

Upper Mantle Processes and Structures Imaged with
Shear-wave Splitting and P_s and S_p Receiver Functions

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Preface

The interaction between Earth's relatively rigid tectonic plates and relatively viscous convecting upper mantle fundamentally influences myriad processes ranging from arc and intraplate volcanism to orogenesis and rifting. The work presented here addresses two of the most basic concepts of plate tectonics regarding this interaction: (1) the pattern of upper mantle flow in subduction zones, and (2) the thickness and internal structure of continental lithosphere.

Shear-wave Splitting and Upper Mantle Flow

The standard view of deformation in the mantle wedge is simple two-dimensional (2-D) corner flow, where mechanical coupling of the subducting slab to the overlying mantle draws hot back-arc wedge material towards the trench and then down with the plate. Fluxing of volatiles off the cool subducted oceanic crust and up into the warm wedge is thought to be the dominant source of arc magmas, and nearly all numerical and petrological modeling efforts to predict the thermal state of the wedge and explain the location and variation in composition of arc lavas assume this 2-D corner flow model. In addition to controlling the temperature distribution in the wedge, flow and the shear deformation that accompanies it are also responsible for the generation of upper mantle fabric. One manifestation of the preferential alignment of anisotropic mantle minerals (e.g., olivine and orthopyroxene) in this fabric is birefringence, or splitting, of shear-waves passing through the affected region. Under several assumptions, measurements of the fast polarization and delay time of split shear-waves can be and frequently are used to

infer the direction of flow in the upper mantle of subduction zones, as well as in other tectonic settings.

Chapters 1-3 are devoted to the observation, modeling, and interpretation of shear-wave splitting in both local-S and teleseismic SK(K)S phases measured by the Tomography Under Costa Rica And Nicaragua (TUCAN) seismic array in the Central American subduction zone. Both the splitting measurements themselves and the tomographic model derived from them demonstrate the need for significant arc-parallel anisotropic fabric, inferred to indicate arc-parallel mantle wedge flow; trends in arc lava radiogenic isotopes (Appendix A) provide independent evidence for the existence of arc-parallel flow and also constrain its rate to be comparable to that of the subducting Cocos Plate. The primary implications of these results are that (1) the simple 2-D corner flow model and any predictions derived from it (e.g., wedge thermal state, melting, melt migration) are not likely to be accurate in southeastern Central America, (2) the three-dimensional and finite nature of subduction (e.g., slab edges, bends, tears, roll-back) probably plays a significant role in determining the pattern of wedge and sub-slab flow, and (3) the degree of coupling between the subducting plate and surrounding mantle may be lower than expected.

Receiver Functions and Lithospheric Structure

The lithosphere is the Earth's conductive outer thermal boundary layer, and the underlying asthenosphere facilitates the relative motion of the faster-moving tectonic plates and slower-moving deeper mantle. Many physical and chemical mechanisms have been proposed to explain the rheological, seismological, petrological, electrical, and

thermal differences between the lithosphere and asthenosphere, and the estimates of lithospheric thickness from each are often inconsistent, especially in continental regions. Knowledge of both the location and character of the lithosphere-asthenosphere boundary (LAB) is vital to our understanding of plate tectonics and lithospheric evolution; in particular, determining the depth and properties of the boundary itself may help constrain the mechanism(s) responsible for it. At a strong discontinuity in density or seismic velocity (i.e., an impedance contrast), such as is expected at the LAB based on many models, seismic energy will be scattered and produce converted phases (e.g., P_s , S_p), and the analysis of these phases through receiver function imaging has become a widely used means of interrogating the LAB, as well as other upper mantle impedance contrasts.

Chapter 4 presents a receiver function study across the continental United States, northwestern Mexico, and southeastern Canada that uses both P_s and S_p data to estimate the depth of the LAB, as well as other discontinuities within the lithosphere. In the western U.S., a single, clear, relatively large amplitude negative phase in the S_p receiver functions (characteristic of the LAB) is observed at depths of 60-90 km, consistent with the low velocity zone imaged by surface and body wave tomography and interpreted as the LAB. Although smaller in amplitude, similar negative S_p phases, at slightly greater depths (up to 120 km), along much of the Phanerozoic Atlantic margin may also represent the LAB. In contrast, the high velocity lithospheric lid in the Proterozoic central and eastern U.S. extends to at least 200 km, but the most prominent, and often only, negative S_p phase is generally only 60-110 km. Along with other geophysical data (e.g., heat flow, xenoliths, reflection profiles, volcanism), the differences in inferred LAB depth between the receiver functions and surface waves and the observed amplitude of

negative S_p energy indicate that (1) the contrast between lithosphere and asthenosphere in the western U.S. is larger, perhaps due to both elevated temperatures and the presence of partial melt in the asthenosphere, (2) the velocity gradient accompanying the transition from lithosphere to asthenosphere in cratonic North America likely occurs over more than 30 km and can be explained purely thermally, and (3) significant layering within the lithosphere likely exists in the central and eastern U.S.

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My involvement in geology would not have gone further than my wonderful high school geology class, taught by Denny Jordan, without both the Architecture School and Harve Waff at the University of Oregon. Had I been accepted to the Architecture School, I would never have met Harve, who convinced me to come to the Department of Geological Sciences, where my first class with Gordon Goles solidified my desire to study some form of geology. Although I was fundamentally influenced by everyone and

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Many other people, including members of the geology departments at Brown and Oregon as well as others unrelated to geology, have been equally important to me. Despite the inherent risk of offending someone by omission, I will attempt to thank individuals...my Dad, my Grandparents, Cadence, and the rest of my family, which includes Dawn's parents, brothers and relatives, as well as Schaefer; the not necessarily geologically oriented here at Brown – Nancy, Bill, Ruthie, Pauline, Caroline, Lisa, Gloria, Margaret – I couldn't have done it without you guys!); former graduate students Amandine, Marshall, Kate, Nick, Caleb, Mike, Mauro, Rob, Yingjie, Christine, Dayanthie, Sarah Z., T.C., Diana, Noah, Reed; all the current graduate students at Brown; Mike M., Brian B., 'Dave Ben Merre'. , and everyone else involved with HOA; the GCB; Doc Taco; all my friends from Wooster and the Kansas tundra. I would be remiss to not thank food and beer as well.

Although I have thoroughly enjoyed my time in academia, I've been told a different life exists beyond seismology, plate tectonics, and Matlab, and I look forward to becoming familiar with it. However, without the knowledge and perspective gained as a direct result of every one of the aforementioned people, I surely would not fully understand, experience, and appreciate it.

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Chapter 1

Resolving Three-Dimensional Anisotropic Structure with Shear-Wave Splitting Tomography

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Abstract

Shear-wave splitting observations are a commonly used tool for inferring anisotropy and flow within the Earth's interior. Here we present the development and validation of a new technique for imaging anisotropy in the upper mantle using local events: shear-wave splitting tomography. The mantle is parameterized as a three-dimensional block model of crystallographic orientations with the elastic properties of olivine and orthopyroxene, and both orthorhombic and hexagonal symmetries are tested. To efficiently forward calculate splitting, the Christoffel equation is used to progressively split the horizontal components of a synthetic wavelet in each block of the model, and predicted shear-wave splitting parameters are obtained with an eigenvalue minimization technique. Numerically calculated partial derivatives are utilized in a linearized, damped least-squares inversion to solve for a best-fitting model of crystallographic orientations. To account for the non-linear properties of shear-wave splitting, the inversion is applied iteratively and partial derivatives are recalculated after each iteration. A starting model that incorporates information from predicted splitting parameters is found by spatially averaging fast directions and the ratio of observed-to-predicted splitting times. Models from inversions utilizing this average starting model reach lower misfit levels than do inversions with a random or uniform starting model. Modeling results using synthetic data from several anisotropic structures (i.e., sharp lateral and vertical variations in anisotropy) within both an idealized and real (Nicaragua-Costa Rica) subduction zone illustrate the capabilities and limitations of shear-wave splitting tomography. With a station spacing of 25 km in an idealized subduction zone containing uniformly spaced events down to 225 km, both the azimuth and dip of crystallographic axes are resolvable

to a depth of 100-150 km, and lateral heterogeneities in anisotropy on a scale of 50 km at arc and fore-arc distances from the trench are retrieved. Spatial resolution of anisotropy at scales of 75 km is possible further into the back-arc above 150 km depth. The geometry of stations and observed seismicity in the Nicaragua-Costa Rica subduction zone yields partial to good resolution at scales of 50-75 km beneath the fore-arc, arc, and limited regions of the back-arc down to 100 km, and resolution at coarser scales is possible in wider regions beneath the back-arc. Given the distributions of seismic sources within many subduction zones and the advances in broadband seismic array deployments, this new method offers a powerful means with which to constrain the orientation of anisotropic fabric in the upper mantle.

1. Introduction

Shear-wave splitting is a widely used seismological observation that provides a powerful diagnostic of mantle anisotropy. However, the resolving power of splitting measurements can be greatly enhanced through the use of tomographic imaging, which allows for a quantitative characterization of the lateral and vertical distribution of anisotropy, thus moving beyond simplified qualitative interpretations of mantle fabric based on individual splitting measurements. Recent studies have made advances towards anisotropy tomography using splitting intensity measurements from long-period teleseismic SKS waves (*Favier and Chevrot 2003; Chevrot et al. 2004; Chevrot 2006; Long et al. 2008*). We have chosen to focus on a different problem - the tomographic inversion of shear-wave splitting parameters (i.e., fast polarization direction, ϕ , and delay time, dt (e.g., *Silver and Chan 1991*) from local events for three-dimensional (3-D) models of anisotropy.

Shear-wave birefringence (i.e., splitting) observed at the Earth's surface represents the integrated effects of all anisotropic structure encountered by a wave during propagation from source to receiver (or core-mantle boundary to receiver for SKS and related phases). Without any *a priori* assumptions regarding Earth structure, a single splitting measurement does not fully constrain the location, orientation, or magnitude of anisotropy along the ray path. Unlike SKS waves, which reach the surface with near-vertical incidence ($<10^\circ$), local S-waves can produce reliably measured splitting at incidence angles up to 35° (i.e., within the shear-wave window), allowing for greater sampling of anisotropy as a function of incidence angle and greater crossing ray coverage

within a volume of interest. Subduction zones offer an ideal environment to employ shear-wave splitting tomography (SWST) because seismic sources within a subducting plate are widely distributed in depth directly beneath the anisotropic mantle wedge, and with a dense seismic array the differential paths from local S-waves allow for more precise imaging than with teleseismic phases alone.

Several subduction zones manifest substantial variations in splitting fast directions, which in turn suggest the presence of complex anisotropic structure and mantle flow. Local S-wave observations in South America (*Polet et al. 2000; Anderson et al. submitted*), New Zealand (*Audoine et al. 2004; Morley et al. 2006*), Tonga-Fiji (*Smith et al. 2001*), the Marianas (*Fouch and Fischer 1998; Volti et al. 2006; Pozgay et al. 2007*), Ryukyu (*Long and van der Hilst 2006*), Japan (*Iidaka and Obara 1995; Nakajima et al. 2006*), Kamchatka (*Levin et al. 2004*), the Aleutians (*Yang et al. 1995*), and Nicaragua-Costa Rica (*Abt et al. 2006*) show evidence for anisotropy that differs substantially from the predictions of flow driven by motion of a subducting plate (i.e., simple 2-D corner flow in the mantle wedge), in particular if the fast symmetry axis of olivine is assumed to be parallel to the flow direction (i.e., A-type dislocation creep)(e.g., *Zhang and Karato 1995*). In addition, many SKS studies have revealed arc-parallel fast directions in subduction zones (*Russo and Silver 1994; Fouch and Fischer 1996; Polet et al. 2000; Peyton et al. 2001; Anderson et al. 2004; Long and van der Hilst 2005; Huang et al. 2006*) while others display fast directions more consistent with 2-D entrained mantle flow (*Currie et al. 2004*). However, the interpretation of SKS fast directions is problematic because of uncertainty as to whether splitting in these phases primarily reflects anisotropy

above or below the slab. Better constraints on shallow anisotropy from local S-wave tomography will help to resolve this ambiguity.

Two problems complicate the interpretation of local shear-wave splitting in subduction zones in terms of mantle flow. First, most paths sample the slab, mantle wedge, and upper plate, making it difficult to simply correlate observed splitting with anisotropy in any one of these regions. Second, the relationship between strain (i.e., flow) and lattice-preferred orientation (LPO) in mantle minerals has been shown experimentally (*Bystricky et al.* 2000; *Jung and Karato* 2001; *Holtzman et al.* 2003; *Mainprice et al.* 2005; *Jung et al.* 2006; *Katayama and Karato* 2006) and in field observations (*Mehl et al.* 2003; *Mizukami et al.* 2004) to be dependent on pressure, temperature, melt and water content, and stress. Resolving the 3-D distribution of anisotropy will allow more informed interpretations regarding deformation processes and strain geometries in the wedge versus the slab and upper plate.

The inverse problem posed in SWST is similar to isotropic velocity tomography in many respects. However, a non-linear dependence exists between the orientation of anisotropy and the polarization and propagation direction of shear-waves. Resolution of anisotropic structure in spite of this non-linearity is the primary focus of this investigation, and we address the problem with the numerical calculation of partial derivatives applied in an iterative, damped, linearized inversion. We present the development of this approach to anisotropy tomography and validate it using a variety of subduction zone examples. Although we apply the method to local S-waves, it can be easily extended to include SKS splitting given assumptions about the maximum depth of anisotropy in the mantle.

Because shear-wave splitting observations require shear-wave particle motions and high signal-to-noise ratios, quality splitting measurements are observed far less frequently than are P and S travel-times. Thus, both the quantity of data available and the spatial coverage offered by shear-wave splitting measurements are significantly reduced relative to velocity tomography. We have therefore conducted inversions using synthetic data from two event/station distributions: (1) an idealized (i.e., uniformly-spaced) distribution of events and stations within an artificial subduction zone and (2) a real distribution of paths in a shear-wave splitting dataset from the Central American subduction zone (*Abt et al.* 2006). In *Hoernle et al.* (2008), as well as a forthcoming manuscript, we present the application of this method and its results to actual shear-wave splitting observations from the TUCAN (Tomography Under Costa Rica And Nicaragua) seismic array in Nicaragua and Costa Rica. The specific objective of this paper is to demonstrate the ability of our combined forward modeling and inversion approach to retrieve different geometries of anisotropic structures in subduction zones.

Following the discretization of a model space (Section 2.1) and definition of how the anisotropic Earth will be represented (Sections 2.2-2.4), the process of testing the shear-wave splitting tomography method begins with the creation of a synthetic (i.e., “known”) structure (Section 5.1) that we hope to recover by inverting synthetic observations calculated using a forward model of shear-wave splitting (Section 3.1). Synthetic splitting parameters from the known model are determined, a starting model which differs from the known model is then defined (Section 4.1), and predicted splitting parameters are calculated from this starting structure. A linearized least-squares inversion (Section 4.2) is used to iteratively find changes to the starting model that result

in a new model which yields predicted splitting parameters that more closely match those from the known structure. Incomplete sampling of the model (common in the real world and to the examples explored here) does not allow for the actual known model to be recovered exactly, but, rather, forces us to accept a best-fitting model. Several aspects of the inversion process can be manipulated to increase the resolution of model parameters and more accurately retrieve the known structure (Sections 4.3 and 5.2). After illustrating the strengths of SWST with several examples (Sections 5.5-5.6), we discuss the scales of structures that are resolvable with the event-station distributions tested here and compare this method with other approaches to imaging anisotropy (Section 6).

2. Model Parameterization

2.1. Model Space

The model space is parameterized into a three-dimensional array of blocks, with each block having a uniform crystallographic orientation and strength of fabric alignment throughout. The boundaries of the model space are defined to contain all sources and stations, although the method could easily be generalized to include SKS and related core phases. In this paper we apply our method to two subduction zone cases. The first is an idealized subduction zone with a station spacing of 25 km in the along-arc and across-arc directions and a surface containing events that represents the top of a subducting plate extending to a depth of 225 km and dipping at either 45° or 60° (Fig 1). In the second case, sources and stations correspond to the shear-wave splitting dataset obtained from

the TUCAN Broadband Seismometer Experiment (*Abt et al.* 2006) in the Nicaragua-Costa Rica subduction zone (Fig. 2); sources extend down to 220 km. The TUCAN experiment was a 20-month deployment that featured a relatively dense array (10-50 km station spacing) designed specifically to image the mantle wedge.

We have tested inversions using a variety of block sizes, and here present results for a cube with sides of 25 km (i.e., a 25^3 km³ block). Although smaller blocks would allow for smaller-scale variations in anisotropic structure and possibly yield a better fit to the data, the constraints on parameters in each block would decrease while computation time would increase. On the other hand, a larger block size may not allow for enough small-scale variation, leading to a poor fit with the data. There are regions, however, in both models (idealized and real) in which ray coverage is sparse, and in these cases it is advantageous to consider non-cubic volumes composed of more than one 25^3 km³ block. The generation of model volumes larger than individual blocks is described in Section 4.3, and the effects on model resolution are discussed in Section 5.4.

2.2. Mineralogy and Elastic Coefficients

We assume a mineralogy of 70% olivine and 30% orthopyroxene (opx) and use the elastic constants of Forsterite ($\text{Mg}_{1.8}\text{Fe}_{0.2}\text{SiO}_4$) (*Anderson and Isaak* 1995; *Abramson et al.* 1997) and Bronzite ($\text{Mg}_{0.8}\text{Fe}_{0.2}\text{SiO}_3$) (*Frisillo and Barsch* 1972), including their pressure and temperature derivatives (Table 1). The olivine coefficients at the surface are the average of those from *Anderson and Isaak* (1995) and *Abramson et al.* (1997), while the pressure derivatives are from *Abramson et al.* (1997) and the temperature derivatives are from *Anderson and Isaak* (1995). *Frisillo and Barsch* (1972) provide both pressure

and temperature derivatives for orthopyroxene. We assume a temperature profile based on a 30 km thick conductive upper layer, roughly approximating the lithosphere, which overlies an adiabatic mantle with a temperature of 1200°C at the base of the lithosphere. We have compared predicted splitting from this combination of elastic coefficients to predictions from other studies of the elastic properties of olivine (e.g., *Graham and Barsch* 1969; *Kumazawa and Anderson* 1969; *Ismail and Mainprice* 1998) and orthopyroxene (e.g., *Kumazawa* 1969; *Chai et al.* 1997). We find that values are typically the same for f and within 2-5% of each other for dt . The inclusion of pressure and temperature derivatives has a larger effect on delay time (6-10% difference) than on fast direction (almost none). At crustal depths, olivine-opx elastic coefficients are not, in general, appropriate. For simplicity, we retain these coefficients throughout the model space, recognizing that within crustal regions they should be taken as a proxy for more likely sources of crustal anisotropy (e.g., stress-induced cracks or deformation fabrics in crustal mineralogies).

2.3. Crystallographic Symmetry

Naturally deformed lithospheric peridotites (e.g., *Mainprice and Silver* 1993) often exhibit an LPO in which the a -axis [100] of olivine is aligned parallel to the c -axis [001] of opx, the b -axis [010] of olivine is parallel to the a -axis of opx, and the c -axis of olivine is parallel with the b -axis of opx. Proportionally adding the stiffnesses ($0.7 * c_{ijkl}^{olv} + 0.3 * c_{ijkl}^{opx}$) in this orientation provides the elastic properties for our orthorhombic model mantle material. For simplicity, from here on the crystallographic axes mentioned will be those of olivine.

Although olivine and orthopyroxene are each intrinsically orthorhombic, naturally deformed rocks and deformation experiments have demonstrated that crystal aggregates may develop overall symmetries that are hexagonal or orthorhombic (*Ismail and Mainprice* 1998; *Bystricky et al.* 2000; *Mehl et al.* 2003; *Jung et al.* 2006; *Michibayashi et al.* 2006). We have therefore carried out inversions assuming both orthorhombic and hexagonal elastic coefficients. In order to simulate hexagonal symmetry using the orthorhombic single-crystal elastic constants of olivine and orthopyroxene, the coefficients corresponding to the b - and c -axes must be combined into an “isotropic plane” perpendicular to the a -axis. We reduce the nine independent orthorhombic coefficients to the five independent hexagonal coefficients following *Montagner and Anderson* (1989). In the event of wave propagation parallel to the hexagonal symmetry axis, no splitting will occur and the polarization direction will simply be that of the wave’s initial polarization. However, for the cases examined here, the impact of this effect is negligible. Predicted fast directions for hexagonal and orthorhombic coefficients are typically nearly identical. Differences in splitting times between the two symmetries vary as a function of propagation direction and polarization, and while in general they are small, they may rise to roughly 25% in certain cases.

2.4. Model Parameters

Crystallographic orientation is defined by three rotation angles: θ , ψ , and γ (Fig. 3). θ represents the horizontal azimuth of the $[100]$ axis, ψ is the plunge of the $[100]$ axis from horizontal, and γ is the plunge of the $[001]$ axis. In the case of hexagonal symmetry, only the azimuth (θ) and dip (ψ) of the $[100]$ axis (i.e., the symmetry axis) are needed, but for

orthorhombic symmetry, the third angle (γ) is required to completely orient the three orthogonal axes. Note that in the inversions discussed in Sections 5.5 and 5.6 we use only two angle parameters (θ and ψ). For the cases where orthorhombic symmetry is used, the c -axis of olivine is assumed to remain horizontal (i.e., $g=0$).

Typical observed splitting times are much smaller than those predicted directly from the single-crystal elastic coefficients, most likely due to misalignment of crystals in real Earth volumes. We define a strength parameter, α , which is simply a scalar that simulates this dilution of anisotropy. Here α is not the same as the percent of shear-wave anisotropy, δV_S , often referenced in work involving seismic anisotropy (i.e., $\delta V_S = 2(V_{Sf} - V_{Ss}) / (V_{Sf} + V_{Ss})$) from *Ismail and Mainprice* (1998), where V_{Sf} is fast S-wave velocity and V_{Ss} is slow S-wave velocity). Differences in the amount of seismic anisotropy may be caused by variations in the percentage of oriented grains within a volume of mantle, but δV_S also depends on the direction of wave propagation.

3. Forward Modeling

3.1. Calculation of Shear-wave Splitting Parameters

To enable the rapid calculation of predicted shear-wave splitting parameters for each path, we use the approximate particle motion perturbation method of *Fischer et al.* (2000), only here we allow for 3-D distributions and orientations of anisotropy. The effects of anisotropy in each block encountered along a ray path are approximated by progressively rotating and time-shifting the horizontal components of a simple wavelet

that has an initially linear horizontal particle motion (Fig. 4a). The non-linear process of splitting is dependent on the order in which anisotropic structure is encountered by a wave, and the method we use accounts for this behavior. Although parameterized differently, this approach is equivalent to the complete (frequency-dependent) splitting operator of *Rümpker and Silver (1998)*. We use a period of 1.5 s, which represents the typical frequency of local S-waves in the Central American subduction zone (*Abt et al. 2006*). Initial polarizations for cases using the idealized event/station distribution are randomly generated, while for the real event/station locations we use the initial polarizations estimated from actual observations (*Abt et al., 2006*). These real initial polarizations are given by the roughly linear horizontal particle motion obtained after correcting for the best-fitting pair of splitting parameters (Fig. 4a, bottom right). As illustrated by the difference between the corrected linear particle motion and the input initial polarization, this measure of initial polarization is not exact, but represents a reasonable approximation when source mechanisms are unknown.

