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Effects of D″ anisotropy on seismic velocity models of the outermost core

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Abstract. We explore effects of shear wave anisotropy in D″ on seismic waves utilized in constructing models of the outermost core. Reflectivity synthetic seismograms approximating transverse isotropy (TI) in D″ yield absolute and differential time perturbations of *S*, *SKS*, and *SKKS* of up to several seconds, especially if laterally varying anisotropy is considered. Several studies have used anomalies of this magnitude to infer a 0.5-2.5% *P* velocity reduction in the outermost 50-150 km of the core, suggesting a stably stratified layer just beneath the core-mantle boundary. Our data samples the mantle and core beneath Alaska and northern Pacific, a region with strong lateral variations in D″ anisotropy. TI models that predict the magnitude of shear wave splitting seen in the data can account for much (but not all) of the anomalies that isotropic models attribute to outermost core velocity reductions. Future outermost core seismic studies should address such D″ anisotropy effects.

Introduction

The radial seismic velocity structure of the outermost core is best probed using *SKS* waves and their higher multiples, e.g., *SKKS*, *S3KS*, etc., collectively called *SmKS*. These waves traverse the mantle as *S* wave energy but convert to *P* wave energy in the core, possibly with multiple underside reflections from the core-mantle boundary (CMB). *SmKS* differential times, such as *S-SKS*, *SKKS-SKS*, *S3KS-SKKS*, and *S4KS-S3KS*, are usually considered when constraining core structure because they minimize effects of source mislocation and unmapped mantle heterogeneity. The most extensively used phase pairs are *S-SKS* and *SKKS-SKS* (Figure 1a). Mantle *P* waves that penetrate the core such as *PKP*, *PKKP* and their associated branches, are less effective probes of the outermost core because the reduction in *P* velocity (V_P) across the CMB causes downward refraction of the energy, with no energy bottoming in the outer 1000 km of the core.

The possibility of a stably stratified layer in Earth's outermost core has been suggested based on V_P reductions (relative to a homogeneous, self-compressed structure) of 0.5-2.5% distributed over the outermost 50 to 150 km of the core [e.g., *Lay and Young*, 1990; *Souriau and Poupinet*, 1991; *Garnero et al.*, 1993; *Tanaka and Hamaguchi*, 1993]. This is of great importance to the geodynamo and creation of Earth's magnetic field [see *Loper and Lay*, 1995]. The dynamics of such a layer may explain some aspects of long-term secular variation of the field [e.g., see *Braginski*, 1993].

A complication for core investigations which has not received much attention is the presence of seismic anisotropy in the D″ region. Waveform modeling studies have detected shear wave splitting for phases that traverse D″, which in most cases is consistent with the presence of transverse isotropy (TI) [see *Lay et al.*, 1998], the special case of hexagonal anisotropy with a vertical symmetry axis. This could arise from lattice preferred orientation of minerals in D″ or from horizontal lamellae of strongly fluctuating velocity contrasts. The observable effect involves different shear wave velocities for horizontal, *SH*, and vertical, *SV*, particle motions [e.g. *Maupin*, 1994]. All seismic waves that enter the core traverse the D″ region at least twice, and *S(SH)-SKS(SV)* differential times [e.g., *Hales and Roberts*, 1971; *Lay and Young*, 1990] should be affected by anisotropy at the base of the mantle. *S* waves, especially at propagation distances well into the core shadow, can have significant path lengths in an anisotropic D″ layer, and while *SmKS* path lengths are much shorter, their times may also be affected (Figure 1b). We address possible effects of TI in D″ on *S* and *SmKS* times, finding a trade-off between isotropic D″ models with reduced outermost core velocities, and anisotropic D″ models lacking strong outer core reductions. A well-studied multi-phase data set sampling D″ beneath Alaska and the north Pacific is compared to synthetics to quantify the trade-offs.

