponding TTI model. The synthetics for the VTI model are shown when this model is the only one that fits. Figure 9 illustrates the quality of the data set and of the synthetic fit.

[35] In most records, the SH component is larger than the SV component. This is due partly to focal mechanisms, and partly to the fact that the data set contains diffracted S waves for which the SV component amplitude decreases faster with distance than SH. In the majority of cases, the SH component has a simple waveform which is matched very well by the synthetics. The SV waveforms are much more variable, and are often delayed with respect to SH (e.g., see records 10 and 23). In many cases, there is no delay between the components but the SV component starts with a small swing which has polarity opposite to the main part of the phase. This is particularly visible on records with very low noise level (see records 15, 40, 42, 49 or 64). The SV component sometimes shows unexplained high-frequency content (e.g., records 50 or 69), a feature which is difficult to produce with D00 structure and current attenuation models of the mantle.

[36] Five records (records 1, 2, 5, 9, 74) are matched equally well by all the model predictions. These are all recordings at epicentral distances less than 90 degrees and for which the propagation through D00 is not long enough for anisotropy in our models to significantly affect the waveforms. In all these cases, the synthetics for the four models are nearly identical and fit the data well. The absence of any anomalous behavior in these data supports the notion that no lower mantle anisotropy above D00 is required by our data set. The only exception is record 17, also recorded at short epicentral distance, which has a small precursor matched by none of our models because the wave turning point is slightly above D00. Westward tilted transverse isotropy
higher up in the lower mantle would produce the precursor and fit this record better.

We investigated whether a 400 km thick D'' would fit the data better than the 250 km thick D'' layer employed above. In the majority of data synthetic comparisons, a 250 versus 400 km thick anisotropic D'' does not change the SV component waveforms, and there is no affect on our conclusions concerning anisotropy. On the other hand, the waveforms of the transverse components are sensitive to the depth of the discontinuity, which in our models coincide with the top of the anisotropic zone. Twenty-five percent of the records, mostly data recorded at large

Figure 9. Summary showing all the data used in this study together with corresponding synthetic seismograms. (bottom) The data and (top) the synthetic seismograms of each comparison, numbered according to Table 1. SH components are shown as dotted lines and SV components as full lines. The data traces are 80 s long. The synthetic seismograms are chosen according to the rules detailed in the text.
epicentral distances, are in better agreement with a 400 km thick D'' layer, whereas 40%, mostly recorded at shorter epicentral distances, favor the thinner model. The implications of this observation will not be discussed further here, since this does not affect our conclusions regarding anisotropy.

We have focused on how the synthetics fit the early part of the observed waveforms. Later parts of the waveform are sometimes poorly matched, but this is not unexpected since they are more sensitive to the thickness of the D'' layer and to lateral heterogeneities. The D'' layer is more strongly heterogeneous than the rest of the lower mantle [e.g., Masters et al., 2000] and lateral heterogeneities will certainly affect the waveforms. For isotropic heterogeneities, Emery et al. [1999] showed that conversion between the SH and SV components occurs only when the heterogeneity is significantly out of the source-station plane. In that case, any waves scattered by the heterogeneities arrive
Table 1. Summary of Correlations Between Data and Synthetics