By tracing each ray through the 1-D velocity/density model AK135 (*Kennett et al. 1995*), path lengths and direction cosines for each ray segment are obtained. Although we are solving for a three-dimensional anisotropic model, these ray paths are utilized throughout the inversion. This simplifying assumption mildly affects ray directions and path lengths within model blocks, but given the relatively short total paths and large model blocks in the cases studied here, in general does not change which blocks a ray samples. Splitting parameters calculated with the 1-D ray paths and the particle motion perturbation method are typically very similar to those calculated with a full waveform method (Section 3.2), suggesting that overall ray bending effects are small.

Polarization directions and phase velocities for each ray segment are given by the eigenvectors and eigenvalues, respectively, of the Christoffel matrix (m_{ij}) (e.g., *Babuska and Cara* 1991)

$$(1) \quad m_{ij} = \frac{1}{\rho(z)} \cdot c_{ijkl}(z) \cdot \hat{n}_k \hat{n}_l$$

where ρ is density and a function of depth z (i.e., pressure and temperature), c_{ijkl} is the elastic stiffness tensor of the olivine-opx crystal (also a function of P and T), and $\hat{n}$ is the propagation direction of the ray. With the exception of the case of hexagonal symmetry and propagation along the symmetry axis, m_{ij} will possess three independent eigenvectors, characterizing the polarization direction of one compressional (P) and two shear (S_{fast} and S_{slow}) waves. In the splitting calculation, fast shear-wave polarization direction in a particular block, Φ , is the horizontal projection of the actual S_{fast} polarization direction in that block (we use F here to distinguish it from the measured ϕ at the surface). Phase velocity is the square root of the eigenvalue corresponding to the eigenvector for that phase, and birefringence of the two shear waves results from the difference in their velocities ($V_{\text{fast}} - V_{\text{slow}}$). The time shift, Δt , for a ray passing through the i^{th} block is

$$(2) \quad \Delta t = L_i (V_{\text{slow}}^{-1} - V_{\text{fast}}^{-1}) \cdot \alpha_i$$

where L_i is the path length in the i^{th} block, V_{fast} and V_{slow} are the fast and slow shear-wave velocities, respectively, and α_i is the strength of anisotropy (i.e., % alignment).

In the first block, the two horizontal components of initially linearly polarized synthetic particle motion are rotated by Φ , time-shifted in the frequency domain by Δt (fast component by $-\Delta t/2$; slow by $+\Delta t/2$), and then rotated back to the original

coordinate system. The resulting perturbed (e.g., elliptical) particle motion is then used as the input for the second block. After applying this procedure successively through each block touched by a particular ray, shear-wave splitting parameters are determined from the resulting horizontal particle motion at the surface using the eigenvalue minimization method (e.g., *Silver and Chan* 1991). An example of the forward calculation and resulting waveforms is shown in Figure 4.

The eigenvalue minimization technique attempts to linearize the horizontal components of particle motion by performing a grid search over all possible azimuths (from -90 to 90 at 1° increments) and a range of splitting times (from 0 to 5 seconds at 0.01 second increments). For each trial combination of parameters, the horizontal components are rotated and time shifted, and their correlation matrix is calculated. If the resulting particle motion were perfectly linear, this matrix would be singular; it would have only one non-zero eigenvalue, and its eigenvector would correspond to the direction of linear particle motion. For uniform anisotropy with a horizontal fast symmetry axis, the splitting fast direction is parallel to the fast axis. When a wave samples multiple regions of anisotropy with different orientations, for example two layers (e.g., *Silver and Savage* 1994) or the three-dimensional cases examined here, the ϕ and dt determined through this method are only apparent splitting parameters that reflect the integrated effects of all anisotropic structure sampled by the wave.

In a process identical to that used for real observations, it is possible to quantify confidence limits for the synthetic splitting parameters calculated here. Each point in (ϕ, dt) -space is assigned the second (smaller) eigenvalue (λ_2) of the corresponding

correlation matrix, and these values are contoured around the minimum (λ_{2min}), with the 95% confidence region given by

$$(3) \quad \frac{\lambda_2}{\lambda_{2min}} \leq \left[1 + \left(\frac{f}{d-f} \right) \cdot F(f, d-f, 95\%) \right]^{1/2}$$

where f is the number of free parameters (in this case 2: ϕ and dt), d is the number of independent pieces of information, and $F(f, d-f, 95\%)$ is the F distribution for 95% confidence (*Draper and Smith* 1966). The number of independent pieces of information, d , is a function of the sampling rate, dominant frequency, and amplitude of both the signal and noise; d is calculated by dividing the waveform window length used in each splitting calculation (10.24 sec here) by the minimum resolvable time (w), and w is the first zero crossing in the auto-correlation of the initial waveform. In a real waveform, the auto-correlated window would contain background noise. Errors (s_ϕ and s_{dt}) are measured on the 95% confidence contour at the maximum distance in the ϕ and dt directions, independently, from the best-fitting pair of splitting parameters (Fig. 4b).

3.2. Full-waveform Comparison

The predicted splitting parameters calculated with the particle motion perturbation method (simple wavelet) have been compared with those obtained from full synthetic waveforms generated using the pseudospectral approach of *Hung and Forsyth* (1998). Figure 5 shows results from two different models for which this comparison has been made. These models are divided into two regions, each with uniform elastic properties, which are separated by a vertical boundary. An event at 150 km depth is located 15 km to the right of the boundary. Wave propagation is three-dimensional and we calculate

splitting using the horizontal components of motion. Splitting is shown (Fig. 5) along a line of surface stations normal to the vertical boundary and located at 80 km from the source in the boundary-parallel direction.

In the first model (Fig. 5a), the left side of the boundary is isotropic, and the right side has a horizontal hexagonal symmetry axis oriented parallel to the y -direction. The anisotropic elastic coefficients have a 5% contrast between the fast and slow shear-wave velocity. Both the fast direction and magnitude of delay time measured with the two methods agree very well from one side of the boundary to the other, with the ϕ being identical (within error) and a maximum dt misfit of only 0.11 s. In the second model (Fig. 5b), 5% anisotropy exists on both sides of the boundary; while the right side is the same as in the first model, the symmetry axis is parallel to the x -direction on the left side, resulting in a larger effective velocity contrast at the boundary. On the right side (same side as the event), there is a good agreement in both ϕ and dt between the two methods. Moving across the boundary, the discrepancy between the two methods becomes larger (when accounting for measurement errors, up to 16° and 0.26 s).

Overall, the simple wavelet method produces a more rapid shift in splitting times across the boundary. Given that the pseudospectral synthetics accurately reflect finite Fresnel zones and the simple wavelet method does not, this result is not unexpected. In the full-waveform case, interaction of the wavefront with the boundary results in some waveform distortion and possible ray bending, leading to the observed differences in measured splitting parameters. These effects are particularly evident in the case with 5% anisotropy on either side of the boundary (Fig. 5c). However, an infinitely sharp contrast in velocity structure (i.e., crystallographic orientation) is not likely to exist in the mantle,

meaning the magnitude of these effects may be overestimated here relative to real datasets. Typical contrasts in anisotropic structure in the models in this paper fall between the two examples in Figure 5. Therefore, we believe the assumptions made in our forward calculation do not significantly undermine the ability of this method to capture the effects of anisotropy on shear-wave particle motion.

4. Inversion Method

4.1. Starting Models

An essential step is the choice of a starting model from which changes will be made to improve the fit between predictions and observations (in this case synthetic observations). We use three different starting models to test the dependence of the final model on this decision, and qualitatively, similar features resolved by inversions with different starting models can be taken as robust. (1) In the first model, we use an averaging technique that, for each block, fixes the a -axis in the horizontal plane (b -axis vertical in the orthorhombic cases) and assigns the azimuth of the a -axis to be the average fast direction ($\bar{\phi}$) of all rays that touch that particular block (Appendix A). The f value for each ray in this case is calculated from a “known” distribution of anisotropy using the forward calculation described in the previous section. Section 5.1 contains a complete explanation of the known structures. This approach differs from the spatial averaging method of *Audoine et al. (2004)* in that the averaging of fast directions is applied over a 3-D, rather than 2-D, model space. The strength of anisotropy in this “averaged” model

is determined with an averaging scheme that utilizes predicted splitting times per kilometer of ray path (Appendix A). In this way, we produce a 3-D starting model that assumes a direct relationship between horizontal a -axis orientation and fast directions, as well as between ray path length and strength of anisotropy. The average starting model for a structure containing a vertical sheet of arc-parallel a -axes surrounded by arc-normal a -axes within the idealized subduction zone is shown in Figure 6; there is a clear visual correspondence between the known structure (Fig. 6a), the predicted splitting parameters (Fig. 6b), and the average starting model (Fig. 6c). (2) In the second starting model, crystal orientations and strengths of anisotropy are randomly generated for each model block (Fig. 7a). (3) The third type of starting model is one with a uniform crystal orientation and strength of anisotropy throughout (Fig. 7b); we choose horizontal a -axes oriented oblique to both arc-parallel and arc-normal (N-S in the geographic coordinate system used for the Central American event and station locations) and a strength of 25%.

4.2. Least-Squares Inversion

The inversion of shear-wave splitting parameters for anisotropy is accomplished by iteratively solving for the orientation and strength of anisotropy that minimize misfit with the synthetic observations through a linearized, damped, least-squares approach (Tarantola 1987)

$$(4) \quad M^{iter+1} = M^{iter} + C_{mm} G^T [G C_{mm} G^T + C_{dd}]^{-1} \delta D^{iter}$$

where changes to a model, $\mathbf{M}^{iter}$, are made to produce a new, better-fitting model, $\mathbf{M}^{iter+1}$.

$d\mathbf{D}^{iter}$ is the data misfit, and $\mathbf{G}$ is a matrix of partial derivatives, which are assumed to be linear with respect to the current model ($\mathbf{M}^{iter}$). The form of $\mathbf{G}$ is

$$(5) \quad \mathbf{G} = \begin{bmatrix} \frac{\delta\phi_1}{\delta\alpha_1} & \frac{\delta\phi_1}{\delta\theta_1} & \frac{\delta\phi_1}{\delta\psi_1} & \dots & \frac{\delta\phi_1}{\delta\alpha_m} & \frac{\delta\phi_1}{\delta\theta_m} & \frac{\delta\phi_1}{\delta\psi_m} \\ \frac{\delta dt_1}{\delta\alpha_1} & \frac{\delta dt_1}{\delta\theta_1} & \frac{\delta dt_1}{\delta\psi_1} & \dots & \frac{\delta dt_1}{\delta\alpha_m} & \frac{\delta dt_1}{\delta\theta_m} & \frac{\delta dt_1}{\delta\psi_m} \\ \vdots & \vdots & \vdots & \ddots & \vdots & \vdots & \vdots \\ \frac{\delta\phi_n}{\delta\alpha_1} & \frac{\delta\phi_n}{\delta\theta_1} & \frac{\delta\phi_n}{\delta\psi_1} & \dots & \frac{\delta\phi_n}{\delta\alpha_m} & \frac{\delta\phi_n}{\delta\theta_m} & \frac{\delta\phi_n}{\delta\psi_m} \\ \frac{\delta dt_n}{\delta\alpha_1} & \frac{\delta dt_n}{\delta\theta_1} & \frac{\delta dt_n}{\delta\psi_1} & \dots & \frac{\delta dt_n}{\delta\alpha_m} & \frac{\delta dt_n}{\delta\theta_m} & \frac{\delta dt_n}{\delta\psi_m} \end{bmatrix}$$

with n observations and m model blocks. Here we show the case for 3 model parameters (α, θ, ψ) , which is applicable to both hexagonal and orthorhombic symmetries, if we assume the c -axis remains horizontal in the orthorhombic case.

The synthetic shear-wave splitting calculations described in Section 3.1 are used to establish the dependence of the splitting parameters (ϕ, dt) on the model parameters $(\alpha, \theta, \psi, \gamma)$. By perturbing each model parameter independently, linear partial derivatives that populate the $\mathbf{G}$ matrix for the inversion are computed by finite difference. The effects of changes in crystallographic axis orientation on splitting parameters are most non-linear when shear-wave polarization is linear and coincident with one of the crystallographic axes. Highly non-linear behavior occurs over only a relatively small portion of the entire parameter space and is most prevalent at the onset of the forward problem (i.e., in the first ray path segment) because the initial wavelet is linearly

polarized. After passing through the first anisotropic block, it is quite unlikely that the resulting particle motion will be both linear and parallel to a crystallographic axis in the next block. If perturbations to axis orientations determined from the inversion are small, then these linear partial derivatives should be valid for most combinations of particle motion and LPO encountered in this analysis. Because crystallographic orientation, as well as propagation direction, in each block touched by a given ray determines the resulting particle motion, partial derivatives must be recalculated when significant changes to the model occur. We use a perturbation of 5° for each angle parameter when calculating finite difference partial derivatives for the inversion and recalculate partial derivatives when a particular angle parameter (θ, ψ, γ) changes by more than 1° from where the partial derivative was last calculated. Strength of anisotropy is simply a scaling factor, and we use perturbations of 5% for this parameter and recalculate partial derivatives when a changes by more than 1%.

C_{dd} is a diagonal data covariance matrix containing the sum of the squares of the 1s errors (see Fig. 4b) of both the observed ($\sigma_{\phi,obs}$ and $\sigma_{dt,obs}$) and predicted ($\sigma_{\phi,pred}$ and $\sigma_{dt,pred}$) splitting parameters

$$(6) \quad C_{dd} = \begin{cases} C_{2n-1,2n-1} = \sigma_{\phi,obs}^2 + \sigma_{\phi,pred}^2 \\ C_{2n,2n} = \sigma_{dt,obs}^2 + \sigma_{dt,pred}^2 \end{cases},$$

where n corresponds to the n^{th} splitting measurement in the data set. The *a priori* model covariance matrix, C_{mm} , may have both diagonal and off-diagonal non-zero values. The function of the diagonal elements of C_{mm} is to damp the changes to model parameters between iterations so that the validity of the linear partial derivatives is not compromised,

and the addition of off-diagonal components establishes a correlation between model parameters. The use of C_{mm} is discussed further in the following section.

4.3. Inversion Parameters

There are several variables that can influence inversion behavior (e.g., starting model, damping, smoothing, block size), and, although each and every combination has not been tested, we have tested many. The inversion parameters discussed here and the range of values tested for each are given in Table 2, and the effects of each on data and model fit are discussed in Section 5.4.

A useful means of assessing the convergence toward a best-fitting model in a damped inversion is to relax damping (i.e., increase the *a priori* value of model parameter variance) after a certain number of iterations. If the inversion has reached the global minimum in misfit prior to relaxing damping, then increasing variance will not result in significant changes to the model. In contrast, if the pre-relaxation model simply occupies a local minimum which has been found by virtue of “over-damping”, then the model may change considerably upon damping relaxation. Low variance values may result in over-damping and require too many iterations to converge, while higher values could result in large changes to the model parameters that may violate the assumption of linearity.

In typical data inversions and the cases studied here, certain model regions are under-sampled and anisotropic parameters in individual model blocks will not be well-resolved. With the real event/station distribution from Central America, this occurs in the back-arc where rays sample the back-arc wedge in only one direction, as well as on the fringes of the array. In the idealized subduction zone, ray coverage is substantially lower in the

deep wedge furthest from the trench. We therefore experimented with two approaches that impose a correlation between specific blocks in poorly sampled regions. The goal is to increase the spatial scales over which anisotropy parameters are allowed to vary to the point where parameters (now representing these larger spatial scales) can be meaningfully resolved.

In the first approach, we combine blocks which are individually sampled by very few rays into larger volumes that act as a single block (Figs. 8,9). Including this additional information within the inversion is accomplished by adding equations that force the changes made to the model parameters in each of the original individual blocks included in a volume to be equal (Appendix B). The large volume constraints are added partway through the inversion, once enough iterations have been completed with the original individual blocks to achieve a stable solution and minimize data misfit; this occurs after iteration 25(30) in the idealized (Central American) subduction zone inversions (Fig. 10). At this point, the individual blocks within a volume are assigned uniform values for crystallographic orientation and strength of anisotropy; we use an average for each model parameter weighted by resolution matrix values (Section 5.2) for these parameters from the iteration prior to combining the blocks. For the remaining iterations, the large volume constraint equations are applied and are assigned small *a priori* values of variance to increase their influence relative to the splitting data.

In the second approach, non-zero off-diagonal components are added to C_{mm} to produce spatial smoothing. This type of smoothing is often used to satisfy some *a priori* information regarding the continuity of physical parameters between model blocks (e.g., Nataf *et al.* 1986). That is, a change made to a certain model parameter will generate a

simultaneous change in that parameter in a nearby block. When used, the form of spatial smoothing is Gaussian and applied only to blocks immediately adjacent to a particular block (m_0) and is based on the distance from the center of each adjacent block to the center of the central block (D)

$$(7) \quad C_{m,m_0} = \sigma_{m_0}^2 \cdot \exp\left(\frac{-\Delta^2}{2\Delta_0^2}\right)$$

where $\sigma_{m_0}^2$ is the *a priori* variance of m_0 and Δ_0 is the minimum distance between two blocks (i.e., blocks sharing a face). In general, we do not see any physical reason for a correlation between the different types of parameters used here, and when discussing spatial smoothing, we are referring only to smoothing between like parameters.

5. Inversion Examples

5.1. Target Structures

To test how well the SWST method resolves different geometries of anisotropic structure (i.e., crystallographic orientation and strength of anisotropy) we have applied the method to synthetic data from five target models. These structures differ only slightly between our two source/station distributions: the idealized synthetic subduction zone and the TUCAN splitting dataset from Nicaragua and Costa Rica. In all examples, the majority of the model will possess a certain crystallographic orientation (e.g., dipping arc-normal α -axes), while a portion of the model will be given a dramatically different

orientation (e.g., horizontal arc-parallel a -axes). Specific examples of these models are shown in Sections 5.5 and 5.6.

First, we conducted simple “control” inversions in which the entire known structure has a uniform orientation (Fig. 11). The second type of known structure contains a vertical column of blocks, 50 km by 50 km laterally, with arc-normal (or arc-parallel) a -axes extending from the top to the bottom of the model, representing a rapid variation in crystallographic orientation over a finite lateral extent (Fig. 12). The third structure is an along-strike sheet of arc-normal (or arc-parallel) a -axes one block wide extending down from the volcanic front in the Central American model space and a horizontal distance of 100-125 km from the trench in the idealized model space (Fig. 13). *Holtzman et al.* (2003) have shown that when melt permeability is reduced in deformed peridotite samples, melt concentrates in oriented bands and olivine a -axes within interband crystal lenses tend to orient perpendicular to the transport direction. The “sheet” structure we use is meant to simulate this behavior in the wedge beneath volcanic centers where large melt fractions are most likely. The development of B-type olivine fabric (*Jung and Karato* 2001; *Jung et al.* 2006) in the “cold” wedge corner (*Kneller et al.* 2005) has been proposed as the source of observed arc-parallel fast directions in the fore-arc of Ryukyu and Japan (*Long and van der Hilst* 2005,2006; *Nakajima et al.* 2006). Therefore, the fourth structure has a region of arc-normal (or arc-parallel) a -axes within the shallow wedge corner (Fig. 14). The final structure is used to test the capability of the method to resolve depth variations in anisotropy and is a layer cake model in which a -axes above 50 km and below 150 km are given one orientation (e.g., arc-parallel) and a -axes between 50 and 150 km have the opposite azimuth.

Structures dominated by arc-normal α -axes would be expected for a standard two-dimensional corner flow model with A-type slip in olivine (e.g., Zhang and Karato 1995). Zones with arc-parallel-fast anisotropy would represent perturbations caused by small-scale flow (*Behn et al.* 2007), melt band formation (*Holtzman et al.* 2003), activation of a different slip system (*Jung and Karato* 2001), or simply fossilized anisotropic fabric. In contrast, the same mechanisms which perturb crystallographic orientation relative to a dominant flow-related fabric could just as easily occur in the presence of principally arc-parallel flow, and we have tested all five known structures using both arc-normal and arc-parallel dominant orientations.

5.2. Quantifying Resolution

Determining the regions of the model space in which parameters are well-resolved is key to interpreting inversion results. In linear inverse problems, a straightforward means of investigating the resolution of a model is through the resolution matrix, $\mathbf{R}_M$ (*Tarantola* 1987),

$$(8) \quad \mathbf{R}_M = \mathbf{C}_{mm} \mathbf{G}^T [\mathbf{G} \mathbf{C}_{mm} \mathbf{G}^T + \mathbf{C}_{dd}]^{-1} \mathbf{G}$$

where the components of $\mathbf{R}_M$ are defined in Section 4.2. The diagonal elements of $\mathbf{R}_M$ reflect how well individual model parameters are resolved, while the trade-offs between different model parameters in different locations of the model are revealed by the off-diagonal terms. If a diagonal element of $\mathbf{R}_M$ is exactly one, then the corresponding model parameter is perfectly well resolved; the diagonal element value declines toward zero as resolution of the model parameter decreases.

The meaning of the resolution matrix $\mathbf{R}_M$ is somewhat more complicated in a non-linear inverse problem such as the inversions for anisotropy studied here. Because the partial derivatives may change significantly with each iteration, $\mathbf{R}_M$ should be considered as reflecting the resolution of the solution only relative to the model from the previous iteration. In addition, when the large volume constraints are employed, the resolution matrix will reflect the values of *a priori* variance assigned to the constraint equations, but not the fact that the constraint equations correspond to zero values in the data vector (Appendix B). Nonetheless, the resolution matrix still contains valuable information about how the data sample the anisotropic parameters in a given model. We use the resolution matrix in two ways: 1) to define model blocks that should be combined into larger model volumes and 2) to define which model regions to display. In presenting the resulting tomographic models (Sections 5.5 and 5.6), we restrict the blocks shown to those in which all model parameters within the block have resolution values greater than 0.25 in the resolution matrix from the final iteration. We distinguish between levels of resolution by assigning different thicknesses to the vectors representing the *a*-axis orientation and anisotropy strength in each block.

5.3. Data and Model Misfit

We characterize the fit of both the data and model with a mean, weighted misfit ($\bar{\epsilon}$), and use the behavior of both measures to choose best-fitting models from the different inversions. Data misfits, $[(\phi, dt)^{obs} - (\phi, dt)^{pred}]$, are weighted by both the observed and predicted errors, calculated as in Section 3.2,

$$(9) \quad \bar{\varepsilon}_{(\phi, dt)} = \frac{\sum_{i=1}^n |(\phi, dt)_i^{obs} - (\phi, dt)_i^{pred}| \cdot (\sigma_i^{(\phi, dt), obs} + \sigma_i^{(\phi, dt), pred})^{-1}}{\sum_{i=1}^n (\sigma_i^{(\phi, dt), obs} + \sigma_i^{(\phi, dt), pred})^{-1}}$$

where σ_i^{obs} are the 1σ errors associated with the i^{th} synthetic observed splitting measurement calculated from the known structure, and σ_i^{pred} are those associated with same ray but calculated from a model during the inversion. Typical errors are quite small ($<5^\circ$ and <0.05 sec) because measurements made here are on synthetic wavelets (Fig. 4).