Synthetic Seismograms

Synthetic seismograms are constructed using the reflectivity method for layered isotropic media [e.g., *Müller*, 1985]. Following the formulation of *Backus* [1962], we approximate a TI medium with a stack of alternating thin layers of contrasting isotropic properties. This method is valid when lamellae thicknesses are much less than the seismic wavelengths of interest. The laminated structure results in distinct effective horizontal and vertical *S* and *P* wave velocities (V_{SH}, V_{SV}, V_{PH}, V_{PV}), i.e., TI (Figure 2). We do not consider *P* wave anisotropy or variations in the parameter η, which affects non-horizontally propagating shear energy (η relates 3 of the 5 independent elastic coefficients that uniquely describe TI). Increasing values of η will decrease V_{SV}. Our purpose here is to address contamination of core phases from simple TI models, and not to specifically model the mantle phases. Thus the added degrees of freedom of perturbing η and V_P anisotropy are not pursued. Doing so could indeed result in further contamination of *SmKS* data. In the example shown in Figure 2, 5 km thick layers with

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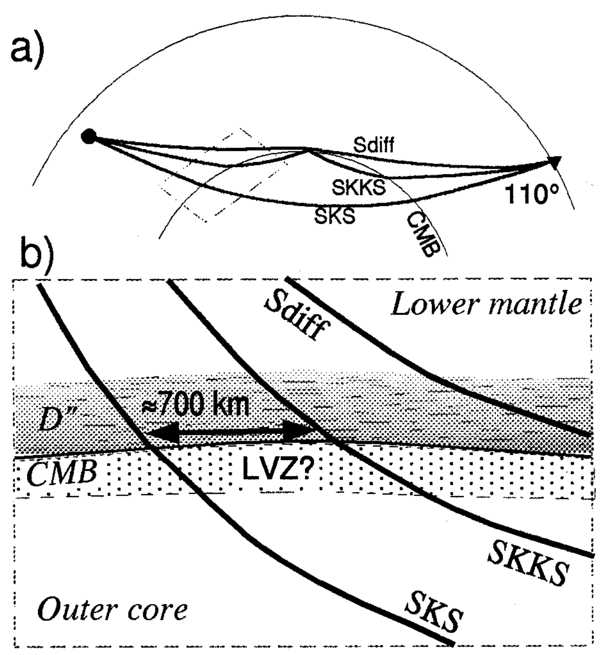

Figure 1. (a) Cross-section of Earth showing raypaths of *SKS*, *SKKS*, and S_{diff}, at 110 deg. Dashed box in lower mantle is shown in (b), which displays D″ anisotropy as shading with horizontal lines. Past studies have imaged the outermost 50-150 km of the core as having reduced Vp gradients (dotted region labeled as LVZ).

V_S alternating between 6.36 and 8.06 km/s give rise to an effective TI medium (Figure 2b) having a difference between V_{SV} and V_{SH} of 2.75%. Thus, thin lamellae with shear velocity differences of over 20% are required to approximate TI of 2.75%, and this proves adequate for modeling long period WWSSN signals (dominant periods of 15 s) quite well. In analyzing seismic energy with wavelengths significantly longer than 5 km, we cannot distinguish between TI caused by intrinsic mineralogical effects or thin lamellae. Therefore, ambiguity in the interpretation of the anisotropy remains. However, for the purpose of this study, the precise mechanism of the anisotropy is unimportant.

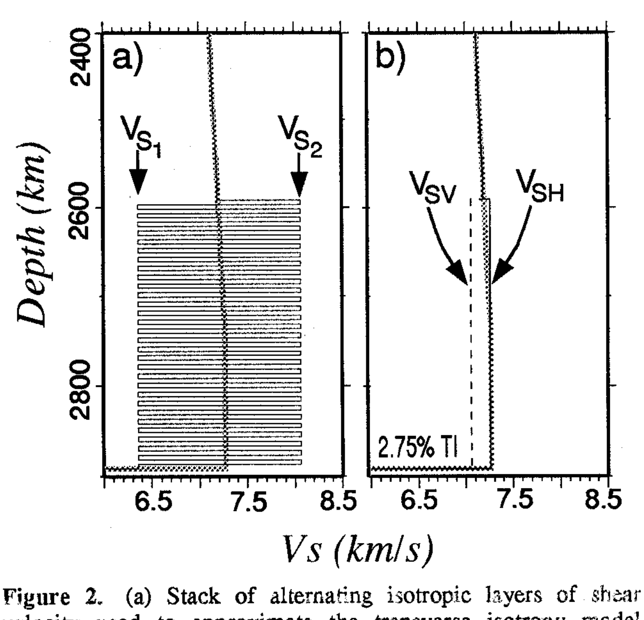

Figure 2. (a) Stack of alternating isotropic layers of shear velocity used to approximate the transverse isotropy model shown in (b). The thicker shaded line is that of PREM. Similar *P* wave fluctuations are necessary to construct the model in (b).