<table>
<thead>
<tr>
<th>Record Event Date</th>
<th>Depth, km</th>
<th>Station</th>
<th>Δ</th>
<th>Azimuth</th>
<th>Midpoint Matches</th>
<th>Matches</th>
</tr>
</thead>
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<tr>
<td>1 20 Jun 2003</td>
<td>555</td>
<td>INK</td>
<td>3.32</td>
<td>-87.31</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>2</td>
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<td>x</td>
</tr>
<tr>
<td>3 12 Oct 2002</td>
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<td>INK</td>
<td>3.29</td>
<td>-87.39</td>
<td>x</td>
<td></td>
</tr>
<tr>
<td>4 10 Jan 1994</td>
<td>596</td>
<td>RES</td>
<td>3.51</td>
<td>-91.06</td>
<td>x</td>
<td></td>
</tr>
<tr>
<td>5 16 Jun 1994</td>
<td></td>
<td>DAWY</td>
<td>3.78</td>
<td>-89.61</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>6 8 Oct 1998</td>
<td>136</td>
<td>RES</td>
<td>3.62</td>
<td>-75.53</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>7 10 Jan 1994</td>
<td></td>
<td>RES</td>
<td>3.73</td>
<td>-76.42</td>
<td>x</td>
<td></td>
</tr>
<tr>
<td>8 30 Nov 1999</td>
<td>128</td>
<td>YKW2</td>
<td>3.78</td>
<td>-83.37</td>
<td>x</td>
<td></td>
</tr>
<tr>
<td>9 29 Jun 2001</td>
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<td>LLLB</td>
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<td>-88.12</td>
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</tr>
<tr>
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<td>PHC</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>11 12 May 2000</td>
<td>225</td>
<td>FRB</td>
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<td>x</td>
<td></td>
</tr>
<tr>
<td>12 19 Aug 1994</td>
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<td>x</td>
<td></td>
</tr>
<tr>
<td>13 10 May 1994</td>
<td>600</td>
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<td>x</td>
<td></td>
</tr>
<tr>
<td>14 24 Sep 2002</td>
<td>119</td>
<td>FCC</td>
<td>3.78</td>
<td>-84.29</td>
<td>x</td>
<td></td>
</tr>
</tbody>
</table>
in the coda of the direct waves and do not affect the polarization of the early part of the S waveform. By focusing on the early part of the SV waves, our conclusions should not be biased by the likely presence of strong isotropic heterogeneities in D00.

[39] Of the 80 records analyzed here (not considering the 5 records which were matched by all models), 34 are compatible with eastward dipping TTI (12 of these are also fit by at least one other model), 16 with westward dipping TTI (13 of these are also fit by at least one other model), and 25 are not matched by a TTI model. The data show spatial trends indicating that no single uniform anisotropy orientation in D00 beneath the Caribbean and Central America can fit all of the data.

### 4.3. Possible Biases Due to Non-D00 Effects

[40] Observation of shear wave splitting provides very little indication of the exact distribution of the anisotropy along the propagation path. Straightforward numerical experiments show that anisotropy in the upper mantle that gives rise to a 2 s shear wave splitting with favorable direction of polarization of the fast and slow waves can produce waveforms very similar to what TTI in the D00 layer produces. In order to isolate the anisotropic region giving rise to the observed splitting, we have to rely on indirect observations and inferences. One example is the identification of the bottoming depth at which deep mantle S waves first exhibit shear wave splitting. This is an extremely difficult measurement to quantify, primarily because of uncertainties of the S wave path (e.g., knowledge of any deep mantle discontinuity structure is required, as this can change the ray path bottoming depth by hundreds of kilometers), and also because the SV component is contaminated by the SKS coda and SKKS emergence at shorter distances (e.g., up to 84–86 degrees).

[41] Figures 10 and 11 show the distribution of best fitting anisotropy models for the distribution of stations and events, respectively. The most robust evidence for