Model misfit is calculated in a slightly different manner from the data, in that we choose to compare only the resolvable portion(s) of the model to the known structure. For blocks in which all model parameters are well-resolved, the mean, weighted strength misfit for each iteration is

$$(10) \quad \bar{\varepsilon}_\alpha = \frac{\sum_{j=1}^m |\alpha_j^{known} - \alpha_j^{model}| \cdot (\sigma_j^{\alpha, model})^{-1}}{\sum_{j=1}^m (\sigma_j^{\alpha, model})^{-1}}$$

where α_j^{model} is the strength of anisotropy in the j^{th} block of the model, m is the number of well-resolved blocks, and the 1σ errors $(\sigma_j^{\alpha, model})$ are given by the square root of the diagonal elements corresponding to α in the *a posteriori* model covariance matrix, C_M , (Tarantola 1987)

$$(11) \quad C_M = C_{mm} - \left(C_{mm} G^T [G C_{mm} G^T + C_{dd}]^{-1} G \right) G C_{mm}.$$

Orientation misfit, $\bar{\varepsilon}_{axis}$, is described by the angular difference between the well-resolved a -axes in a model retrieved with the inversion and those in the known structure, and the weight assigned to each axis misfit is the mean 1σ error for all angle parameters used

$$(12) \quad \bar{\epsilon}_{axis} = \frac{\sum_{j=1}^m (\cos^{-1}(\mathbf{a}^{known} \cdot \mathbf{a}^{model})) \left(\frac{\sigma_j^\theta + \sigma_j^\psi}{2} \right)^{-1}}{\sum_{j=1}^m \left(\frac{\sigma_j^\theta + \sigma_j^\psi}{2} \right)^{-1}}$$

where $\mathbf{a}^{known}$ is the a -axis in the known structure and $\mathbf{a}^{model}$ is the a -axis in the j^{th} well-resolved model block. Note that here, and in the inversion examples shown in Sections 5.5 and 5.6, both the azimuth (θ) and dip (ψ) of the a -axis are solved for, but for inversions where only θ is used, σ^ψ would be zero and the 2 in the denominator of the weight would become 1. Similarly, if γ is solved for, then σ^γ would be added and the denominator would be 3.

5.4. Optimal Inversion Parameters

Through extensive testing, the inversion parameters which most consistently result in the lowest data and model misfits have been determined. The range of values tested and their preferred values are given in Table 2.

Model Parameters: As described in Section 2.4, three angle parameters (θ, ψ, γ) are needed to fully orient the three crystallographic axes of the olivine-orthopyroxene “crystal”, but resolution testing has shown that γ is not as well-resolved as θ and ψ with the limited ray coverage available in Central America, and, therefore, we assume the [001] axis to be horizontal (i.e., $\gamma=0$) and only consider θ and ψ for the inversions with orthorhombic symmetry. This assumption does not affect the inversions using hexagonal symmetry.

Initial Model Parameter Variance: It is found that an *a priori* model parameter variance of 1 provides the best compromise between stability, convergence rate, and number of resolvable blocks. In terms of model parameters, such a value means that we assume each parameter (either an angle or strength) is known within ± 1 (degree or %). Additionally, we find that relaxing damping (variance) to a value of 10 results in a noticeable increase in the number of well-resolved blocks without significantly decreasing inversion stability.

Damping Relaxation: Relaxing damping from an initial value of 5 to a value of 20 results in an increase in the number of well-resolved blocks, but both data and model misfit also increase. Such behavior demonstrates that relaxing to a value of 20 permits model parameter changes that violate the assumption of linearity to a degree that prevents convergence to a stable solution. However, relaxation of damping from an initial value of 1 to a value of 10 produces no significant change in misfit while still increasing the number of well-resolved blocks. Therefore, we choose to use an initial model parameter variance value of 1 and relax damping to 10. For the idealized subduction zone, damping is relaxed after 15 iterations, and for the Central American inversions, relaxation occurs after 20 iterations.

Large Volume Constraints: The generation of larger model volumes from individual model blocks (Figs. 8,9 and Appendix B) allows well-resolved anisotropic parameters to be retrieved in portions of the model where parameters for individual 25^3 km^3 blocks are not as well-resolved (Fig. 10); possible spatial variations in anisotropy are obviously

made coarser in the regions where these constraints are applied. Large volume constraints are introduced at intermediate iterations (once stable solutions for the original individual blocks were obtained). The effect of the large volume constraints can be partially assessed by comparing values of the resolution matrix diagonals for a -axis azimuth between the intermediate iteration just before the large volume constraints are introduced (Figure 10a,c) and the next iteration (Figure 10b,d); this example corresponds to the case of uniform anisotropy with a dipping a -axis (e.g., Figure 11). In back-arc mantle regions where the large volume constraints combine groups of individual blocks, the resolution matrix diagonal for each large volume is larger than those of the individual blocks that comprise it. Note that if a portion of a large volume is untouched by any shear-wave paths, that region is shown as having zero resolution (dark blue).

We find that model structure at the iteration where the large volume constraints are introduced plays a significant role in how well the final retrieved model matches the known structure. In particular, inversions using the average starting model achieve a much lower model misfit than those with either random or uniform starting models. This is not surprising given that the regions of the model space influenced most by adding the large volume constraints are those in which model parameters are not individually well-resolved. Unlike with the average starting model, which is initially similar to the known structure (Fig. 6), inversions with the random and uniform starting models are less able to retrieve the known structure in the more poorly sample portions of the model space. The degree to which the volumes affect the inversion is controlled by the “variance” assigned to the added equations in $\mathbf{G}$ (see Appendix B, equation B4). Although the “data” added with these constraints do not influence the resolution matrix, their associated variance

does. We find that an *a priori* variance of 10^{-1} reduces model and data misfit more than values of 10^{-3} and 10^{-5} . We conclude that using larger model volumes is very valuable in terms of expanding the region where model parameters can be meaningfully resolved, but that these constraints should be applied in conjunction with an average starting model.

Spatial Smoothing: As an alternative to the large volume constraints described in the previous section, we instead applied spatial smoothing through the addition of non-zero off-diagonal terms to C_{mm} for several length scales ($\Delta_0 = 12.5$ to 25 km). Because D_0 is the distance at which the value of C_{mm} drops by $e^{-0.5}$ (see Eq. 7), light spatial smoothing is applied even when Δ_0 is less than the individual block dimensions. For all D_0 , no noticeable improvement in model convergence is observed, and at $\Delta_0 = 25$ km (the distance between facing blocks) a significant increase in data and model misfit also occurs. Overall, this approach was much less effective at extending the region where anisotropic parameters are well-resolved while keeping data misfit low. It was therefore discarded in favor of the large volume constraints.

5.5. Idealized Subduction Zone

We first present several inversion examples from idealized subduction zone models to demonstrate the capability of the SWST method to resolve crystallographic orientations from local shear-wave splitting measurements. Inversions for the target structures described in Section 5.1 are shown in Figures 11-14. Each figure includes the “known” model, the splitting parameters from the “known” model, the starting model and splitting parameters, and the final model and associated splitting parameters. The initial and final

data and model misfits are also given in a table with each figure. The pseudo 3-D figures are produced by projecting the entire 3-D model onto each side of a cube, allowing visualization of the azimuth and dip of each a -axis in both the along-arc and across-arc directions. Overlapping axes are seen because all vectors in the model are displayed. In the cases where a -axes are horizontal and exactly arc-normal (or arc-parallel) no vector will be visible in the along-arc (or across-arc) cross-sections. The location of the trench and projection of the events used to represent the slab surface are displayed in the plot containing the splitting parameters from the target structure (subplot b). Each inversion has an initial model parameter variance of 1 for α , θ , and ψ ; all models have a horizontal c -axis (i.e., $\gamma=0$). A total of 35 iterations are completed, damping is relaxed to 10 after 15 iterations, spatial smoothing is not applied, and larger volumes (Section 4.3) are formed after 25 iterations with a variance of 0.1 (Fig. 10, Appendix B).

All model blocks are shown in the known and initial models, while only well-resolved blocks are shown in the final models. The vectors representing a -axes and splitting parameters are all the same thickness except for subplot e, in which vector width corresponds to the level of resolution (see Section 5.2) of model parameters in a particular block (e.g., thicker = better-resolved). An example of the resolution of two model parameters (α and θ) is illustrated in Figure 10 and is representative of resolution in other inversions; model parameter resolution does not change noticeably between the different inversion examples because propagation direction influences resolution more than model orientation.

The “control” inversion example contains all arc-parallel a -axes dipping at 30° (Fig. 11a), uses the average starting model (Fig. 11c), and assumes hexagonal symmetry. The

initial model is very close to the known structure in terms of a -axis azimuth and strength of anisotropy, but because a -axes are horizontal in the starting model the initial angular misfit is 30.7° . Initial fast direction and splitting time misfit are also quite low. The inversion decreases data misfit significantly (6.83° to 2.58° for ϕ and 0.50 s to 0.12 s for dt), and although strength misfit increases from 9.43% to 16.99% , angular misfit is reduced to 6.90° . Importantly, a -axis dip is accurately recovered nearly everywhere in the model down to 150 km. This example demonstrates the effectiveness of the inversion and provides context to assess the inversions that attempt to retrieve heterogeneous anisotropic structure.

The target structure with a vertical column of horizontal arc-normal a -axes surrounded by dipping arc-parallel a -axes is shown in Figure 12a. In map view, the starting model in this example displays only a slight perturbation in the dominantly arc-parallel a -axes where the column exists and almost no indication of the column in either cross-section (Fig. 12c). Following the inversion, the column is clearly visible in both map view and cross-section (Fig. 12e). Note that the arc-normal a -axes in the column remain nearly horizontal while the dip of the surrounding arc-parallel a -axes is retrieved down to 75 km. There is a noticeable increase in strength of anisotropy misfit, but angular misfit is cut by more than half and data misfit is also reduced.

The next example is for a structure in which dipping arc-parallel a -axes dominate and a one block wide sheet of horizontal arc-normal a -axes extends down from the surface (Fig. 13a). In this case, the dip of the slab is 60° (Fig. 13b) and the symmetry used is orthorhombic. The average starting model here (Fig. 13c) exhibits, in map view, a closer resemblance to the known model than for the column structure (Fig. 12c), but there is no

indication of a vertically continuous feature. The inversion, however, is able to accurately reproduce both the horizontal arc-normal a -axes and dipping arc-parallel a -axes down to depths of at least 150 km (Fig. 13e). Angular misfit reduction is commensurate with the visual agreement between the target and initial/final models and data misfits improve, but again strength misfit increases.

For the wedge corner target structure, we use a dominantly arc-normal structure (dipping) with horizontal arc-parallel a -axes above 100 km and out to 100 km from the trench, resulting in a “notch” of arc-parallel structure in the wedge corner (Fig. 14a). Such a geometry allows us to examine the ability of the inversion to retrieve vertical variations in anisotropy close to the slab. The starting model displays vertically continuous structure that does not reflect the notch in the wedge corner (Fig. 14c); the final model clearly reveals the notch, and the dip of arc-normal a -axes further from the trench is reasonably well-recovered down to 150 km (Fig. 14e).

5.6. Central American Subduction Zone

The inversion results in Figures 15-17 illustrate that the SWST method can also be utilized with less uniform event and station distributions that are more typical of real data. These inversions use the event/station locations from the TUCAN shear-wave splitting dataset in Nicaragua and Costa Rica. They are run for 40 iterations, damping is relaxed from a value of 1 to a value of 10 at iteration 20, no spatial smoothing is used, and larger volumes are generated in less well-resolved regions (Fig. 8) after iteration 30. The variance applied to the equations forcing the correlation between these blocks is 0.1 (see Appendix B). As with the idealized subduction zone, adding the large volume

constraints increases the joint resolution of model parameters within groups of blocks (Fig. 10)

For the “control” inversion (i.e., uniform structure), in this case arc-normal a -axes that dip 30° towards the back-arc (Fig. 15a), we solve for a , q , and y using the average starting model and assume orthorhombic symmetry. [100]-axis azimuths in the starting model match the known structure quite well (Fig. 15c), with the exception of a handful of more arc-parallel a -axes in Costa Rica. However, because the known model contains dipping a -axes, the initial model has an average angular misfit of 31° . A large decrease in data misfit (7.89° to 3.31° in ϕ and 0.68 s to 0.16 s in dt) occurs as a result of the inversion. The well-resolved portion of the model space extends far into the back-arc region, and resolution of a -axis dip is good to depths of roughly 75 km (Fig. 15e). In this case, both angular and strength misfits are reduced (31.27° to 21.41° and 17.81% to 15.23%), although the decrease in angular misfit is not as dramatic as in the idealized inversion (Fig. 11) owing to the less dense sampling of the model.

The first example with heterogeneous structure is for the model with columns of horizontal arc-parallel a -axes surrounded by dipping arc-normal a -axes (Fig. 16a). One column is found in Nicaragua beneath the dense cross-arc line of stations and the other is beneath the Costa Rican cross-arc line. Both columns are visible in the starting model (Fig. 16c), but the column in Costa Rica is smeared in the along-arc direction and the strength of arc-parallel anisotropy is too low in the column in Nicaragua. The columns in the final model are better-defined, and a -axis dip is recovered beneath the fore-arc and arc and down to a depth of 75 km (Fig. 16e). Except for a , both data and model misfit are

substantially reduced: 33.63° to 22.43° for model axis misfit and from 12.52° to 7.6° and 0.52 s to 0.17 s for the data misfit.

The layered structure is not as well resolved as any of the laterally varying structures (columns, sheet, wedge corner), both for the event and station distribution in the Central American subduction zone (Fig. 17) and the idealized subduction zone (not shown). While the final model (Fig. 17e) does not match significant portions of the known structure (Fig. 17a), it is considerably closer to the input structure than the uniform starting model. Despite its deviations from the known structure, the final model does reproduce the predicted splitting parameters quite well (Fig. 17b,f). The difficulty resolving this type of layered anisotropy with SWST is not unexpected given that splitting is the integrated consequence of all anisotropic structure along the ray path. Structures similar to this simple layer cake model would be better resolved with sampling from near-horizontal rays or with a more complete range of initial polarizations at typical incidence angles in subduction zones. Nonetheless, because this inversion example employs the uniform starting model, it illustrates that the SWST method still functions well with starting models farther from the true structure.

6. Discussion

The inversion examples presented in Figures 11-17 clearly demonstrate that the SWST method is able to accurately retrieve heterogeneous, dipping crystallographic orientations in a mantle wedge-type environment. This result suggests that we can use

SWST to image mantle fabric and make interpretations regarding deformation. Here we discuss several aspects of the inversion process and provide rationale for some of the decisions we have made.

Although all three types of starting models have been tested for each structure, we have chosen to show the results from inversions using the average starting model (except in Fig. 17) because they reach lower levels of both data and model misfit and also display a much more consistent behavior in terms of misfit reduction. An important benefit of the average starting model is that the portions of the model space which are sampled less (i.e., deeper areas further from the slab) begin the inversion with a model orientation that results in better-fitting splitting parameters. The random and uniform starting models produce the same structure as the average starting model in the best-resolved areas but are unable to do so in more under-sampled regions because of the non-linearity and non-uniqueness of the problem.

Within the more densely sampled shallow wedge corner of the idealized subduction zone, variations in anisotropy at the scale of 50 km can be resolved to depths of 100-150 km (e.g., Figs. 12,13). Partial resolution at scales of 50-75 km exists beneath the fore-arc, arc, and limited regions of the back-arc down to 100 km in the Nicaragua-Costa Rica models (e.g., Figs. 15,16). Importantly, in both cases, a -axis dip is at least partially recovered in these regions. The well-resolved portion of the model space can be extended to the back-arc in both the real and idealized models by coarsening model block size through the generation of larger volumes (Section 4.3, Figs. 8,9; Appendix B), but in doing so, the scale of resolvable variations in anisotropy increases to at least 75 km. In general, at distances of more than 50 km beyond the volcanic arc in Central America,

lateral resolution is dramatically reduced (Figs. 15-17), and in both subduction zone models (idealized and Central America), retrieval of structure in deeper regions which are illuminated by fewer sources is difficult. Multiple horizontal layers of differently oriented anisotropy are less well-resolved, given the event and initial polarization distributions considered here. However, as demonstrated by the inversion example in Figure 14, this limitation does not preclude the recovery of all vertical variations in anisotropy.

Although data misfit and model angular misfit are systematically reduced in the inversions, model strength misfit (a) increases in many cases. The a misfit increase is possible because changes made to the starting model and all subsequent models are based solely on the misfit of predicted splitting parameters and not on how well a particular model matches the known structure. After the first iteration, both ϕ and dt misfits decrease until leveling out. These data misfits are slightly higher in regions where variations in anisotropic structure are greater (i.e., near the column/sheet/wedge corner). Overall, angular misfit of the a -axis also decreases smoothly. In contrast, a misfits typically increase in the first iteration and then decrease overall with subsequent iterations. We find that strength misfit does not improve for any of the damping values examined (1,5,10) in the testing of model parameter variance.

The strength of anisotropy in the final models is often larger than in the known models, and this result appears to be due to trade-offs between non-uniform anisotropic structure (e.g., alternating arc-parallel and arc-normal a -axes) and strength of anisotropy. That is, if the delay time for a particular ray is greater than that predicted from the known structure, then two types of changes to the model could reduce data misfit equally well:

(1) the overall strength of anisotropy could be decreased or (2) the orientation of a -axes in two blocks sampled by the ray could be made more different, essentially canceling out a certain amount of birefringence for that ray. In other cases, the strength misfit appears to be due to trade-offs in strength with depth. For example, with the uniform anisotropy target structure in the idealized subduction zone (Fig. 11), a -axis orientation is fairly well-retrieved throughout the model, but anisotropic strength is underestimated in the deeper, less sampled portion of the model and overestimated at shallower depths. Most regions in the models are sampled well enough that these effects are minimized, but such trade-offs, coupled with the non-linearity of shear-wave splitting, appear to allow for the existence of multiple data misfit minima in model parameter-space. Overall for the six successful inversion examples presented here (Figs. 11-16), model non-uniqueness contributed to average uncertainties of 15-25% in anisotropic strength. When the subduction zone is uniformly well-sampled (the idealized case) the average a -axis misfit is quite small (7° - 15°), but even in the case of the path distribution from Central America, the average a -axis misfits are less than 25° . These uncertainties represent the average errors only in the models tested here and should not necessarily be taken as a more general measure of the accuracy of this method. Inversions carried out with real data should be interpreted in the context of resolution experiments that employ the paths, initial polarizations, and splitting parameter uncertainties of the particular dataset. Nonetheless, the retrieval of target structures demonstrated here indicates that the SWST approach is capable of capturing broad variations in anisotropy over significant subduction zone volumes.

The SWST method presented here is best-suited to applications in subduction zones because of the fundamental need for sampling of the mantle with waves from different directions (i.e., crossing rays) and the additional depth resolution provided by sources with varying hypocentral depths. This method should also work well in crustal volumes with widely distributed seismicity. In contrast, finite-frequency scattering effects are thought to provide sufficient overlap between the Fresnel zones of near-vertical, long-period SKS waves (periods of 5-15 s) to allow finite-frequency vectorial tomography (FFVT) to be used more widely throughout the mantle (*Chevrot 2006*). The Fresnel zones of short-period local S-waves (periods of 0.5-1.5 s), like those used in SWST, are much narrower than those of SKS waves and overall local-S sensitivity to structure far from the geometrical ray is much smaller.

The SWST and FFVT methods differ in several ways. The data inverted in FFVT are splitting intensity measurements (e.g., *Chevrot 2000*) made over a range of azimuths and not the splitting parameters (f, dt) used in SWST. Splitting intensity (i.e., a measure of transverse energy) is not an exploitable observation in local-S splitting studies because, unlike SKS phases, local S-waves will, in general, have significant transverse energy in the absence of anisotropy simply due to the earthquake source. In principle, the source radiation pattern could be accounted for, but with relatively small events without an identified focal mechanism, as is often the case, the initial polarization of local shear-waves is unknown. In FFVT, the anisotropic mantle is modeled as deviations from a reference homogeneous isotropic elastic model as opposed to using the single-crystal elastic stiffnesses we utilize here. The model parameterization used in FFVT employs a dimensionless anisotropic parameter (γ) to represent the strength of anisotropy (negative

values for a fast symmetry axis, positive for slow) as opposed to our strength parameter (α), which directly scales splitting time. The orientation of the hexagonal symmetry axis in FFVT is parameterized with two angles (θ and ψ), similar to SWST, but in SWST we also allow for the more general orthorhombic case to be modeled as well. Although the inverse problem is solved in the same fashion (*Tarantola* 1987) the forward modeling procedure differs between FFVT and SWST. In FFVT, splitting intensity is calculated by integrating analytical sensitivity kernels for the anisotropic parameters over the model volume (*Favier and Chevrot* 2003; *Chevrot* 2006). In SWST, we progressively split, from source to receiver, a sinusoidal wavelet by rotating and time-shifting the horizontal components of motion (*Fischer et al.* 2000) in accordance with the polarization directions and phase velocities obtained from the Christoffel matrix (equation 1) (*Babuska and Cara* 1991), and then use the eigenvalue minimization method (*Silver and Chan* 1991) to calculate apparent splitting parameters (ϕ, dt).

Both the FFVT method described in *Chevrot* (2006) and the SWST method presented here possess advantages and disadvantages. If the object of study is the anisotropic structure of the upper mantle in subduction zones, then SWST provides a more direct means of interrogating the region of interest. The FFVT method is much more suited to regions where local sources either do not exist or are limited in depth and where an assumption of the lower limit of the anisotropic mantle can be comfortably made. In a broad regional study that includes both plate margins and interiors, a combination of these two analyses would provide a powerful means of constraining both lateral and vertical variations in anisotropy.

In addition to Nicaragua-Costa Rica, studies in several other subduction zones have produced event-station distributions and high quality splitting data sets necessary for the successful application of SWST. For example, New Zealand, the Marianas, Japan, Kamchatka, Alaska, and South America (Central Andes) have all been densely sampled (e.g., *Audoine et al.* 2004; *Levin et al.* 2004; *Nakajima et al.* 2006; *Pozgay et al.* 2007; *Anderson et al.* submitted; *Abers et al.* in prep), and their tectonic settings possess a degree of complexity that would likely generate complex upper mantle flow, making them intriguing locations to image mantle fabric.

7. Conclusions

Shear-wave splitting tomography is a robust method capable of resolving three-dimensional anisotropic structure in subduction zones given dense enough sampling of the mantle wedge with local S-waves. Although aspects of shear-wave interaction with anisotropic structure are inherently non-linear, successful inversions are possible through the iterative application of a linearized, damped, least-squares inverse solution if partial derivatives are recalculated after each iteration. The use of the approximate particle motion perturbation method to rapidly predict shear-wave splitting parameters allows for efficient calculation of partial derivatives. When tested with synthetic data obtained from known target structures, good convergence to accurate and stable solutions is found in inversions with widely varying starting models. However, known target structures are

typically best-retrieved with starting models constructed by averaging the splitting measurements that sample a given model block.