Synthetic seismograms are displayed in Figure 3 for *S* waves at 110 deg for the anisotropic model and the isotropic PREM reference structure [*Dziewonski and Anderson*, 1981] of Figure 2b. PREM is used for the outer core in both cases. As expected, the TI synthetic (solid) shows a delay of the SV_{diff} pulse relative to that of PREM (dashed), whereas the SH_{diff} pulses have peaks that are coincident in time. The lag in SV_{diff} for the TI trace relative to PREM is 10.6 s, which is larger than reported observations (which are typically less than 3 s for diffracted arrivals at this distance). This suggests that D″ anisotropy of this magnitude is not laterally extensive; however, it may be present on localized scales within a laterally varying TI medium [see *Lay et al.*, 1998]. Figure 3b also brings to light an inherent difficulty in quantifying shear wave splits: the SH_{diff} wave shapes are not the same for the two models (due to the strong structural differences). Each of the various methods for quantifying splits (e.g., peak-to-peak times, onset times and waveform correlations) yield differing split estimates. Nonetheless, relative changes in travel time behavior for a given method are easily obtained. We explored a wide range of equivalent TI structures, including TI zones restricted to the top or bottom 75 km of a 300 km thick D″ layer, as well as uniform TI throughout the D″ layer. Magnitudes of TI up to 5% were considered, for various offsets of the V_{SV}:V_{SH} baselines relative to PREM. In general, larger magnitude TI models result in larger splits. Here we do not address the details of the model space explored (owing to the uncertainties involved), and instead focus on some simple TI structures that illustrate the structural trade-off between D″ anisotropy and outermost core V_P.

In Figure 4, predictions from shear wave synthetics for a 300 km thick D″ layer having TI of 1.5%, 2.75%, or 5% are compared to raw (small black dots) and 5 deg distance-bin averaged (larger circles) observations from the data set of *Young and Lay* [1990], which samples D″ beneath Alaska and the northern Pacific. This includes differential times of ScS_{SV}-ScS_{SH} (Figure 4a), SV_{diff}-SH_{diff} (Figure 4b), S_{SH}-SKS_{SV} (Figure 4c), and *SKKS-SKS* (Figure 4d). The first two sets of times provide the evidence for anisotropy at the base of the mantle, while the last two sets provide information about core structure (after possible mantle biases are properly accounted

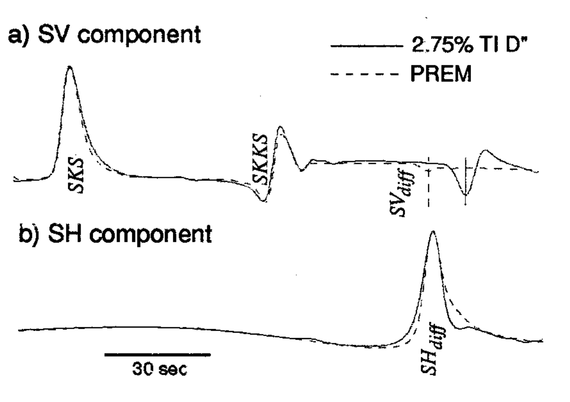

Figure 3. Synthetic seismograms for the TI model of Figure 2 (with η=0.95) and PREM, calculated with the reflectivity method. (a) *SV* component; (b) *SH* component.

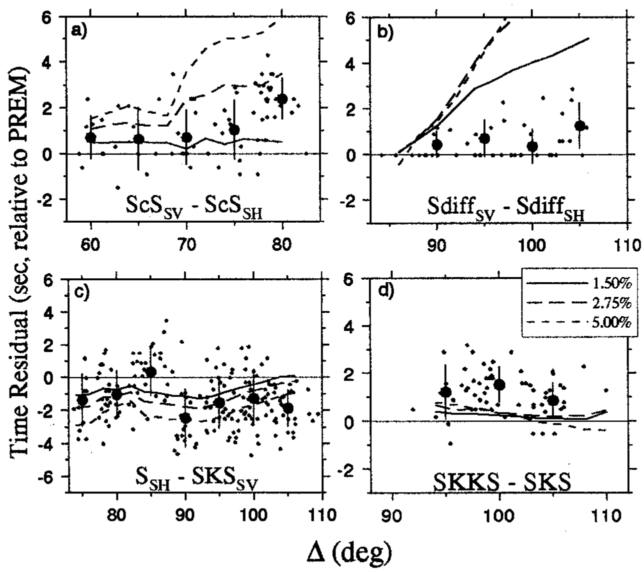

Figure 4. Observed differential travel times (black dots) and distance bin averages (larger filled circles with vertical bars of one standard deviation) for (a) ScS_{SV}-ScS_{SH}; (b) SV_{diff}-SH_{diff}; (c) *S-SKS*; and (d) *SKKS-SKS*, from *Young and Lay* [1990]. Predictions for TI anisotropy of 1.5%, 2.75%, and 5% in a 300 km thick D″ layer are shown by the solid, dashed, and dotted lines, respectively.