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**Table 1. (continued)**

<table>
<thead>
<tr>
<th>Record</th>
<th>Event Date</th>
<th>Depth, km</th>
<th>Station</th>
<th>Δ</th>
<th>Azimuth</th>
<th>Midpoint</th>
<th>Matches</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Lat Long</td>
<td>iso vti east west</td>
</tr>
<tr>
<td>73</td>
<td>RES</td>
<td>107.32</td>
<td></td>
<td>353.07</td>
<td>21.79</td>
<td>21.79</td>
<td>-75.24</td>
</tr>
<tr>
<td>74</td>
<td>16 Jun 2000</td>
<td>120</td>
<td>ULM</td>
<td>86.82</td>
<td>343.76</td>
<td>8.21</td>
<td>-81.29</td>
</tr>
<tr>
<td>75</td>
<td>SCHQ</td>
<td>88.40</td>
<td></td>
<td>361.88</td>
<td>10.30</td>
<td>-68.76</td>
<td>x x</td>
</tr>
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<td>91.32</td>
<td></td>
<td>332.90</td>
<td>7.99</td>
<td>-89.30</td>
<td>x</td>
</tr>
<tr>
<td>77</td>
<td>FCC</td>
<td>94.44</td>
<td></td>
<td>347.73</td>
<td>12.52</td>
<td>-79.28</td>
<td>x</td>
</tr>
<tr>
<td>78</td>
<td>LLLB</td>
<td>95.73</td>
<td></td>
<td>329.91</td>
<td>9.13</td>
<td>-92.21</td>
<td>x x</td>
</tr>
<tr>
<td>79</td>
<td>FRB</td>
<td>97.30</td>
<td></td>
<td>360.69</td>
<td>14.77</td>
<td>-69.55</td>
<td>x</td>
</tr>
<tr>
<td>80</td>
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<td>111.66</td>
<td></td>
<td>333.73</td>
<td>17.63</td>
<td>-92.69</td>
<td>x x</td>
</tr>
</tbody>
</table>

---

**Figure 10.** Histogram showing the distribution of satisfactory anisotropy models at the different stations.
azimuthal anisotropy involves data from several events and a wide variety of stations. Stations recording many events show evidence for models spanning several types of anisotropy. For our restricted source-receiver geometry, this would not be likely if upper mantle anisotropy under the station were responsible for the observations, although complex interplay with initial polarizations must be considered.

On the source side, a strong correlation between events and type of best fitting model is apparent. The correlation is even more striking if we look at the S wave polarization. For our data set, events with a focal mechanism producing SH and SV with the same polarity are primarily associated with eastward dipping TTI models, whereas the few events showing opposite polarities are associated dominantly with westward tilting models. These observations consist of delayed SV components for which the delay gradually transforms into reversed polarity onsets. This correlation prompted us to assess whether anisotropy close to the source could be responsible for the observations of the SV onset reversals.

Our six deep events (500 to 600 km depth) are located below the zone of expected anisotropy in the mantle in general. There is further evidence for the absence of anisotropy below 500 km depth in this particular region [Polet et al., 2000]. Our data from the deepest events should therefore not be contaminated by source-side anisotropy. For events in the 100 to 300 km depth range, anisotropy under the slab may produce some source-side splitting. For our data set, source-side anisotropy with fast axes oriented to the east of the source-to-station azimuth, that is N-S to NE-SW, produce waveforms similar to those produced by eastward TTI in D'', whereas NW-SE to E-W oriented fast axes produce waveforms similar to those produced by westward TTI in D''. Our analysis shows that splitting times of about 1–2 s are sufficient to give waveforms similar to those obtained with anisotropic models of D'' at epicentral distances up to about 92 degrees, whereas stronger source-side anisotropy is necessary at the largest epicentral distances. The onset of the synthetic SV waves produced with source-side anisotropy have usually a slightly lower frequency content, due to their longer mantle path after splitting, than those produced by D'' anisotropy.

Our shallow events are located in two main regions. For both of them, anisotropy in the upper mantle around the slab has recently been analyzed using data from permanent as well as temporary seismological arrays. Polet et al. [2000] has analyzed the northern region, corresponding to our records 8 to 37, and Anderson et al. [2004] has studied the southern region, records 69 to 80. The northern region shows the most complex pattern of anisotropy, interpreted as anisotropy with an E-W trending fast axis below the slab and N-S trending fast axis in the wedge above the slab. The E-W trend of the anisotropy below the slab is almost perpendicular to the source-to-station direction in our data, and should therefore contribute to delay the SV waves compared to the SH waves, but not to transfer energy between the two components. In addition, if source effects are responsible for our observations of anomalous onsets of SV waves, we expect to observe similar reversals of the early part of the SV component of the SKS wave. This is not observed and most of the SKS waves have a clear simple onset on our recordings. This is particularly noticeable for event 981008, which is one of the few events indicating...
westward TTI anisotropy in D″, and for event 010629, for which the SKS waves present at large epicentral distances do not show anomalies. For this event, source-side anisotropy with fast axis in the NW-SE direction could explain the polarization anomalies observed at short epicentral distances, but not the fact that they do not appear at the largest epicentral distances. Event 991130, which has a depth of only 128 km, has very weak SKS waves. It is therefore not
possible to rule out that source-side anisotropy contaminate these data with a fast polarization in a different direction than the one given by Polet et al. [2000].