In settings comparable to the case of the idealized subduction zone, three-dimensional variations in anisotropy should be retrievable at spatial scales of 50 km in much of the shallow mantle wedge. With less uniform station spacing (as in the Nicaragua-Costa Rica cases), resolution decreases, but partial to good retrieval of anisotropic structure at horizontal scales of 50-75 km is still possible. Laterally varying anisotropic structures are typically more clearly retrieved by the inversion than are vertically layered structures. Resolution of individual model parameters decreases into the back-arc and the deeper part of the mantle wedge where sampling by local shear-waves is less dense, but this issue can be addressed by directly coupling model perturbations between blocks in the inversion, thus creating effectively larger volumes in which individual model parameters converge on a stable solution and attain a high enough resolution to be meaningfully interpreted.

The SWST method shows promise for imaging broad variations in anisotropy (e.g., arc-parallel vs. arc-normal) at spatial scales that will enable testing of various hypotheses for the processes that govern subduction zone anisotropy (i.e., deformation, melt, fluids). For example, how does the pattern of wedge flow evolve given the competing effects of slab-entrainment, along-arc pressure gradients due to slab shape, and surface plate motions? Does a shift in the dominant olivine slip system and LPO development occur between the cold wedge corner and the hotter outer wedge (e.g., *Jung and Karato* 2001; *Kneller et al.* 2005)? Do melt porosities in the wedge and upper plate reach levels where melt affects the apparent anisotropy (e.g., *Holtzman et al.* 2003)? An ultimate goal is to

combine local-S SWST with the inversion of SKS datasets (e.g., *Chevrot 2006; Long et al. 2008*), integrate the inferred mantle fabric with geochemical, geologic, and other geophysical observations, and build a more comprehensive model of mantle deformation in subduction zones. More generally, SWST also has the potential for resolving anisotropy in any crust and mantle region that is densely sampled by local S-waves.

8. Appendices

8.1. Appendix A: Calculating the “Average” Starting Model

Here we find a three-dimensional model of crystallographic orientation and strength of anisotropy based on a set of shear-wave splitting observations. The simplifying assumption is made that the a -axis of olivine is horizontal and that the b -axis is vertical. The a -axis azimuth (θ) is found by taking an average of the fast directions (ϕ) from all n rays that touch block b , weighted by the path length ($L_{i,b}$) of each ray within that block

$$(A1) \quad \bar{\theta}_b = \frac{\sum_{i=1}^n \phi_i^{obs} \cdot (L_{i,b})}{\sum_{i=1}^n L_{i,b}}.$$

The first step in determining strength of anisotropy (α) is calculating the predicted amount of birefringence per kilometer from the single-crystal elastic coefficients. To do this we assume a vertically propagating shear wave through a crystal oriented such that the a -axis and c -axis are in the horizontal plane. In the case of hexagonal symmetry, the a -axis is still in the horizontal plane. Given this crystal orientation and wave propagation direction and the elastic coefficients used (Table 1), predicted splitting times are 0.0158 s

km^{-1} for orthorhombic symmetry and 0.0197 s km^{-1} for hexagonal symmetry. Therefore, the total predicted splitting time (dt_{total}^{pred}) for a ray is found by multiplying the predicted splitting per kilometer by the total path length, and dividing the observed splitting time (dt^{obs}) by the dt_{total}^{pred} yields a mean strength of anisotropy for that ray. Similar to θ , the average strength in block b is found by averaging the strength of each ray that touches it, weighted by path length within block b

$$(A2) \quad \bar{\alpha}_b = \frac{\sum_{i=1}^n \left(\frac{dt_i^{obs}}{dt_i^{pred}} \right) \cdot (L_{i,b})}{\sum_{i=1}^n L_{i,b}} * 100 .$$

8.2. Appendix B: Generating Larger Model Volumes

In the event that a region of the model space is under-sampled and model parameters in individual blocks are not well-resolved, then correlating groups of blocks will increase resolution of parameters representing the region as a whole. In effect, this is simply a way of coarsening the spatial resolution of the model in particular locations. The choice of which blocks to group is made by assessing both the number of rays that hit a particular block and the resolution matrix described in Section 5.2.

The larger volumes constructed for the Central American model are illustrated in Figure 8. The lateral extent of each volume does not vary with depth, and each of the volumes is 50 km thick (i.e., 2 blocks thick) except for a 25 km-thick “crustal” layer. The volumes in the idealized subduction zone are cubes with 75 km sides (27 smaller blocks) and are located in the back-arc and extend to the base of the model (Fig. 9).

The crystallographic orientation $(\theta_V, \psi_V, \gamma_V)$ and strength of anisotropy (α_V) in each block within the volume is set by a weighted mean of the parameters in all b blocks touched by at least one ray, with the weight given by the diagonal element of the resolution matrix, $R_M(\alpha_i, \alpha_i)$,

$$(B1) \quad \alpha_V = \frac{\sum_{i=1}^b [\alpha_i^{iter-1} \cdot R_M(\alpha_i, \alpha_i)]}{\sum_{i=1}^b R_M(\alpha_i, \alpha_i)}$$

where α_i^{iter-1} is the strength of anisotropy in the i^{th} touched block within volume V from the previous iteration. An identical scheme is used for the angle parameters. Upon generating a uniformly oriented volume of blocks, new constraints must be added to the inversion (equation 4), the form of which are rows in $\mathbf{G}$, $\mathbf{C}_{dd}$, and $d\mathbf{D}$. To illustrate this procedure, we give the example of adding a volume in which only three individual blocks with the volume of interest are touched by rays, and a and q are the only parameters solved for. The new constraints are

$$(B2) \quad \begin{array}{ll} \delta\alpha_1 = \delta\alpha_2 = \delta\alpha_3 & \text{or} \quad \begin{array}{l} \delta\alpha_1 - \delta\alpha_2 = 0 \\ \delta\alpha_1 - \delta\alpha_3 = 0 \\ \delta\theta_1 - \delta\theta_2 = 0 \\ \delta\theta_1 - \delta\theta_3 = 0 \end{array} \\ \delta\theta_1 = \delta\theta_2 = \delta\theta_3 & \end{array}$$

and the $\mathbf{G}$ matrix will gain one row for each equation, with the values being +1 or -1 as shown below in B3

$$(B3) \quad G = \begin{bmatrix} \frac{\delta\phi_n}{\delta\alpha_1} & \frac{\delta\phi_n}{\delta\theta_1} & & \dots & & \dots & \frac{\delta\phi_n}{\delta\alpha_m} & \frac{\delta\phi_n}{\delta\theta_m} \\ \frac{\delta\phi_n}{\delta\alpha_1} & \frac{\delta\phi_n}{\delta\theta_1} & & \dots & & \dots & \frac{\delta\phi_n}{\delta\alpha_m} & \frac{\delta\phi_n}{\delta\theta_m} \\ \frac{\delta dt_n}{\delta\alpha_1} & \frac{\delta dt_n}{\delta\theta_1} & & \dots & & \dots & \frac{\delta dt_n}{\delta\alpha_m} & \frac{\delta dt_n}{\delta\theta_m} \\ \frac{\delta dt_n}{\delta\alpha_1} & \frac{\delta dt_n}{\delta\theta_1} & & \dots & & \dots & \frac{\delta dt_n}{\delta\alpha_m} & \frac{\delta dt_n}{\delta\theta_m} \\ & & \ddots & & & & & \ddots \\ 0 & \dots & 0 & 1 & 0 & -1 & 0 & 0 & 0 & 0 & \dots & 0 \\ 0 & \dots & 0 & 1 & 0 & 0 & 0 & -1 & 0 & 0 & \dots & 0 \\ 0 & \dots & 0 & 0 & 1 & 0 & -1 & 0 & 0 & 0 & \dots & 0 \\ 0 & \dots & 0 & 0 & 1 & 0 & 0 & 0 & -1 & 0 & \dots & 0 \\ & & & & & \vdots & & & & & & \end{bmatrix}$$

and the location (i.e., column) of the (+1,-1) are those corresponding to the global block number. Here we assume that the three blocks within the volume are neither the first nor the m^{th} block in the entire model space. Similarly, the data vector, $\delta\mathbf{D}$, is extended by adding zeros and the data covariance matrix, $\mathbf{C}_{dd}$, is also enlarged by adding prescribed variances, σ_V^2 ,

$$(B4) \quad C_{dd} = \begin{bmatrix} \ddots & & & & & & & & & & & \\ & \sigma_{\phi_n}^2 & & & & & & & & & & \\ & & \sigma_{dt_n}^2 & & & & & & & & & \\ & & & \ddots & & & & & & & & \\ & & & & \sigma_V^2 & 0 & 0 & 0 & & & & \\ & & & & 0 & \sigma_V^2 & 0 & 0 & & & & \\ & & & & 0 & 0 & \sigma_V^2 & 0 & & & & \\ & & & & 0 & 0 & 0 & \sigma_V^2 & & & & \\ & & & & & & & & \ddots & & \end{bmatrix}.$$

We have tested values of 10^{-5} , 10^{-3} , and 10^{-1} for σ_V^2 (Table 2) and find that 10^{-1} results in the best data and model misfit reduction. Smaller values of variance (10^{-5} and

10^{-3}) give the large volume constraints too much influence, while the larger value of 10^{-1} allows the actual data (splitting parameters) to contribute at a non-trivial level.

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10. List of Figures and Tables

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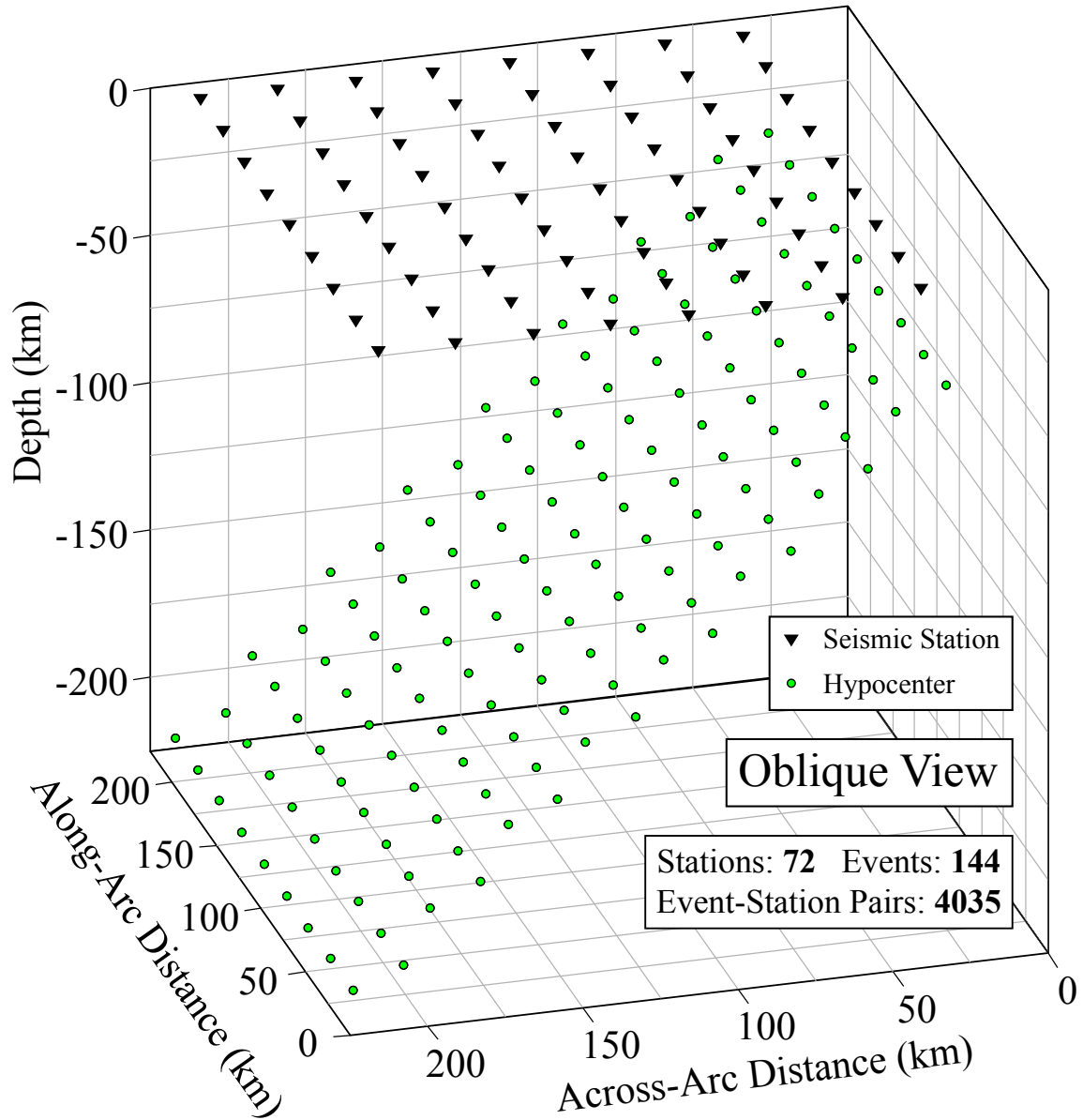


Figure 1. Event and station locations in an idealized subduction zone model. Starting in the second layer, each block along a 45° angle contains either 1, 2, 3, or 4 events per block; 60° slab dips are also employed. The example shown here has 2 events per block, but the inversion results presented in Section 5.5 and 5.6 all contain 4 events per block. We vary the number of events per block to assess the effects of increasing or decreasing the number and distribution of crossing rays. Only ray paths that have incidence angles within the shear-wave window ($<35^\circ$) are actually used, and therefore, splitting measurements from a fraction of all ray paths are included in the inversion. For reference in the 45° case, one event per block yields 2022 event-station pairs within the shear-wave window, 4035 for 2 events per block, 6035 for 3, and 8070 for 4. Initial polarizations are randomly generated and are the same for all rays coming from a single event. The cubic grid illustrates the extent of the model space.

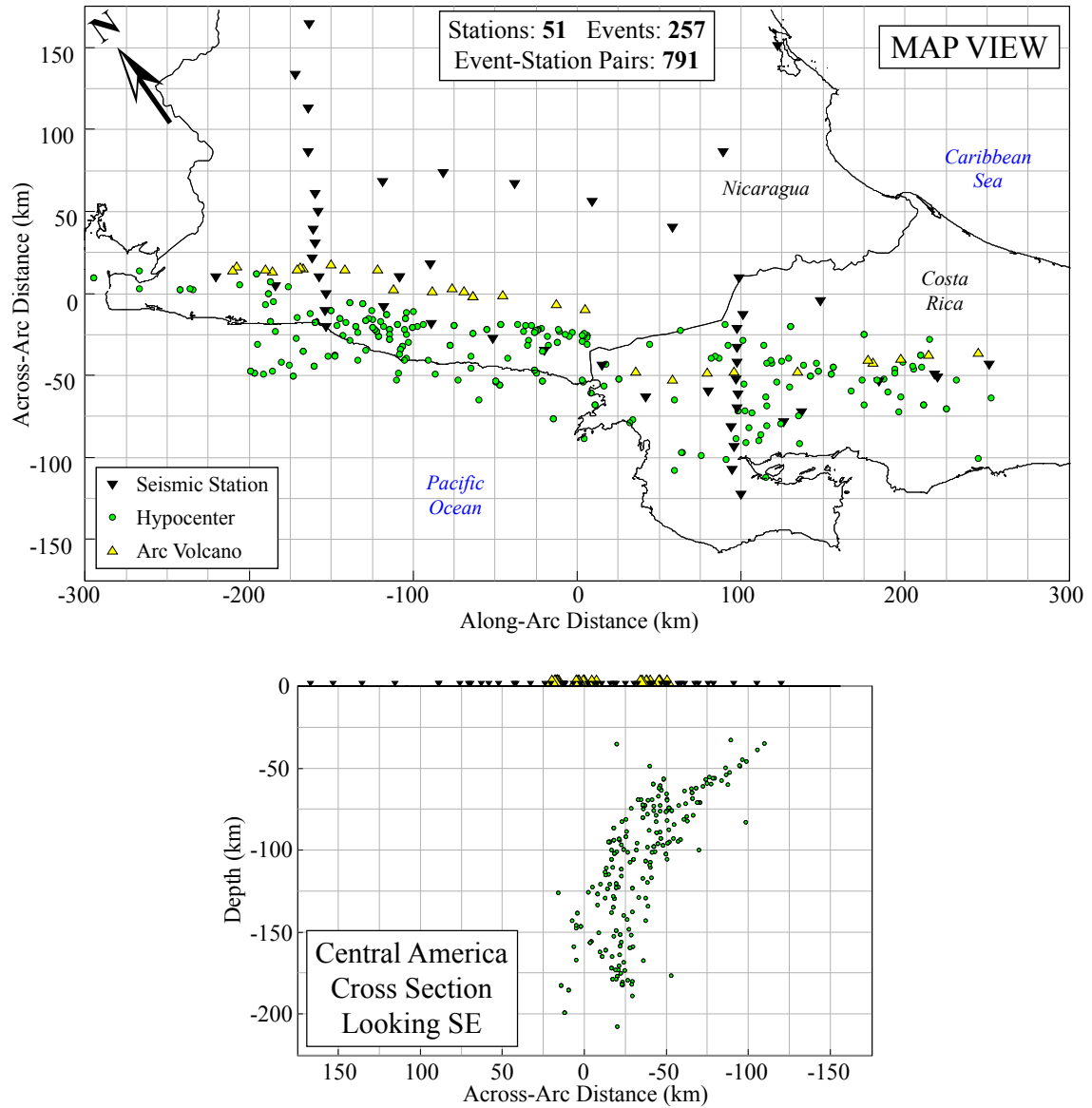


Figure 2. Events here correspond to those from actual high quality shear-wave splitting measurements in Central America (*Abt et al. 2006*) and stations are from the recent TUCAN seismic experiment. The coordinate axes of the model space are rotated 35° west of north in order to characterize distances in approximately along-arc and across-arc directions. Note the distinct jogs in the volcanic arc and the non-uniform distribution of events relative to the idealized subduction zone. Slab seismicity appears very thick in the cross-section (bottom) because its strike changes along the arc and is not always aligned with the approximate along-arc/across-arc coordinate axes used here.

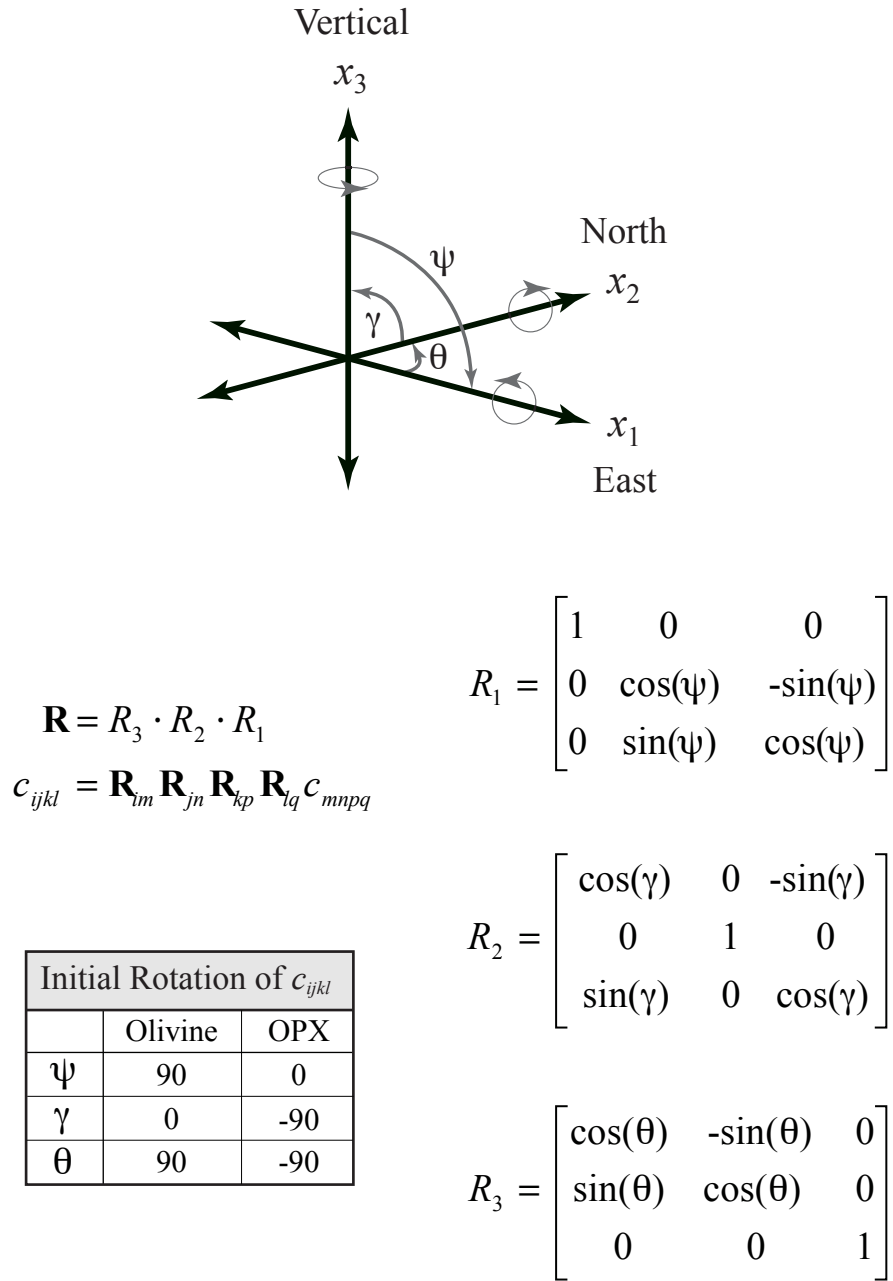


Figure 3. Model parameters and rotation scheme used in the forward modeling for orienting the crystallographic axes of olivine and orthopyroxene. ψ , γ , and θ rotate around the x_1 , x_2 , and x_3 axes, respectively. The angles used to orient olivine and opx as described in the text (e.g., *Mainprice and Silver 1993*) are given in the table; in this case, the a -axis of olivine is parallel to x_2 and the b -axis is vertical.