for). Differential times for the synthetics were measured in exactly the same way as for the data, with anomalies being computed relative to PREM synthetics. We explored a wide range of TI magnitudes over various D″ thicknesses. Figure 4 demonstrates key points regarding effects of TI in D″: (a) the *ScS* splits (Figure 4a) are well reproduced by TI with magnitudes of 1.5 to 2.75%; (b) the diffracted wave splits (Figure 4b) are overpredicted by all of these models, suggesting that TI of these magnitudes is not uniform along the whole diffracted wave path [*Garnero and Lay*, 1997]; (c) *S-SKS* perturbations (Figure 4c) can be on the order of seconds, with predictions matching the general sign and distance trend of the data averages; and (d) predicted *SKKS-SKS* anomalies are small (< 1 sec), and while shifted in the same direction as the observations, 1-D TI structures do not account for all the anomalous signal in the data.

Outer Core Implications

Previous studies have used *S-SKS* times to infer outer core radial V_P structure once a mantle model (usually isotropic, at least in the lower mantle) is determined. As Figure 4c shows, laterally uniform D″ anisotropy can perturb these differential times by causing relative delays of SKS_{SV} and advances of SH_{diff}, which result in negative differential anomalies. *SKKS-SKS* times can be significantly perturbed by D″ anisotropy if small scale lateral variations in anisotropy are present (*SKS* and *SKKS* raypaths are separated by around 700 km at the CMB) given that the absolute times of *SKS* and *SKKS* vary by 1-3 s for the models considered in Figure 4. For our study region, a 1.5% reduction in the outermost 150 km of the core can be invoked to fit the *S-SKS* and *SKKS-SKS* anomalies, if D″ anisotropy is ignored when determining the reference mantle structure [*Lay and Young*, 1990]. The presence of D″ anisotropy at the base of the mantle in this region [*Lay and Young*, 1991; *Matzel et al.*, 1996; *Garnero and Lay*, 1997] weakens the case for a slow outermost core; but unless there is significant variability in anisotropy at much shorter length scales than in the model of *Garnero and Lay* [1997] (e.g., < 500-1000 km), some *SKKS-SKS* anomaly remains to be accounted for. Thus a reduction of V_P in the outermost core is still consistent with the data. The most important point is that in order to model outermost core structure with these data, the *S-SKS* and *SKKS-SKS* times must first be corrected for D″ anisotropy and its lateral variations. For a D″ layer containing TI, the general tendency is for *SKS* to be slightly delayed relative to PREM. Such delays can alternatively be accounted for by lowering the *P* velocity of the outermost core.

SmKS phase pairs with higher multiple reflections in the core have raypaths containing much closer proximity throughout the mantle, and are longitudinally polarized (SV) while traversing D″. Hence differential measures should be less sensitive to D″ anisotropy (unless anisotropy variations have correspondingly smaller spatial wavelengths). For a region that expands beyond the area studied by *Young and Lay* [1990], a reduction of V_P in the outermost 50 km of the core (-1.5% at the CMB to 0% 50 km below the CMB) was proposed using *SKKS-SKS*, *S3KS-SKKS*, and *S4KS-S3KS* differential times [*Garnero et al.*, 1993]. *S3KS* and *S4KS* raypaths are only separated by 150-200 km at the CMB; their differential times are the strongest constraint on the core model, as they are less dependent on either isotropic or anisotropic mantle structure.

A model that qualitatively explains the *S* and *SmKS* data sets sampling the area below Alaska, contains the following features: (1) anisotropy in the D″ layer that begins at the discontinuity (the top of D″, ≈250 km above the CMB) extending downward, perhaps all the way to the CMB [e.g., *Lay and Young*, 1990; *Matzel et al.*, 1996]; (2) lateral variations in D″ anisotropy [*Garnero and Lay*, 1997]; and (3) a mild V_P reduction in the outermost 50 km of the core [*Garnero et al.*, 1993]. More data are necessary to firmly establish the overall model, which is left for future work, but we note that the majority of published compilations of *SmKS* observations are consistent with such a core model. Our intent here is to demonstrate the need to account for D″ anisotropy effects when modeling outermost core structure.