[45] Records 69 to 80 also originate from events at rather shallow depth in a region studied by Anderson et al. [2004] further south in the subduction zone. In the vicinity of events 020924 and 930608, they observe strong anisotropy with fast axis in the NE-SW direction and 2 s delay time. They infer that this anisotropy is in or below the slab. This could therefore bias our results toward eastward TTI in D′. The data from these two events show, however, little evidence for splitting. Event 000616 is located in a region where the source-side anisotropy is also large, but with a direction of the fast axis very close to the source-to-station direction for our data. This should contribute to early SV waves, which we have no evidence for in our data.

[46] These considerations suggest that the location of the azimuthal anisotropy reported here is not solely near the sources, and that the data are more readily explained by D′′ anisotropy. The data from the deep events should not suffer from contamination from source-side anisotropy. Ideally, data from intermediate events should be corrected for possible source-side anisotropy, in particular if this introduces large delays between SV and SH waves which cannot be detected by analyzing SKS waves. Although such delays cannot explain our data, their combination with D′′ anisotropy can modify the waveforms in quite an intricate way. Considering the pattern in the S waves and in the SKS waves, there is, however, not enough evidence for a coherent effect of source-side anisotropy, and it is not possible at the present time to apply an appropriate source-side anisotropy correction.

4.4. Geographic Distribution of D′′ Anisotropy

[47] We now turn our attention to the distribution of ray paths and their best fitting models. Figure 12 summarizes the modeling of all records, with the preferred model(s) being indicated for each ray path. The two upper frames show the whole data set. In order to improve clarity, we show on the upper left frame the paths requiring anisotropy and on the upper right frame those which can be fit by isotropic models as well. In order to account for the fact that the data from the deeper events are more robust than those from the intermediate events, the paths which give evidence for anisotropy, and shown in the upper left frame, are shown separately for the intermediate and for the deep events in the two lowest frames. Data from the deeper events clearly favor models with eastward TTI. However, we should keep in mind that most of these data come from three events with very similar locations and focal mechanisms. Taking also into account the data from the more shallow events, the picture gets more complicated. Although there is overlap between paths favoring different models, models with tilt to the east are consistently located toward the southeast of the study area, whereas those with tilt to the west enter D′′ more toward the north or northwest.

[48] There are only three data which are solely fit by westward TTI, and their ray paths are somewhat scattered on the map. They all come from the 30 November 1999 event, which is one of our shallowest events, with depth estimates of 128 km by the National Earthquake Information Center (NEIC) and 138 km for the Harvard CMT solution. At stations INK and DAWY, where the match with westward TTI is very good, the sSKS wave should arrive 9 s after Sdiff for a source depth of 128 km, which would produce interference with the later part of the diffracted wave. If the depth of the event has been overestimated, sSKS and Sdiff could arrive even more closely in time (for a source depth of 108 km, they would arrive simultaneously). There is therefore some uncertainty concerning these data, and a definite conclusion cannot be drawn before the depth and the source-side anisotropy of the event has been confirmed. For the rest of our data, which span a large range of epicentral distances and focal depths, similar interference problems have been ruled out.