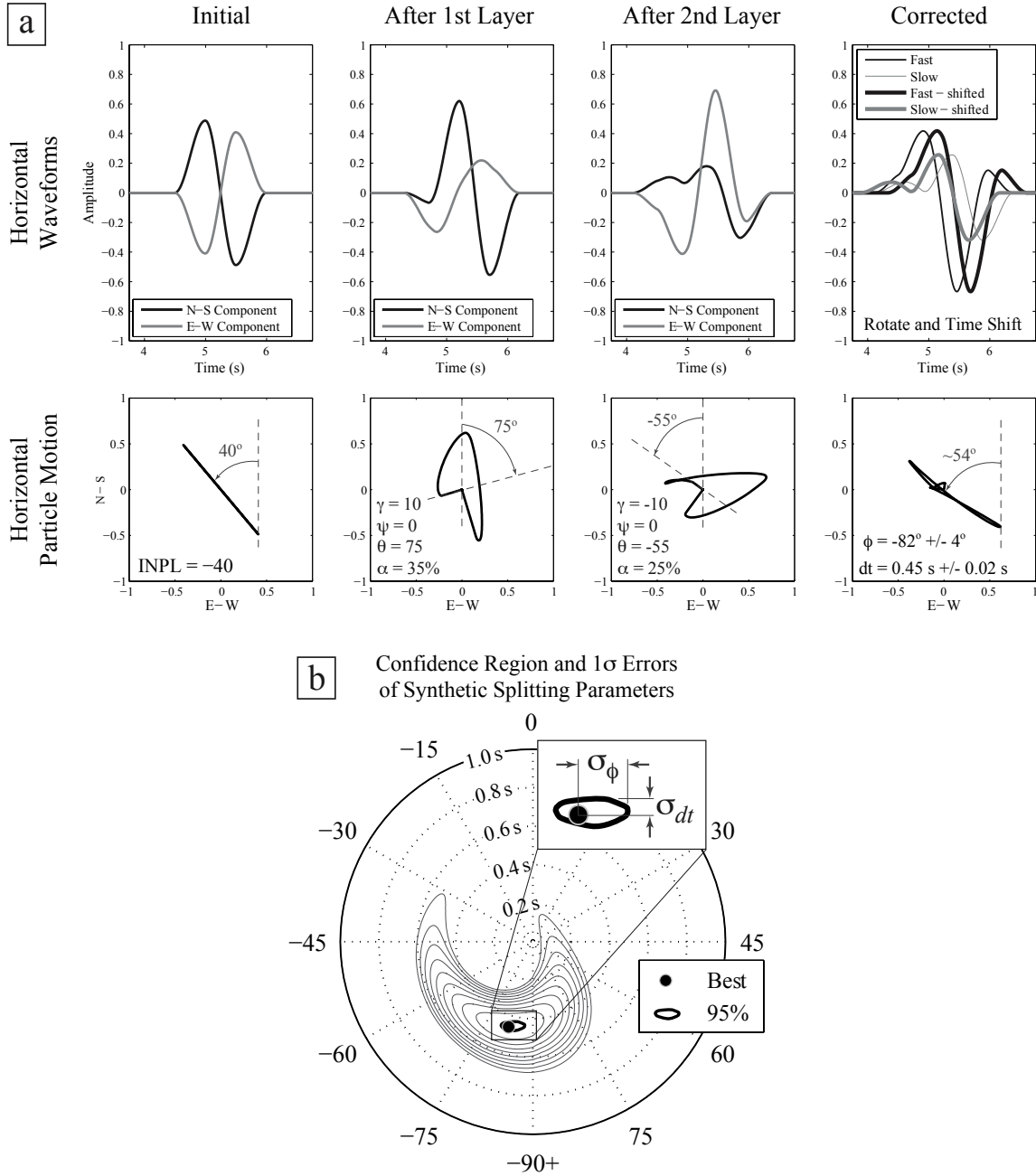


Figure 4. Example of synthetic shear-wave splitting used in the forward modeling (Section 4.1). **(a)** From left to right, an initial wavelet with linear horizontal particle motion (-40°) passes through the first anisotropic layer (orientation given by the three angles in the second lower box) and is split. The characteristic elliptical particle motion of a split wave is generated. In the third column, the result of passing the already split wave through a second anisotropic layer of different crystallographic orientation and strength of anisotropy can be seen. Owing to the nearly horizontal orientation of the a -axis, the fast and slow direction in both layers can easily be distinguished in the particle motion plots. The steps of rotating and time-shifting the waveforms are illustrated in the top right graph. The thin lines represent the result of rotating the E-W (grey) and N-S (black) components by ϕ (-82°). The thick line shows the fast waveform shifted by $+dt/2$ and the thin is the slow waveform shifted by $-dt/2$. Note that the corrected linear particle motion does not exactly match the input particle motion. **(b)** Confidence contours around the best-fitting pair of splitting parameters and 1σ error bars for this example. North is up (i.e., 0), and $\pm 90^\circ = \text{E-W}$.

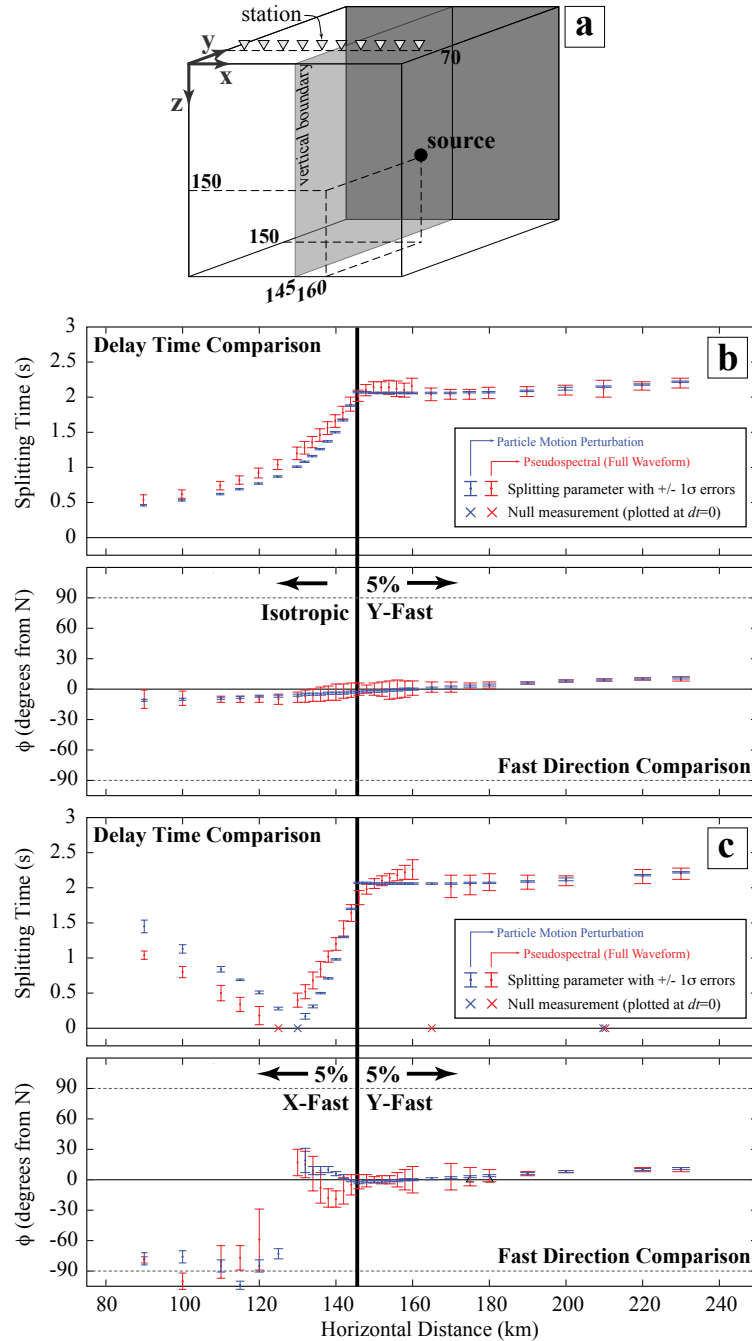


Figure 5. Comparison of splitting parameters calculated from the particle motion perturbation method (blue) and full-waveform (red) pseudospectral synthetics (*Hung and Forsyth 1998*). (a) Each example has a double-couple source (strike=0°, dip=45°, rake=45°) located at a depth of 150 km ($x=160$ km, $y=150$ km) and a vertical boundary (thick black line in (a) and (b)) at a horizontal distance of 145 km ($x=145$ km) between the two sides of the model. Splitting is calculated along a line of stations at $y=70$ km. (b) Model is isotropic on the left side of the boundary and 5% fast in the y -direction on the right. The splitting parameters show good agreement (≤ 0.11 s for dt , fast directions all within error of each other when measurement uncertainties are included). (c) Model with 5% fast velocity in the x -direction on the left side of the boundary and 5% fast in the y -direction on the right side. This large, sharp velocity contrast produces some ray bending and wavefront interaction with the boundary, which leads to moderate differences ($\leq 16^\circ$, ≤ 0.26 s when measurement uncertainties are included) in splitting parameters between the two methods.

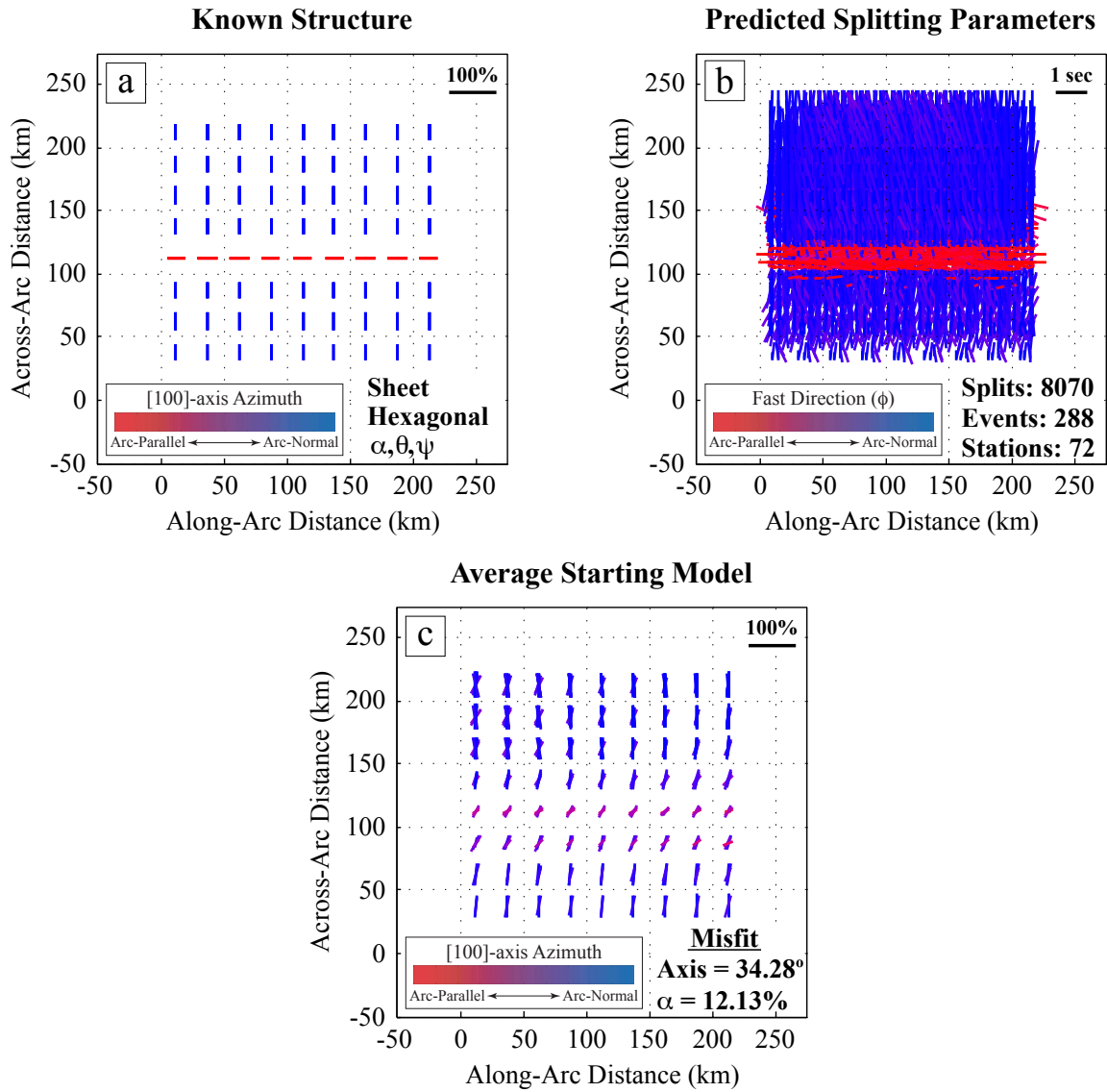


Figure 6. Illustration of the average starting model described in Section 4.1 and Appendix A. (a) The known structure in the example shown here is a vertical sheet of horizontal arc-parallel a -axes within the idealized subduction zone (45° slab dip angle) surrounded by arc-normal a -axes dipping 30° away from the trench, which is at $y=0$ km. The symmetry assumed here is hexagonal, and the model parameters used are α , θ , and ψ . Vectors represent the a -axis of olivine and are plotted at model block centers with length corresponding to strength and color indicating azimuth. (b) Predicted splitting parameters from the known structure, plotted as standard shear-wave splitting vectors at ray path midpoints, with length corresponding to delay time and color (and orientation) indicating fast direction. (c) a -axis orientations in the starting model after averaging the fast directions from (b) and determining the average strength from the splitting times. The mean strength misfit with the known structure is quite small (12.13%), while the mean angular misfit of the a -axes is larger (34.28°) because the a -axes in the starting model are horizontal.

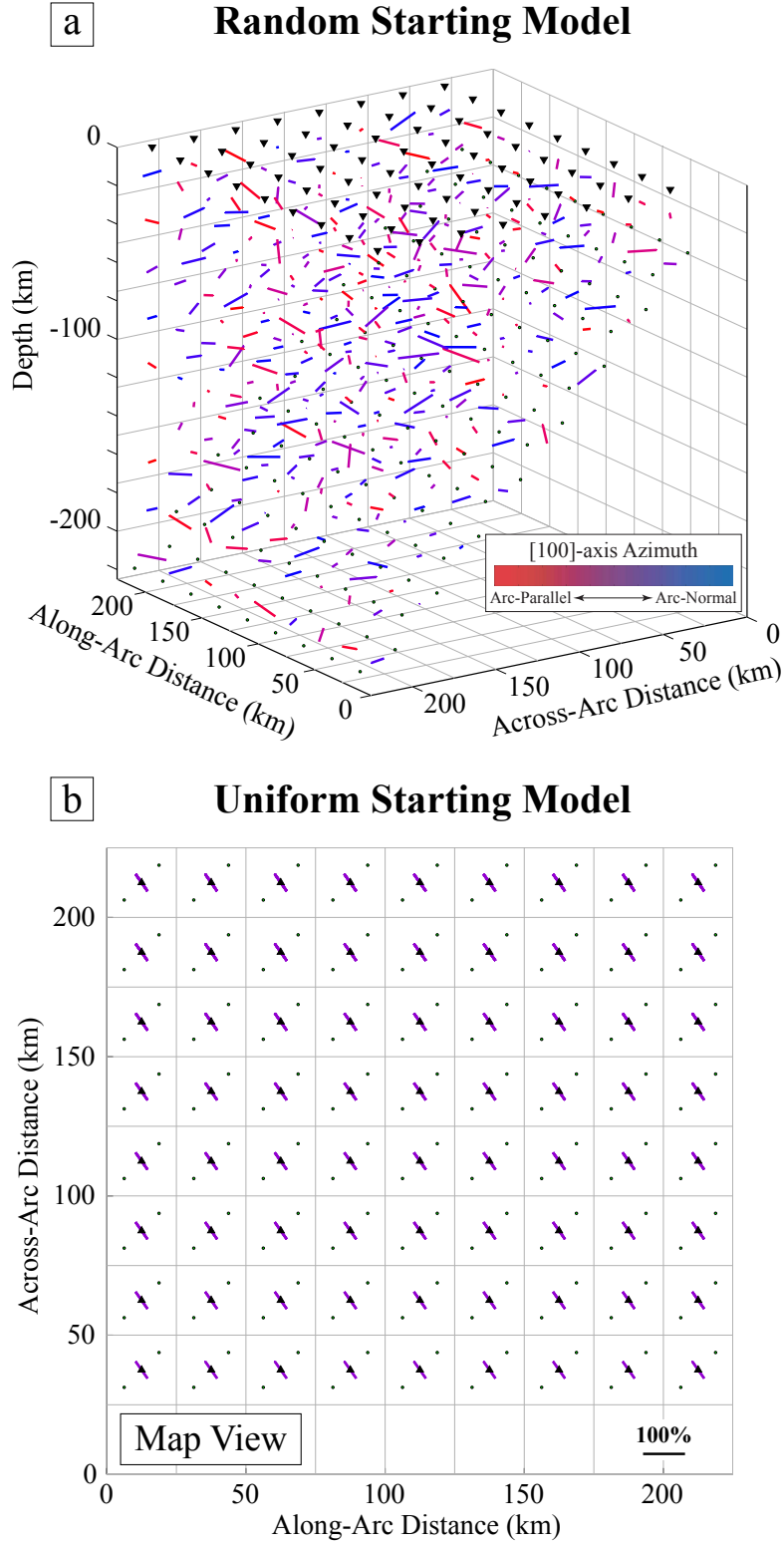


Figure 7. (a) Random and (b) uniform starting models in the idealized subduction zone. The view in (a) is down and oblique to the dipping surface of events. Unlike in the average (Fig. 6) and uniform (b) starting models, the a -axes in the random starting model are not restricted to the horizontal plane. The small circles are the synthetic hypocenters and the black triangles are the synthetic stations.

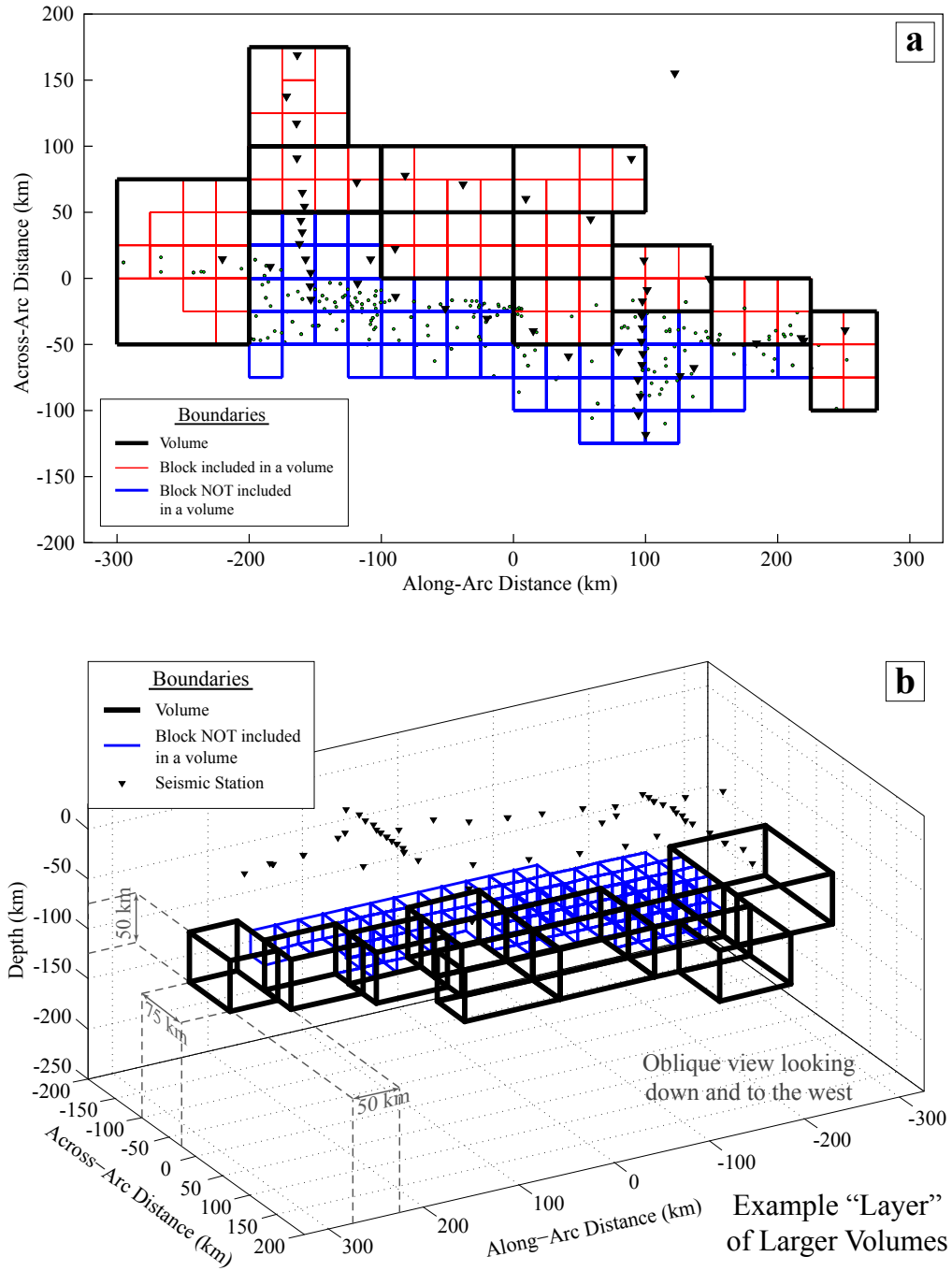


Figure 8. Larger volumes generated to increase resolution of individual parameters in the Central American subduction zone. Blue blocks are those that remain independent after volumes (thick black lines) are formed. (a) Volumes used in the Central American subduction zone are not uniform and are based on hit counts. The red blocks are those within the volume that are actually touched by a ray and used in the inversion. (b) Example “layer” of larger volumes and independent blocks. The lateral extent of each larger volume does not change with depth, and each layer is 50 km thick, except for the surface layer, which is only 25 km. For clarity, the red blocks within the volumes are not shown here.

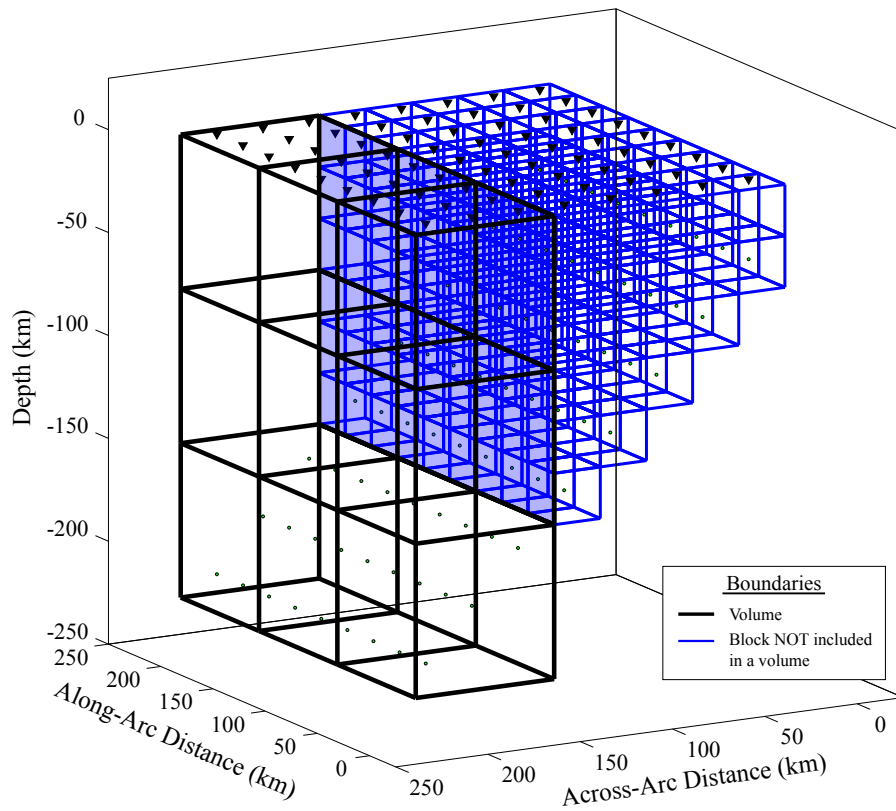


Figure 9. Three-dimensional view of the larger volumes used in the idealized subduction zone. 75^3 km^3 volumes are made out of blocks furthest from the trench where the fewest crossing rays exist. Nearly all blocks within the volumes are touched, but are not shown here for clarity.