Discussion and Conclusions

In order to constrain the seismic structure of the outermost core, a variety of travel time effects from mantle structure must first be peeled away. Most work has emphasized development of a reliable 1-D mantle model, but it is clear that this is not adequate for mapping 1-2 s anomalies into anomalous outermost core structure [*Garnero and Helmberger*, 1995]. While this has been understood in general for a long time, the effects of small scale isotropic heterogeneity and lower mantle anisotropy and its variations are now entering the picture. It is noteworthy that no past outer core model (including PREM which utilized the *Hales and Roberts* [1971] observations) has accounted for possible contamination due to D″ anisotropy. While allowing for anisotropy does reduce the case for anomalous outermost core structure made by *Lay and Young*, [1990], it does not eliminate the evidence for a somewhat thinner zone of reduced V_P in the outermost 50 km of the core, as suggested by *S4KS-S3KS* times. Thus, some evidence favoring a stably stratified zone in the outer core persists, although it may be possible to account for the observations with small scale patterns of mantle heterogeneity.

Constraining heterogeneity and magnitude of anisotropy in D″ is very challenging. To even establish the existence of D″ anisotropy, upper mantle anisotropy corrections must first be made (which is certainly not trivial, e.g., *Vinnik et al.* [1998]). Mapping the spatial distribution of any D″ anisotropy requires reliable velocity models for predicting the ray path geometry. The magnitude of TI in D″ is usually inferred using the observed travel time splits along with D″ path length estimates from geometric ray theory (using an assumed reference model). We tested possible inaccuracies in this approach using synthetic seismograms of the 2.75% TI model in Figure 2. SV_{diff}-SH_{diff} splits were measured from the synthetic waveforms and combined with geometric ray path length estimates in the lowermost 300 km of the mantle for PREM, SYLO (from *Young and Lay* [1990]), and M1 (an average model for the central Pacific from *Ritsema et al.* [1997]). The resulting TI strength estimates address the error introduced by using isotropic raypaths and also from using ray paths predicted from different models (Figure 5). The thin horizontal dashed line at 2.75% represents the input model. Several important points emerge: at distances less that 98 deg, ray theory based estimates of the magnitude of anisotropy using the different models poorly predict the starting model, and models PREM and SYLO underpredict the input model throughout the distance range. M1 does better because its negative velocity gradient results in the shortest D″ raypaths and thus highest TI estimate. This result demonstrates the importance of the reference structure, as well as the drawbacks of using isotropic ray theory for anisotropic path length estimation. A more reliable approach is to compare observations directly with TI synthetic seismograms, where synthetic computation is currently viable (i.e., 1-D models).

Our considerations have been limited to the end member case of TI and its effect on shear wave phases, but clearly general anisotropy in D″ can have comparable effects, although possibly with different sign of anomalies. It is important to constrain the precise geometry of mantle anisotropy before reliable corrections can be made. Of the many possible causes of D″ anisotropy [see *Lay et al.*, 1998], the scenario of quasi-horizontal lamellae of strongly contrasting properties, has received particular attention [e.g., see *Kendall and Silver*, 1996]. Such a medium may alter waveforms and travel times of P wave phases, such as P_{diff}, PKP, and $PKKP$, possibly resulting in contamination of P structure studies (e.g., the inner core). Assessing and quantifying such effects is left for future work.

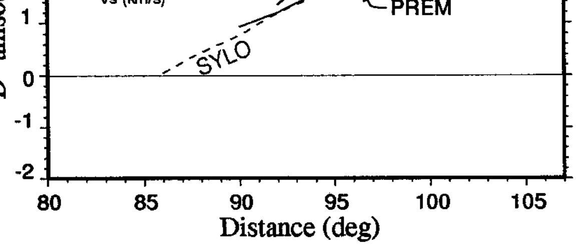

Figure 5. Ray theory based estimates of the magnitude of D″ anisotropy using raypaths calculated for models PREM, SYLO, and M1 (inset) applied to interpret the SV_{diff}-SH_{diff} differential times measured from synthetics for a TI medium of 2.75%.

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