[49] While only three data require westward dipping TTI, it is clear that different data prefer different models throughout the study region, and lateral variations in D′′ anisotropy must be present. We have modeled the data using 1-D structures; the results clearly show that the azimuthal anisotropy in D′′ varies at a scale which is smaller than the path length of our data in D′′, which extends up to 45 degrees in epicentral arc. Our analysis is primarily based on SV waveforms. Because of the destructive interference of SV and ScSV at diffracted distances, accompanied by continuous leakage of SV into P energy propagating into the core, SV diffracted waves decrease rapidly with distance into the core shadow. The presence of azimuthal anisotropy results in coupling between SV and SH which can offset the relative amplitude decay of the two phases. The SV wave energy generated by SH-SV coupling early in the diffraction process will have an opportunity to leak into the core more than SV energy generated further along the diffraction path. Thus the early part of the SV waveforms should be more sensitive to the anisotropy distributed along the latter portion of the D′′ path than at its beginning, if the path traverses laterally varying anisotropic structure [e.g., Vinnik et al., 1989; Maupin, 1994]. The separation between preferred anisotropy models is, however, not made clearer by considering the last segment of the diffracted paths in D′′ for each ray path. A better separation of observations is obtained by considering the ray paths which span the 50 km above the turning point for each wave [Garnero et al., 2004a]. It is not clear, however, why the structure close to the turning point and in a rather thin region would affect the waves more strongly.

[50] It is also important to keep in mind that tilts of the symmetry axis directly along the ray path cannot be discriminated from VTI (Figure 7), and that only moderate changes in azimuth are needed to cause variation from reversed onsets to expected onsets, so our data could be accounted for by small fluctuations in azimuth (not flipping back and forth by 180 degrees, but maybe changing by 30 or 40 degrees only).

5. Discussion and Conclusions

[51] This study presents an analysis of a large data set of S and Sdiff recordings, and demonstrates the presence of azimuthal anisotropy in the D′′ layer. Evidence for anisotropy is primarily based on anomalous SV waveform onsets which cannot be explained by simple isotropic or VTI models, but can be explained by models with azimuthal anisotropy, which couples the SH and SV waves.
The data correspond to paths in essentially one source-receiver corridor, and do not have the resolving power needed to uniquely characterize the parameters of the anisotropy. We can distinguish only between two families of azimuthal anisotropies, represented in our analysis by eastward and westward tilted transverse isotropic models. We stress that these two model types are not unique, but they allow us to explore geographical variations in our two parameter (tilt and azimuth) approach.

Detecting azimuthal anisotropy in D″ is based on analysis of small features in the waveforms which can easily go undetected or masked in the presence of noise. It is therefore not surprising that our data set is complex and does not give a very simple picture of D″ anisotropy. We have strongest evidence for eastward tilted type anisotropy, which matches nearly 50% of the total data set. However, there is also evidence for lateral variation of the anisotropy at a scale of a few degrees (i.e., 200–300 km at the CMB).

Our study area corresponds to an anomalously high-velocity region at the base of the mantle, which is sometimes attributed to the remnants of the Farallon plate [e.g., Grand et al., 1997], which subducted in a direction perpendicular to our ray paths [e.g., Lithgow-Bertelloni and Richards, 1998]. This raises two possibilities; either the anisotropy is directly associated with subducted slabs material, or the slab downwelling may induce the development of anisotropy in the D″ boundary layer as a result of shear flow.

Some recent studies suggest that magnesiowustite, likely a major mineral component of the lower mantle, has significant intrinsic anisotropy and a propensity to develop preferred orientation in strongly deformed regions [Yamazaki and Karato, 2002, McNamara et al., 2002] found that the required deformation may be reached in convection models of the mantle for a wide range of rheological parameters, and that LPO is thus likely to occur at the base of the mantle in regions beneath past or present subduction where temperatures are relatively low and shear stresses are relatively high. LPO of magnesiowustite is therefore a plausible candidate to contribute to D″ anisotropy. Using the elastic coefficients given by Karki et al. [1997, Yamazaki and Karato 2002] calculated the total elastic coefficients for polycrystalline samples of magnesiowustite they had deformed under different conditions. Using the elastic coefficients of their sample PI805, which they refer to as representative for samples at lower mantle conditions, we tested whether magnesiowustite LPO could account for the anisotropy required to explain our data.