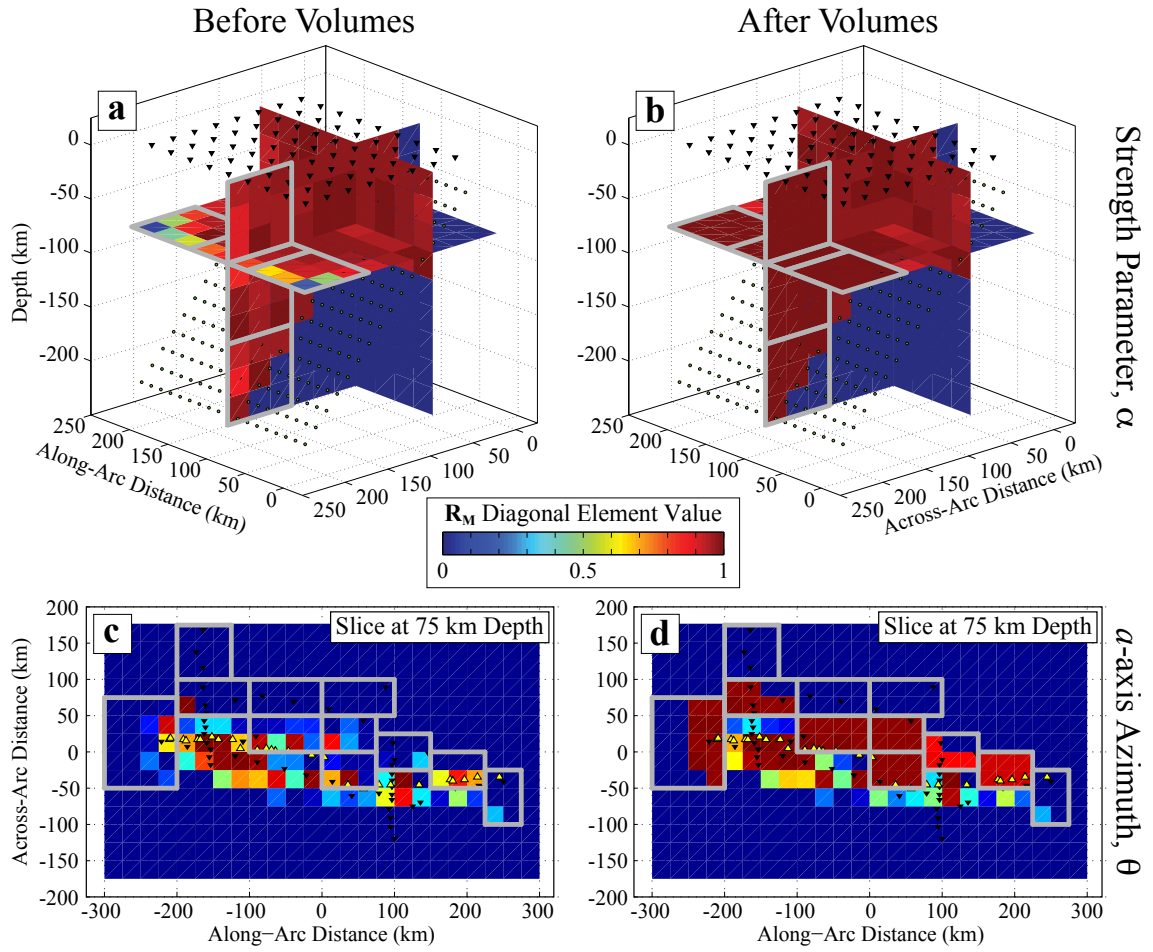


Figure 10. Effects of large volumes on model parameter resolution. Values shown are those of the diagonal element of the resolution matrix corresponding to the strength parameter, α (a-b), and the a -axis azimuth, θ (c-d) at iteration 25 (a) and 30 (c) just prior to adding the large volume constraints and at iteration 26 (b) and 31 (d) just after applying the constraints. The outline of the volumes is shown with thick, light grey lines. Note that in (c) and (d), some volumes do not have blocks at 75 km that are touched by a ray, and therefore, the region displays a resolution value of zero.

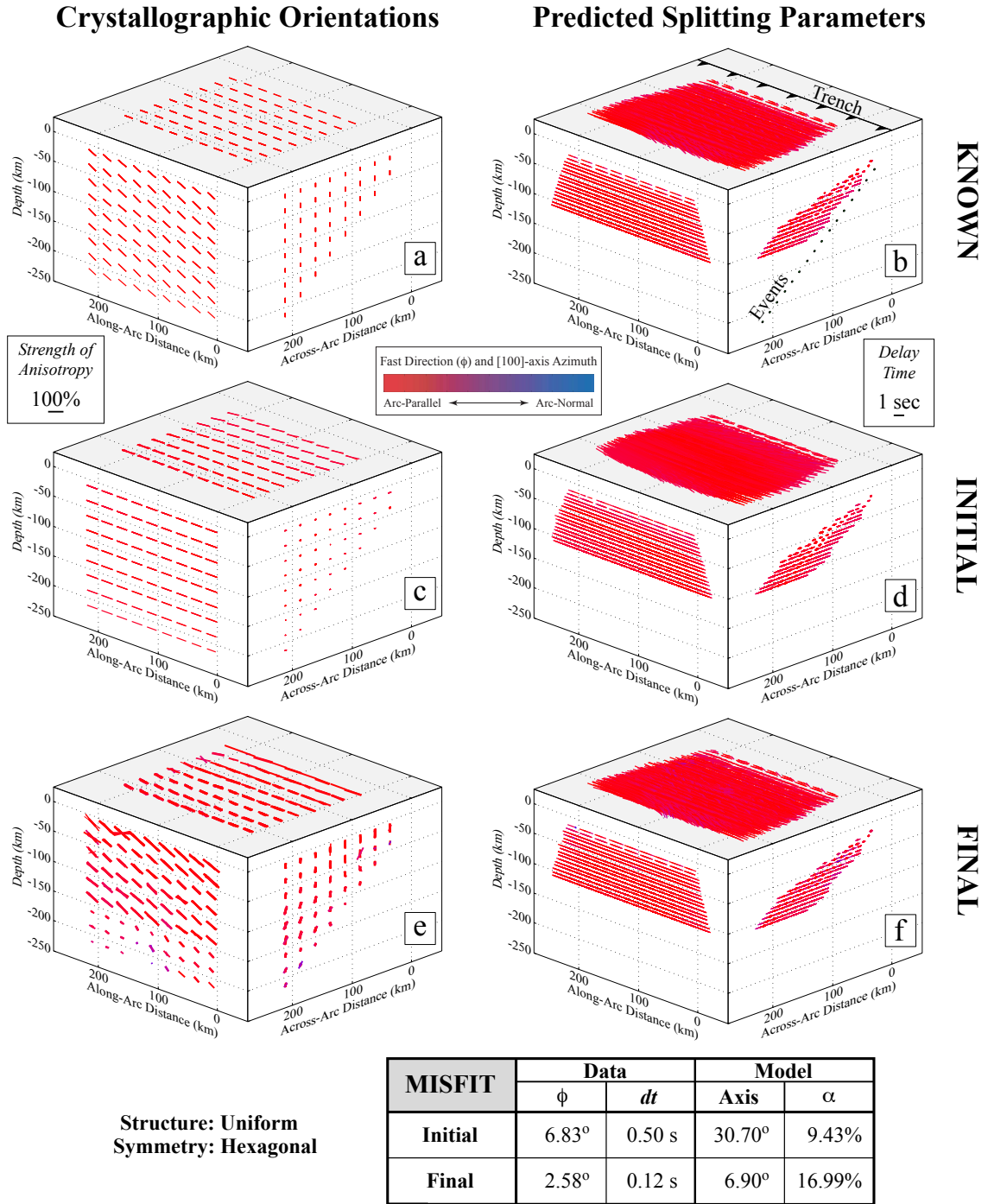


Figure 11. Inversion results from the uniform structure for the idealized event-station distribution with a slab dip of 45°. Vectors are projected onto a surface, along-arc, and across-arc plane to provide a quasi 3-D view of the model and data. The models of anisotropy (i.e., olivine a -axes) are on the left and the predicted splitting parameters (i.e., splitting vectors plotted at ray path midpoints) from each model are on the right. The middle row shows the average starting model and splitting parameters; note that all blocks touched by at least one ray are displayed in the known and starting models. The bottom row shows the well-resolved blocks in the final model along with the predicted splitting parameters. Thicker vectors in the final model represent higher resolution of model parameters in that block, and vectors are shown only when all model parameters in a given volume achieve a resolution matrix diagonal value greater than 0.25. The table indicates initial and final data and model misfits.

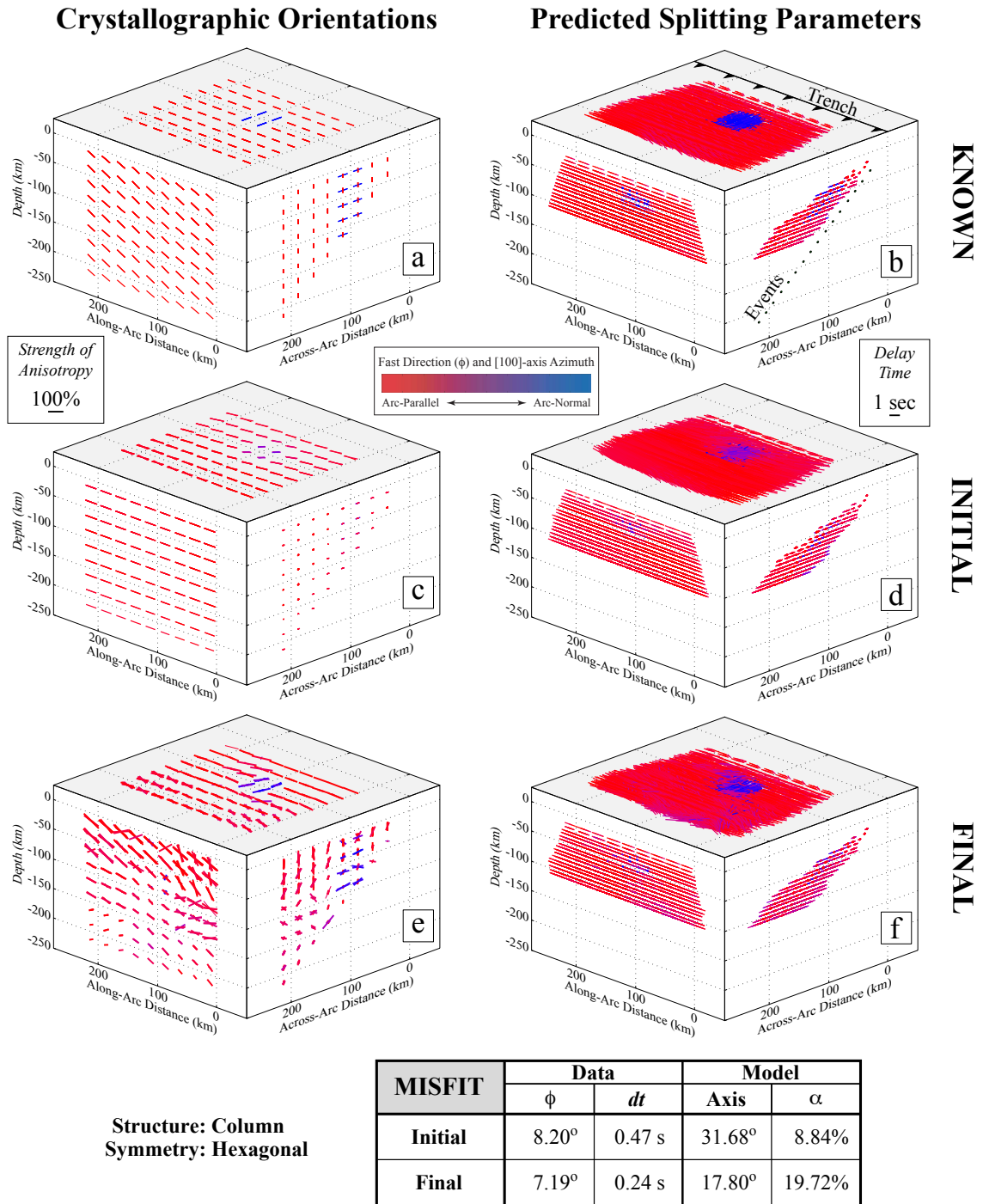


Figure 12. Inversion results from the column structure for the idealized event-station distribution with a slab dip of 45°. See caption from Figure 11 for full description.

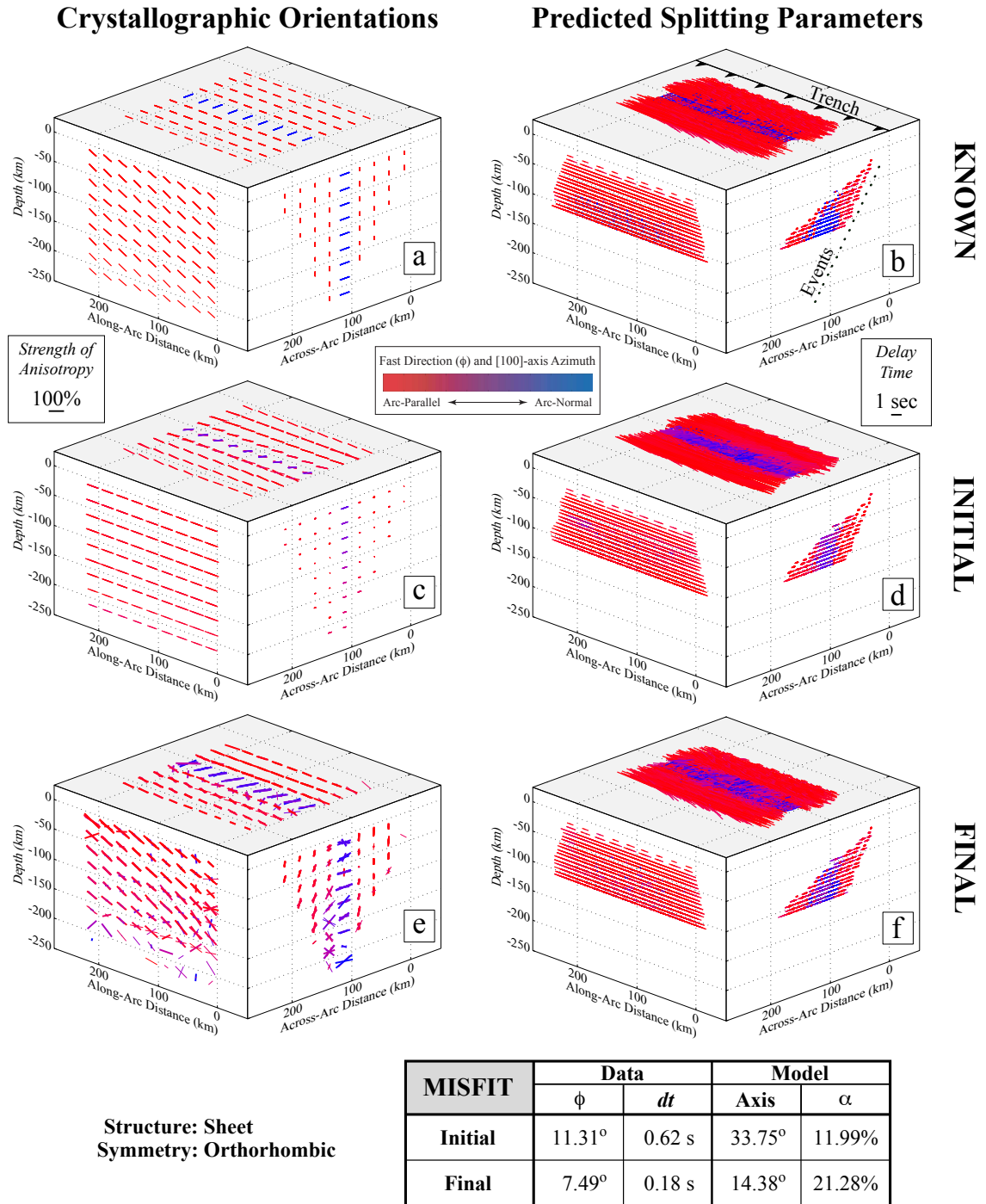


Figure 13. Inversion results from the sheet structure for the idealized event-station distribution with a slab dip of 60°. See caption from Figure 11 for full description.

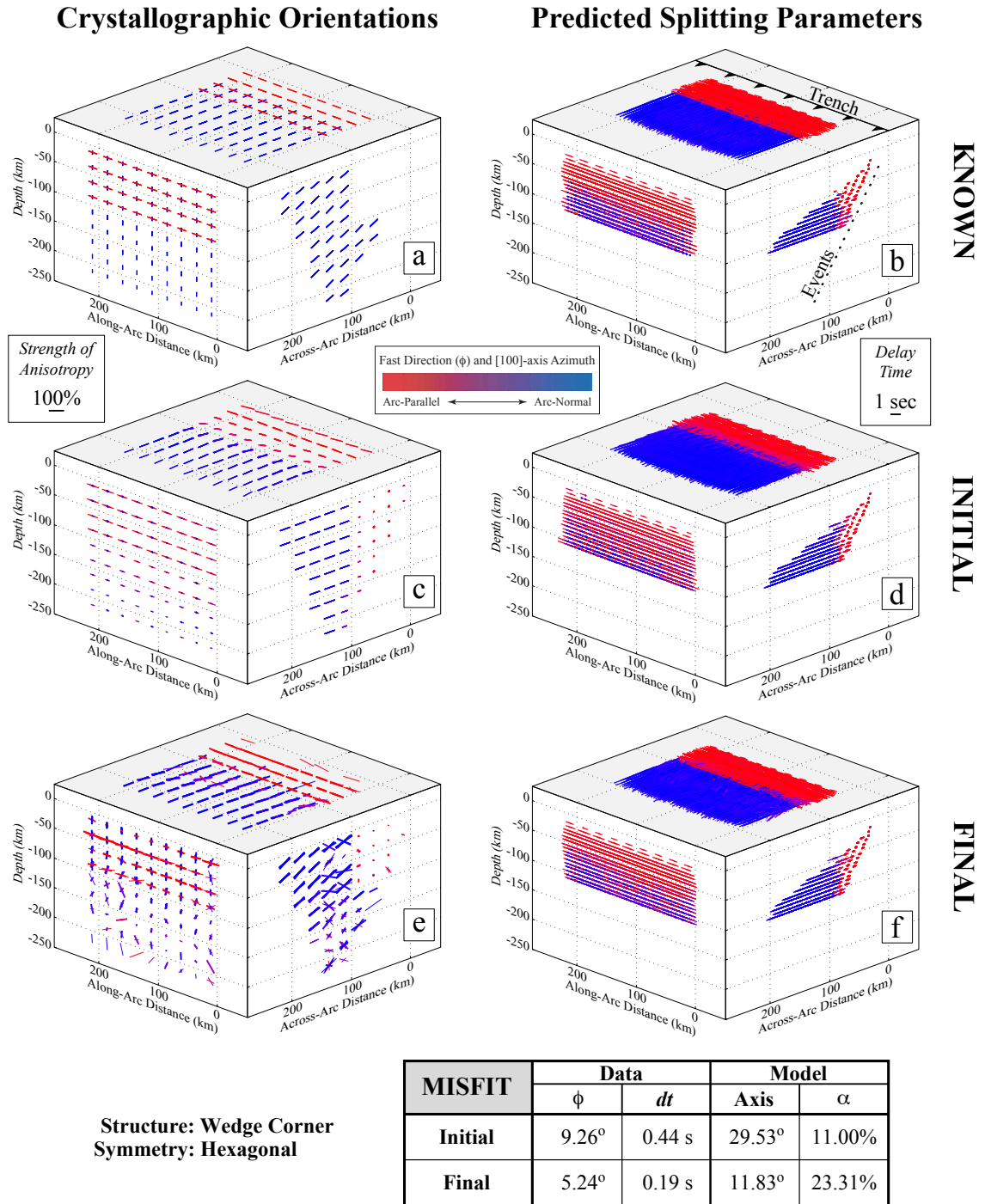


Figure 14. Inversion results from the wedge corner structure for the idealized event-station distribution with a slab dip of 60°. See caption from Figure 11 for full description.