Magnesiowustite has a cubic symmetry, leading to a variation of the velocities and polarizations dominated by 40 terms, where 0 is the azimuth. For horizontal shear the largest degree of S wave anisotropy in the horizontal plane is for propagation at 0, 90, 180, and 270 degrees from the shear direction. For these azimuths, the strength of anisotropy is about 6% and the fast S waves are polarized nearly as SH, the fast polarization making an angle of 5 degrees with the horizontal plane at azimuths 0 and 180 degrees, and 8 degrees at azimuths 90 and 360 degrees. If the shear strain is not purely horizontal but at an angle of 10 to 20 degrees from the horizontal plane, inclinations of the order of 20 to 30 degrees for fast polarizations will be easily achieved in some directions, leading to azimuthal anisotropy. The strength of anisotropy in those directions would be about 1.5% if D″ is 25% magnesiowustite, assuming the dominant phase, perovskite-structure silicate, does not contribute to the anisotropy at lower mantle conditions. Distributed uniformly over a 400 km thick layer, this magnitude of anisotropy produces the same waveform effects as those presented in this study.

Allowing for the uncertainty still attached to the elastic coefficients of lower mantle materials at very high temperature and pressure conditions, we can state that LPO of magnesiowustite is a reasonable candidate for explaining our observations. The anisotropy of cubic crystals varies more rapidly with azimuth than the anisotropy of hexagonal or orthorhombic crystals. For sample PI805, the fast S waves at azimuths 45, 135, 225 and 315 degrees are dominantly SV, with a polarization direction at 8 and 22 degrees from the vertical direction. The anisotropy for these azimuths is half of that in the directions discussed above, and might be difficult to detect. If LPO of magnesiowustite is the principal cause of D″ anisotropy in regions of paleosubduction, the higher degree of anisotropy in the directions where fast waves are dominantly SH may explain why, on average, SH waves appear faster than SV waves, but SV waves can be faster than SH waves for a few cases. This mechanism could also provide an apparent high degree of anisotropy heterogeneity since small variations in the strain azimuth significantly modify the apparent anisotropy seen by waves propagating in a given direction.

A phase transition has recently been proposed at D″ pressure and temperature conditions for the dominant component of the lower mantle, MgSiO₃ silicate perovskite [Murakami et al., 2004]. This phase transition may be an excellent candidate to explain the discontinuity at the top of the D″ layer, as it results in a high-pressure postperovskite phase with significant anisotropy of orthorhombic symmetry. Initial estimates of the elastic parameters [Murakami et al., 2004] indicate that LPO could develop under shear flow, with the slow axis oriented perpendicularly to the flow and the two fast axes in the flow plane. In a horizontal flow, this would produce anisotropy close to the usual VTI models used in most D″ anisotropy studies, with SH wave velocity higher than SV velocity. Deviations of the flow from the horizontal plane could easily orient the crystals as in our TTI structures. In order to discriminate between LPO in magnesiowustite or LPO in postperovskite material versus other possibilities such as layering or orientation of inclusions, much more extensive ray path coverage will be needed, with a good variety of initial S wave polarizations.

The presence of azimuthal anisotropy in D″ has consequences for interpretation of other S phases, in particular that of SKS splitting. The anisotropy that produces splitting of SKS (or other mantle and core phases) is usually assumed to be located in the upper mantle, or at least above the 660-km discontinuity. The possibility that some of the splitting occurs at much greater depth complicates the analysis significantly [e.g., Hall et al., 2004]. The S waves used in this study have long D″ paths, and they traverse the lowermost mantle with much larger incidence angles than SKS waves. Depending on the precise geometry of the D″ anisotropy parameters, it is possible that any contribution to SKS splitting will be small. This question
can only be confidently answered when the anisotropy orientation is better constrained. This study shows the feasibility of detecting the presence of general anisotropy in the D″ layer. It raises the future possibility of using the orientation of anisotropy in D″ for constraining deep mantle convective flow patterns, as is often done in the upper mantle.

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