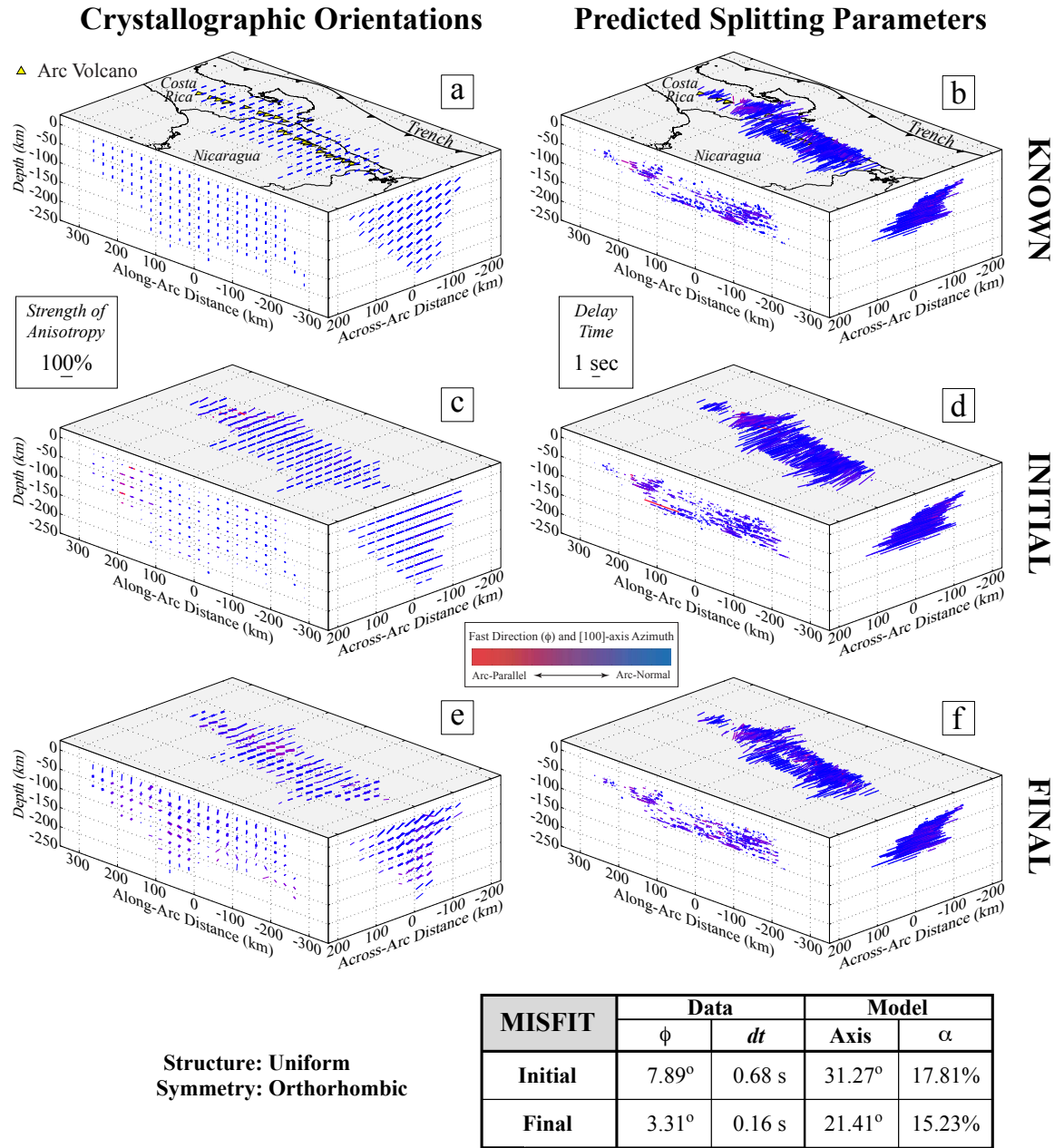


Figure 15. Inversion results from the uniform structure with the Central American event-station distribution. See caption from Figure 11 for full description.

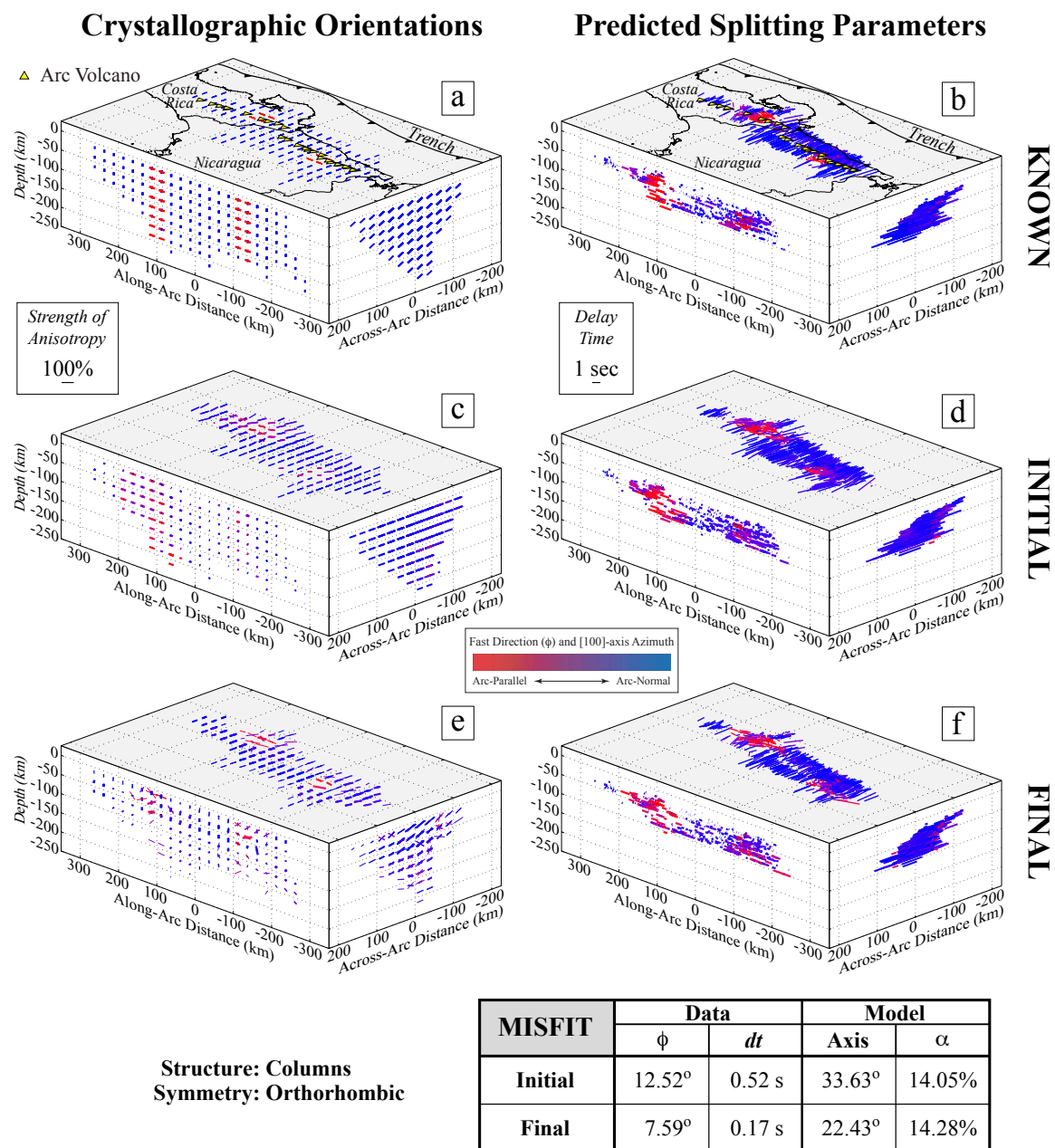


Figure 16. Inversion results from the columns structure with the Central American event-station distribution. See caption from Figure 11 for full description.

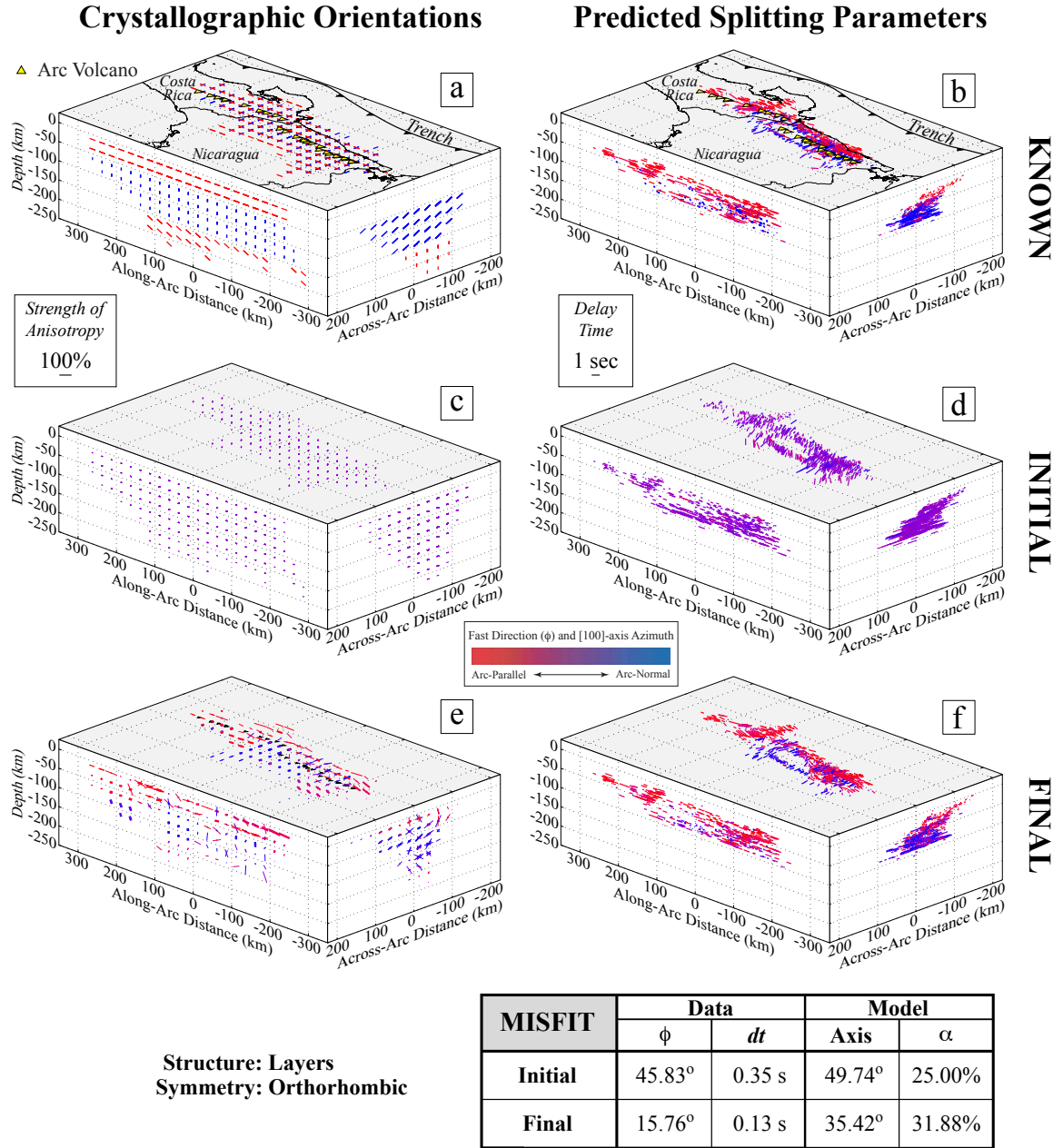


Figure 17. Inversion results from the layers structure with the Central American event-station distribution. See caption from Figure 11 for full description.

ij	Olivine*, Fo ₉₀ (Mg _{1.8} Fe _{0.2} SiO ₄)			OPX**, Bronzite (Mg _{0.8} Fe _{0.2} SiO ₃)		
	C_{ij} (at 1 bar)	$\frac{\delta C_{ij}}{\delta P}$	$\frac{\delta C_{ij}}{\delta T}$	C_{ij} (at 1 bar)	$\frac{\delta C_{ij}}{\delta P}$	$\frac{\delta C_{ij}}{\delta T}$
	units GPa	-	GPa · K ⁻¹	GPa	-	GPa · K ⁻¹
11	323.7	7.98	-0.034	228.6	11.04	-0.0352
12	66.4	4.74	-0.0105	79.9	6.97	-0.0212
13	71.6	4.48	-0.0094	63.2	9.09	-0.0318
22	197.6	6.37	-0.0285	160.5	9.19	-0.0328
23	75.6	3.76	-0.0051	56.8	8.73	-0.0107
33	235.1	6.38	-0.0286	210.4	16.42	-0.0516
44	64.62	2.17	-0.0128	81.75	2.38	-0.0131
55	78.05	1.64	-0.013	75.48	2.92	-0.0138
66	79.04	2.31	-0.0157	77.6	2.75	-0.0145

*Anderson and Isaak (1995) and Abramson et al. (1997); **Frisillo and Barsch (1972)

$$\begin{array}{lll}
 C_{11} = c_{1111} & C_{12} = c_{1122} = c_{2211} & C_{44} = c_{2323} = c_{2332} = c_{3232} = c_{3223} \\
 C_{22} = c_{2222} & C_{23} = c_{2233} = c_{3322} & C_{55} = c_{1313} = c_{1331} = c_{3131} = c_{3113} \\
 C_{33} = c_{3333} & C_{13} = c_{1133} = c_{3311} & C_{66} = c_{1212} = c_{1221} = c_{2121} = c_{2112}
 \end{array}$$

Table 1. Single-crystal anisotropic elastic coefficients (C_{ij}) at 1 bar, pressure derivatives ($\delta C_{ij} / \delta P$), and temperature derivatives ($\delta C_{ij} / \delta T$) used to represent the upper mantle in the forward calculations of shear-wave splitting. The orthorhombic symmetry of olivine and opx requires 9 elastic coefficients, which are typically given as elements in the 6-by-6 matrix C_{ij} . By symmetry $C_{12}=C_{21}$, $C_{13}=C_{31}$, and $C_{23}=C_{32}$. The C_{ij} are part of the fourth-order stiffness tensor, c_{klmn} , that fully describes the response of an elastic material to stress, and here, C_{11} corresponds to the [100] axis, C_{22} is the [010] axis, and C_{33} is the [001] axis. In our analysis, we assume a starting orientation of the olivine-opx aggregate such that the [100]-axis of olivine is initially parallel to the x_2 direction, and the C_{ij} presented here must be rotated as described in Figure 3. Olivine C_{ij} are averaged from Anderson and Isaak (1995) and Abramson et al. (1997), $\delta C_{ij} / \delta P$ are from Abramson et al. (1997) and $\delta C_{ij} / \delta T$ are from Anderson and Isaak (1995). The C_{ij} , $\delta C_{ij} / \delta P$, and $\delta C_{ij} / \delta T$ of opx are from Frisillo and Barsch (1972). For completeness, the mapping between C_{ij} and c_{klmn} is provided here (e.g., Babuska and Cara 1991).

Inversion Parameter	Testing Conditions	Values Tested	Preferred Value
Initial Model Parameter Variance $\sigma_{m_0}^2$	All 3 starting models, uniform known structure, hexagonal symmetry, solving for α and θ , 25^3 km ³ blocks, both subduction zones, 2 events/block with a 45° slab in the idealized, 15 iterations	1 5 10	1
Damping Relaxation	All 3 starting models, columns known structure, hexagonal (α, θ, ψ) and orthorhombic ($\alpha, \theta, \psi, \gamma$) symmetry, 25^3 km ³ blocks, Central America subduction zone, 20 and 40 iterations	1 to 5 1 to 10 5 to 20	1 to 10
Variance for Larger Volumes σ_V^2	All 3 starting models, columns known structure, hexagonal symmetry, solving for α and θ , both the idealized and Central America subduction zones, 25^3 km ³ blocks, 20 iterations	10^{-5} 10^{-3} 10^{-1}	10^{-1}
Correlation Length (km) Δ_0	Average starting model, column(s) known structure, damping=5, hexagonal symmetry, solving for α and θ , 25^3 km ³ blocks, 1 and 2 events/block in the idealized subduction zone, 45° slab, 15 iterations	0 12.5 15 20 25	0 (no smoothing)

Table 2. Parameters tested in order to optimize the behavior and results of the inversion. The basis for each variable is described in Section 4.3, and their effects on data and model misfit are discussed in Section 5.4. A brief description of the different conditions under which each of the variables was tested is also given here. As illustrated, we have tested multiple scenarios to understand the impact of these inversion variables.

Chapter 2

Shear-Wave Anisotropy Beneath Nicaragua and Costa Rica: Implications for Flow in the Mantle Wedge

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Abstract

We present new shear-wave splitting data from local events in Costa Rica and Nicaragua recorded by the temporary (July 2004-March 2006) 48 station TUCAN broadband seismic array. Observed fast polarization directions in the fore-arc, arc, and back-arc range from arc-parallel to arc-normal over very short distances (<5 km when plotted at ray path midpoints) making the direct interpretation of individual splitting measurements in terms of flow tenuous, even when considering variations in the relationship between lattice-preferred orientation and deformation (e.g., B-type dislocation creep in olivine). Therefore, we tomographically invert the splitting measurements to find a three-dimensional model of crystallographic orientation in the wedge. We assume the elastic constants of olivine and orthopyroxene with hexagonal symmetry and use a damped, iterative least-squares approach to account for the non-linear behavior of splitting when considering three-dimensional ray propagation and distributions of anisotropy. The best-fitting model contains roughly horizontal, arc-parallel olivine a-axes in the mantle wedge down to at least 125 km beneath the back-arc and arc, which we interpret to indicate along-arc flow in the mantle wedge. Pb and Nd isotopic ratios in arc lavas provide additional evidence for arc-parallel flow and also constrain the direction (northwest, from Costa Rica to Nicaragua) and minimum flow rate (63-190 mm/yr). With only slightly oblique subduction at 85 mm/yr of the relatively planar Cocos Plate, the most likely mechanism for driving along-arc transport is toroidal flow around the edge of the slab in southern Costa Rica, generated by greater slab roll-back in Nicaragua. Two important implications of this arc-parallel flow are the progressive depletion of the mantle source for arc lavas from Costa Rica to Nicaragua and the possible need for significant decoupling between the wedge and down-going plate.

1. Tectonic and Geochemical Variations Across Nicaragua and Costa Rica

The pattern of solid mantle flow in subduction zones fundamentally influences wedge thermal structure, melt generation and transport, and slab dynamics. Seismic anisotropy offers a means of imaging mantle wedge flow, and a growing database of shear-wave splitting measurements from subduction zones [e.g., *Wiens and Smith*, 2003; *Pozgay et al.*, 2007; *Long and Silver*, 2008] has revealed flow patterns that are considerably more complex than those predicted by 2-D entrainment of the wedge with the down-going plate (2-D corner flow) [e.g., *McKenzie*, 1979; *Ribe*, 1989; *Fischer et al.*, 2000; *Blackman and Kendall*, 2002; *Lassak et al.*, 2006]. This conclusion holds true when the [100]-axis of olivine is assumed to align parallel to flow, as suggested by experimentally and naturally deformed dunites and harzburgites at relatively dry conditions [e.g., *Zhang and Karato*, 1995; *Ismail and Mainprice*, 1998], or when taking into account recent deformation experiments [*Bystricky et al.*, 2000; *Jung and Karato*, 2001; *Holtzman et al.*, 2003; *Mainprice et al.*, 2005; *Jung et al.*, 2006; *Katayama and Karato*, 2006] and field observations [*Mehl et al.*, 2003; *Mizukami et al.*, 2004] that have shown olivine fabrics depend on pressure, temperature, melt and water content, and stress.

For example, in the Japan and Ryukyu subduction zones, a shift from arc-parallel (perpendicular to plate motion) fast directions in the fore-arc to arc-normal (plate motion parallel) fast directions in the back-arc is observed in both local-S [*Long and van der Hilst*, 2006; *Nakajima et al.*, 2006] and teleseismic-SKS [*Long and van der Hilst*, 2005] splitting measurements. This pattern is consistent with 2-D corner flow, if a B-type fabric [*Jung and Karato*, 2001; *Jung et al.*, 2006] exists in the cold wedge corner [*Kneller*

et al., 2005]. However, in other subduction zones, fast directions are arc-parallel or vary from arc-parallel to arc-normal for paths that sample the wedge well outside its cold corner, indicating significantly 3-D flow [*Wiens and Smith*, 2003; *Long and Silver*, 2008]. Examples include South America [*Russo and Silver*, 1994; *Polet et al.*, 2000; *Anderson et al.*, 2004; *Anderson et al.*, in review], New Zealand [*Matcham et al.*, 2000; *Audoine et al.*, 2004; *Morley et al.*, 2006], Tonga-Fiji [*Smith et al.*, 2001], Taiwan [*Huang et al.*, 2006], the Marianas [*Pozgay et al.*, 2007], Izu-Bonin [*Anglin and Fouch*, 2005], Kamchatka [*Peyton et al.*, 2001; *Levin et al.*, 2004], the Aleutians [*Yang et al.*, 1995], Alaska [*Christensen et al.*, 2003], and Central America [*Hoernle et al.*, 2008; this paper].

Physical factors that could contribute to 3-D flow include oblique convergence [*Hall et al.*, 2000], trench retreat [*Buttles and Olsen*, 1998; *Kincaid and Griffiths*, 2004; *Funiciello et al.*, 2006], slab edges [*Kincaid and Griffiths*, 2003; *Funiciello et al.*, 2006] or tears, upper plate deformation or delamination [*Hall et al.*, 2000; *Blackman and Kendall*, 2002; *Kincaid and Hall*, 2003, *Lowman et al.*, 2007; *Behn et al.*, 2007], non-planar slab geometry [*Kneller and van Keken*, 2007], and a low viscosity wedge channel [*Conder and Wiens*, 2007]. In addition, shape-preferred orientation (SPO), the alignment of heterogeneous features such as melt or fluid-filled cracks within the mantle and lithosphere/crust, should be accounted for when interpreting measurements of anisotropy in complex regions such as subduction zones.

In this study we present shear-wave splitting measurements from waveforms collected by the Tomography Under Costa Rica and Nicaragua (TUCAN) Broadband Seismic Experiment (Fig. 1) and a 3-D model for anisotropy in the Nicaragua-Costa Rica subduction zone obtained by inverting the splitting measurements using a new

tomographic approach [Abt and Fischer, 2008]. Our objective was to place constraints on the pattern of wedge flow, and we find evidence for significant anisotropy with a fast symmetry axis parallel to the arc in the mantle beneath the arc and back-arc. These results, when combined with Pb and Nd isotope data (discussed in Section 2), strongly suggest along-arc flow to the northwest in the mantle wedge beneath Nicaragua and Costa Rica [Hoernle *et al.*, 2008]. Although the 3-D anisotropic model has been previously published [Hoernle *et al.*, 2008], in this study we explore its uniqueness and implications in greater depth.

2. Tectonic and Geochemical Variations Across Nicaragua and Costa Rica

The Nicaragua-Costa Rica subduction zone offers a unique environment to investigate geodynamic, seismogenic, and petrophysical processes in the mantle wedge because distinct changes in arc lava geochemistry, seismic activity, and geomorphology are observed over a relatively short distance (~500 km). Here we review features of the tectonic setting that are particularly relevant to understanding regional mantle flow.

The Cocos Plate, which is subducting beneath the southwestern edge of the Caribbean and North American Plates (Fig. 1), formed at both the East Pacific Rise (EPR) and the Galapagos Spreading Center (GSC). Seafloor magnetic anomalies place the junction between these two components of the Cocos plate at the southern end of the Nicoya Peninsula. Convergence occurs at a rate of ~85 mm/yr, with obliquity varying from ~7° in central Costa Rica to ~17° in Nicaragua [Barckhausen *et al.*, 2001; DeMets, 2001].

The shape of the subducting slab may be an important driver for subduction zone flow, and along-strike changes in Cocos slab shape have been suggested [Protti *et al.*, 1994; Husen *et al.*, 2003; Syracuse and Abers, 2006]. However, velocity tomography and hypocentral relocation using data from the TUCAN array [Syracuse *et al.*, 2008] indicates that the slab below 50 km depth dips at 50-60° beneath most of Nicaragua and Costa Rica, with a moderate shallowing of slab dip to roughly 45° beneath the southeastern end of the array in central Costa Rica. No resolvable change in slab geometry occurs where the arc shifts 40 km further from the trench at the Nicaragua-Costa Rica border. In southeastern Costa Rica, significant thrust faulting occurs in the absence of large-volume arc volcanism [Carr *et al.*, 2003] and deep Cocos Plate-related seismicity [Protti *et al.*, 1994]. Abratis and Wörner [2001] contend that the lack of deep seismicity, a unique progression of arc geochemistry, and shallow subduction represent the opening of a slab window caused by the subduction of the Cocos Ridge, allowing Galapagos hot spot-enriched mantle to pass into the Caribbean side of the subduction zone. Another reasonable explanation for the termination of volcanism and deep seismicity would be that this is the location of the slab edge, with the absence of a lithospheric barrier between Cocos-Nazca (i.e., Pacific) upper mantle and Caribbean (i.e., Atlantic) upper mantle [e.g., Russo and Silver, 1994]. In addition, trench motion (e.g., slab roll-back during trench retreat) near the slab edge could produce a prominent deviation from a simple 2-D corner flow pattern [Kincaid and Griffiths, 2003, 2004]. Estimates of trench motion in Central America vary depending on the plate motion model used from roughly 2 cm/yr of trench retreat to 2.5 cm/yr of trench advance [Lallemant *et al.*, 2008; Schellart *et al.*, 2008]. However, for plate motion models where along-arc

variations in trench motion exist, the trench in northern Nicaragua tends to be moving seaward relative to the trench in southern Costa Rica [Schellart *et al.*, 2008].

A variety of upper plate features have been proposed as evidence for the evolution of slab shape and subduction zone flow. Overall, the position of the arc has migrated trenchward (roughly southwest) since the early Miocene, with the largest offsets near the northern border of Nicaragua with Honduras [Weinberg, 1992; Plank *et al.*, 2002; Alvarado *et al.*, 2007]. Since the middle Miocene, the arc has been relatively stationary in Costa Rica, but continued trenchward offsets occurred in Nicaragua [Weinberg, 1992; Plank *et al.*, 2002], and have been attributed to slab roll-back or an increase in subduction angle [Weinberg, 1992]. These changes in arc position are consistent with the trench migration pattern in Schellart *et al.* [2007]. The recent and present-day arc in Nicaragua is located in the Nicaragua Depression, a large lowland area containing Lake Managua and Lake Nicaragua. Sediment thicknesses have been estimated at 2 to 2.5 km from potential field data [Elming and Rasmussen, 1997] and at roughly 3 km from receiver function modeling of TUCAN Ps waveforms [MacKenzie *et al.*, 2008]. Weinberg [1992] proposed that the 50-km wide depression formed in the Pliocene to Middle Pleistocene, possibly as a dynamic response of the upper plate to slab steepening.

However, despite this history, recent extension and subsidence appear to be small. GPS data indicate that present-day upper plate deformation is dominated by the northwestward translation of a fore-arc sliver that includes the Nicoya Peninsula and the Pacific coast of Nicaragua [Lundgren *et al.*, 1999; Iinuma *et al.*, 2004; Norabuena *et al.*, 2004; Turner *et al.*, 2007] at rates of 7-8 mm/yr in Costa Rica [Norabuena *et al.*, 2004] and an average of 15 mm/yr in Nicaragua [Turner *et al.*, 2007]. Arc-normal extension

occurs only northwest of Lake Managua (5 mm/yr on average) with a small component of arc-normal compression in southeastern Nicaragua [Turner *et al.*, 2007]; again, such a pattern is consistent with the signs of the differential trench motion rates estimated by Schellart *et al.* [2007]. In regard to recent subsidence, data from a borehole on Momotombo volcano in the middle of the Nicaragua Depression indicated only 150 m of Quaternary volcanoclastic sediments on top of Tertiary volcanics [van Wyk de Vries, 1993].

Geochemical and isotopic variations also exist along strike in this subduction zone. The presence of Ocean Island Basalt (OIB) signatures related to the Galapagos plume in Costa Rican and Nicaraguan arc magmas is particularly relevant to anisotropy, as it has been used to argue for flow in the mantle wedge parallel to the slab. Proposed scenarios include flow of Pacific mantle from beneath the Nazca slab in South America that enters the wedge beneath Panama [Russo and Silver, 1994; Herrstrom *et al.*, 1995] or flow into the wedge from a slab window beneath central Costa Rica [Abratis and Wörner, 2001]. Alternative models invoke melting of mantle that was influenced by the Galapagos plume during the Cretaceous formation of the Caribbean Large Igneous Province [Feigenson *et al.*, 2004], or melting of eroded fore-arc complexes that had been accreted from the Cocos Ridge [Goss and Kay, 2006]. Recently, mixing trends defined by $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios have pointed to the Galapagos seamounts, which are subducting beneath central Costa Rica (Fig. 1), as the source of the OIB signature in volcanic-front magmas from central Costa Rica and Nicaragua [Hoernle *et al.*, 2008]. This model indicates northwest flow in the mantle wedge at a minimum velocity of 63-190 mm/yr [Hoernle *et al.*, 2008], whereas the Abratis and Wörner [2001] model implies flow at

roughly 40 mm/yr. Regardless, both rates are significant relative to the rate of subducting plate motion (~ 85 mm/yr [*DeMets*, 2001]).

3. Shear-wave Splitting Observations

We analyze shear-wave splitting from local events, and this approach has both advantages and disadvantages for studying anisotropy in the mantle wedge. Unlike SKS waves, which are sensitive to anisotropy from the core-mantle boundary to the station, local S-waves, generated by earthquakes in the subducting slab, sample anisotropy only in the slab itself, the mantle wedge, and the overriding lithosphere and crust. Incidence angles for SKS waves are typically close to vertical, but for local phases they can range from vertical to nearly horizontal. However, in order to avoid waveform complications from the free-surface and shallow interface phase shifts and converted phases, we restrict our analysis to events within the “shear wave window” ($<35^\circ$ from vertical for straight-line ray paths). Because shear-wave splitting is dependent on propagation (and polarization) direction, this broader range of incidence angles (even when limited to the shear-wave window) allows local S-waves to detect variations in anisotropic structure which SKS waves cannot. One downside to the local events used in this study is that they are generally small (magnitudes between 2.7 and 5.4), and therefore, the initial polarizations are unknown because source mechanisms are not readily available. This is in contrast to SKS-waves, where upon conversion to a shear-wave at the core-mantle boundary, all

shear energy is in the radial direction, making the initial polarization parallel to the azimuth of the ray path.

3.1. Splitting Analysis Method

We apply a band pass filter (0.01-2.0 Hz) to each waveform, a window 10-20 s before and ~ 1.5 -2 periods after the S-wave arrival is manually cut, and shear-wave splitting parameters (i.e., fast horizontal S-wave polarization azimuth, ϕ , and splitting time, dt) are calculated from the windowed waveform using the eigenvalue minimization method [Silver and Chan, 1991]. This technique attempts to linearize the horizontal components of particle motion by performing a grid search over all possible azimuths and a range of splitting times (Fig. 2a); we use -90° to $+90^\circ$ at 1° increments for ϕ and 0 to 5 seconds at 0.01 sec increments for dt . The horizontal components (Fig. 2b) are rotated and time-shifted (Fig. 2d) for each trial combination of splitting parameters, and a cross-correlation matrix is calculated. This matrix would possess only one non-zero eigenvalue if the resulting particle motion (corrected for splitting parameters) were to be perfectly linear, and the corresponding eigenvector would characterize the direction of linear particle motion. However, complexities can arise from background noise, lateral heterogeneities, and the possibility that the seismic wave being analyzed may have sampled multiple layers of anisotropy [e.g., Silver and Savage, 1994]. In practice, there will be two non-zero eigenvalues, and the best-fitting pair of splitting parameters is the one that minimizes the second eigenvalue [Silver and Chan, 1991] and produces the most linear particle motion (Fig. 2e). Therefore, splitting parameters determined with this method are apparent splitting parameters that approximate the integrated effects of all anisotropic

structure along the ray path. Errors (σ_ϕ and σ_{dt}) are measured on the 95% confidence contour at the maximum distance in the ϕ and dt directions, independently, from the best-fitting pair of splitting parameters (Fig. 2a).

3.2. Data Set

Shear-wave splitting was measured in local-S phases recorded by the stations of the TUCAN Broadband Seismic Experiment. The TUCAN array was operational from July 2004 to March 2006 and consisted of 48 broadband stations that formed a fore-arc, back-arc, and two cross-arc lines (Fig. 1), with station spacing varying between 10 km and 50 km. Splitting was also measured at three permanent stations: JTS (Global Seismograph Network, IRIS-Scripps Institution of Oceanography), HDC (GEOSCOPE, Institut de Physique du Globe de Paris), and BOA (GEOFON, Program of GFZ Potsdam). From ~20 months of data, we analyzed 8581 local-S waveforms, resulting in 791 high quality splitting measurements (Fig. 3 and Supplementary Table 1). We do not define “high quality” splits exclusively by the size of the 95% confidence region, but rather, each measurement is assessed based on four attributes. A “high quality” splitting measurement (1) is made from a clear phase with a high signal-to-noise ratio, (2) displays elliptical or circular particle motion, (3) has a small, single 95% contour (preferably enclosed by multiple contours), and (4) yields roughly linear particle motion when the waveform components are corrected for splitting. See Table S1 (Supplementary Information) for a complete spreadsheet of event, station, and splitting parameter information.

This data set is complex and shows a high degree of spatial variability in both ϕ and dt , not unlike other subduction zones [e.g., *Morley et al.*, 2006; *Pozgay et al.*, 2007]. There is no obvious simple correlation between event characteristics (e.g., depth, path length, across-arc/along-arc position) and splitting fast direction (Fig. 4), making it difficult for any simple 1-D or 2-D model of anisotropy to account for all observations. At first glance, it may be tempting to attribute such a distribution to short wavelength structural variations in the subducting or overriding crust/lithosphere. However, path segments through the subducting slab do not typically constitute a significant fraction of the total path length, and delay times show a clear increase with increasing path length when averaged in depth bins (Fig. 5), indicating the presence of anisotropy in the mantle wedge beneath the overriding plate.

Despite the complexity of the data set as a whole, sorting by depth and geographic location does reveal underlying structure in the measurements. In the northwest Nicaraguan fore-arc, a transition from arc-normal fast directions in the extreme fore-arc to arc-parallel fast directions towards the arc occurs (Fig. 6a); this pattern becomes less apparent moving along the arc towards and into Costa Rica (Fig. 3). Another example is in the Nicaraguan back-arc where measured fast directions rotate from arc-parallel in central Nicaragua to more arc-normal in the northwest (Fig. 6b). One of the challenges posed by a large, complicated shear-wave splitting data set, such as this one in Central America, is finding a model of anisotropy that can account for short-wavelength variability and, sometimes, even contradictory splitting observations at similar locations. We therefore turn to tomography to produce a model of anisotropic structure that best explains the splitting observations.

4. Shear-wave Splitting Tomography

If three-dimensionally heterogeneous anisotropic structure exists above the local-S sources, then the large scatter of ϕ and dt , when plotted at a single point (e.g., hypocenter, ray path midpoint, station) is not surprising considering that ray paths with a common point may sample distinctly different structure outside the shared region and that ϕ and dt are dependent on path length and both propagation and polarization direction. The spatial averaging technique developed by *Audoine et al.* [2004], and later utilized by *Nakajima et al.* [2006] and *Pozgay et al.* [2007], begins to address the possibility of heterogeneous anisotropy in subduction zones by projecting individual splitting parameters to the surface, but does not incorporate the effects of multiple layers of anisotropy on predicted particle motions.

The distribution of local seismic sources and receivers in Nicaragua-Costa Rica (Fig. 7) and the large number of high quality splitting measurements (Fig. 3) provide the opportunity to employ the use of a formal inversion to resolve anisotropic structure in this subduction zone. A more detailed discussion and validation of this technique is described in *Abt and Fischer* [2008], but the principle assumptions regarding mantle composition, elastic properties, and Earth structure are restated here. Brief explanations of the forward modeling and inversion are also provided before presenting the modeling results.

4.1. Mineralogy

A model mantle mineralogy of 70% olivine and 30% orthopyroxene (opx) was assumed, and we used the elastic constants of olivine ($\text{Mg}_{1.8}\text{Fe}_{0.2}\text{SiO}_4$) [*Anderson and*

Isaak, 1995; Abramson et al., 1997] and opx ($\text{Mg}_{0.8}\text{Fe}_{0.2}\text{SiO}_3$) [*Frisillo and Barsch, 1972*], including their pressure and temperature derivatives (see Table 1 in *Abt and Fischer [2008]*). Naturally deformed lithospheric peridotites [e.g., *Mainprice and Silver, 1993*] exhibit orthorhombic symmetry in both olivine and opx and a LPO in which the a -axis [100] of olivine is aligned parallel to the c -axis [001] of opx (b -axis [010] of olivine is parallel to the a -axis [100] of opx). Adding the stiffness tensors ($0.7 * c_{ijkl}^{\text{olivine}} + 0.3 * c_{ijkl}^{\text{opx}}$) in this orientation provides the elastic properties for our orthorhombic model mantle material.

Although olivine and opx are intrinsically orthorhombic, several studies have demonstrated that crystal aggregates display hexagonal symmetry, rather than purely orthorhombic symmetry [*Mainprice and Silver, 1993; Bystricky et al., 2000; Mehl et al., 2003; Michibayashi et al., 2006*]. For example, *Bystricky et al. [2000]* and *Michibayashi et al. [2006]* observe LPO in which olivine a -axes are aligned ~parallel with the shear direction and the b - and c -axes form a “girdle” perpendicular to the shear direction (i.e., D-type slip, $\{0kl\}[100]$). In order to simulate hexagonal symmetry using the orthorhombic single-crystal elastic constants of olivine and opx, the coefficients corresponding to the b - and c -axes must be combined. We prescribed the olivine a -axis [100] (opx c -axis [001]) as the axis of symmetry and reduced the nine independent orthorhombic coefficients to five independent hexagonal coefficients [*Montagner and Anderson, 1989*]. As discussed in Section 4.4, here we only present models that assume this hexagonal symmetry.

We made the simplifying assumption that the entire model is composed of the olivine-opx material described above and, therefore, do not account for crustal

mineralogies or the elastic properties of cracks/fractures at crustal depths. A more realistic characterization of crustal anisotropy is one aspect of the modeling that could be addressed in the future to help isolate the effects of mantle wedge fabric.

4.2. Model Parameterization

The model region (i.e., a volume encompassing all events and stations) was parameterized into blocks 25 km on each side (Fig. 8). Crystallographic orientation in each model block was set by two (θ, ψ - hexagonal case) or three (θ, ψ, γ - orthorhombic case) angle parameters. θ represents the horizontal azimuth of the [100]-axis, ψ is the plunge of the [100]-axis from horizontal, and γ is the plunge of the [001]-axis (Fig. 8). We employed a scalar parameter, α , to allow for variable strength of anisotropy. This parameter varies from 0% to 100% and reduces the effect of anisotropy on predicted delay time in each model block relative to that from its single crystal value. Conceptually, α can represent a percentage of randomly oriented grains mixed with grains aligned by the specified orientation parameters for a particular block.

4.3. Forward Calculation

We calculated predicted shear-wave splitting parameters (ϕ, dt) for each ray path in the data set (Fig. 7) using a particle motion perturbation method [e.g., *Rümpker and Silver, 1998; Fischer et al., 2000*]. The effects of three-dimensionally varying anisotropy on the horizontal components of shear-wave particle motion were approximated by progressively rotating and time-shifting a simple wavelet according to the eigenvectors (polarization directions) and eigenvalues (phase velocities) of the Christoffel matrix

[*Babuska and Cara, 1991*] for each model block touched by a ray. (See *Abt and Fischer [2008]* for more details. Note the typographical error in the indices of equation 1 in *Abt and Fischer [2008]*. The Christoffel Equation should be: $m_{ik} = 1/r(z) * c_{ijkl} * n_j * n_l$). This technique accounts for the order in which anisotropic structure is encountered by a wave and is sufficiently efficient to allow for its repetitive use in calculating partial derivatives needed for the inversion. The particle motion perturbation approach does not consider Fresnel zones of the local S phases. However, a comparison [*Abt and Fischer, 2008*] of the splitting parameters calculated with this approach to those obtained from full-waveform pseudospectral synthetics [*Hung and Forsyth, 1998*] for moderately large variations in anisotropy showed they were in close enough agreement that any inaccuracy in predicted splitting parameters introduced by the particle motion perturbation method will likely be much smaller than the uncertainties typical of the observed shear-wave splitting parameters inverted in this study.

4.4. Inversion

A best-fitting model of anisotropy (i.e., crystallographic orientation and strength of fabric alignment) was determined by minimizing the misfit between observed and predicted shear-wave splitting parameters through an iterative, damped, linearized, least-squares inversion [*Tarantola, 1987*]. The effects of model parameter changes ($\delta\theta, \delta\psi, \delta\gamma, \delta\alpha$) on predicted splitting parameters ($\delta\phi, \delta dt$) were calculated numerically by simple finite difference. However, to account for non-linearities in the dependence of splitting on model parameters, we recalculate partial derivatives when model parameters changed by more than 1° (for θ, ψ, γ) or 1% (for α).

The data were weighted by their individual errors (Section 3.1, Table S1) and the inversion was damped in order to keep changes in model parameters small enough such that the linear partial derivatives used in each iteration are valid. The *a priori* variance was increased (i.e., damping was decreased) after initial model convergence to test the stability of the model. . Based on tests of inversion accuracy with synthetic data [*Abt and Fischer, 2008*], inversions were run for a total of 100 iterations, with variance increased from 1 (degrees or %) to 10 after iteration 40. The event and station distribution in the current dataset is such that regions of the back-arc contain fewer crossing rays than beneath the arc and fore-arc (Fig. 7), resulting in reduced resolution in the back-arc. To allow modeled crystallographic orientations to be interpreted, at a coarser resolution, in larger portions of the back-arc, we combined multiple individual 25^3 km^3 blocks into larger effective volumes following convergence of the model after damping relaxation. The extent of these larger volumes is shown in Figure 9, and their incorporation into the inversion was accomplished by averaging crystallographic orientation and strength of anisotropy in each block within a volume and then introducing constraints into the inversion that forced subsequent model parameter changes in blocks within a volume to be the same. This approach is described in more detail in *Abt and Fischer [2008]*. The large volume constraints were emplaced after iteration 70 with an associated variance value of 0.1. This variance value controls the degree to which the individual blocks within a volume are forced to maintain the same parameters (i.e., act as one coherent block). Relative to the splitting data variances, which are the errors from the splitting measurements, a value of 0.1 is sufficiently small for the large volume constraints to

dominate over any change in the model parameters for an individual block that would otherwise be required by trying to match the observed splitting parameters alone.

Although we conducted inversions with all four model parameters ($\alpha, \theta, \psi, \gamma$) assuming orthorhombic symmetry, we found that γ , the parameter controlling c -axis dip, is less well-resolved than the other two angle parameters. Furthermore, there are only minor differences in the predicted splitting parameters between an LPO with hexagonal symmetry and orthorhombic symmetry with the c -axis fixed in the horizontal plane ($\gamma=0$). Therefore, we present results from inversions that assumed hexagonal symmetry and solved for three parameters: fabric strength (α) and the azimuth (θ) and dip (ψ) of the symmetry axis (i.e., the olivine a -axis).

An “average” starting model was used for the inversion of the real data. This model was generated by setting θ in each block to be the average fast direction of all rays that touch that particular block and by setting $\psi=0$ (i.e., horizontal a -axis) and α to be that predicted from the average delay times of all rays that touch that particular block. (See Appendix A in *Abt and Fischer* [2008] for a full description). The starting model assumed does affect the inversion results, but during testing of the tomographic method with synthetic data we found that inversions with the average starting model, used in concert with the inversion parameter values given above, retrieve synthetic known structures with the greatest accuracy [*Abt and Fischer*, 2008].

4.5. Tomographic Modeling Results

The final model of crystallographic orientation obtained from the inversion is first shown in Figure 10. Model parameter resolution was determined with the resolution

matrix [Tarantola, 1987], and in Figure 10a we show results for model blocks which have resolution matrix diagonal element values greater than 0.25 for all model parameters. In this view, the a -axes of olivine are projected onto horizontal planes, thereby showing azimuthal orientation but not dip. Arc-parallel and nearly arc-parallel a -axes dominate a significant portion of the model space. Three separate “pockets” of arc-normal a -axes exist, but as demonstrated in section 4.6, they are not all uniquely resolved. The dips of olivine a -axes are revealed in a pseudo 3-D view of the model (Fig. 11). In general, a -axes (arc-parallel, arc-normal, and intermediate) are close to horizontal, with only a slight tendency towards dipping to the north-northwest. As demonstrated by *Abt and Fischer* [2008], a -axis dip is partially resolvable, particularly beneath the two arc-normal station lines where crossing ray coverage is most dense. Therefore, along with the dominantly arc-parallel a -axis azimuths, the overall roughly horizontal orientation of a -axes displayed in Figure 11 is also a significant feature of the model. The accuracy of a -axis azimuth and dip recovery is discussed further in section 4.6.

Over the course of the inversion, both average fast direction misfit and delay time misfit decrease relatively smoothly until reaching the point where the model discretization is effectively coarsened in the back-arc, with no indication of instability caused by the relaxation of damping (Fig. 12). When the larger volumes are first added to the inversion (iteration 70), the model parameters from the individual blocks contained in a single larger volume are averaged, resulting in a jump in average data misfit. As the inversion progresses, average delay time misfit returns to the same level (~ 0.08 s), while average fast direction misfit retains a slightly higher value than before the volumes were

generated ($\sim 31^\circ$ vs. $\sim 29^\circ$). In the final model, the 95% confidence limits for the observed and predicted fast directions either overlap or are mismatched by less than 8° for 50% of the data, with mismatches of less than 28° for 75% of the data. Relative to their corresponding starting values, the final fast direction confidence limit mismatches represent a reduction of 22% for 50% of the data and 8% for 75% of the data. The 95% confidence limits for the observed and predicted times either overlap or are mismatched by less than 0.01 s for 50% of the data, and by less than 0.07 s for 75% of the data; these values represent 85% and 52% reductions from starting values, respectively.

4.6. Recovery Tests

We tested the uniqueness of our model (Figs. 10,11) by conducting two recovery tests in which this model is used as a “known” input model of anisotropy. In conjunction with the characterization of resolution from the resolution matrix, comparing the recovered structure from these inversions with the “known” model allowed us to better define the location(s) in the model space that are particularly well-constrained. Additional tests using simple input structures and synthetic data for the paths in the TUCAN dataset were carried out by *Abt and Fischer* [2008]. They found that structures with lateral scales of 50-75 km are at least partially recovered down to depths of 100 km, particularly in the portions of the model space with the densest crossing ray coverage near the two arc-normal lines. At greater depths and at horizontal distances of more than 50 km into the back-arc, spatial resolution rapidly decreases, particularly where the larger volume constraints are imposed.

The first recovery test uses an average starting model, generated from the predicted splitting parameters in the same fashion as with the real data, and the inversion is run with the same parameters as the real data inversion (i.e., iterations, damping relaxation, larger volumes generated). In *Abt and Fischer* [2008], resolution testing with synthetic models showed that input models were typically best retrieved when average starting models such as this one were employed. The final model from this recovery test is displayed in Figure 13a as the pseudo 3-D figure and in Figure 14a with the a -axes projected onto horizontal planes. Overall, the patterns of a -axis orientation in the input model and the retrieved structure are similar, although a few blocks show large differences. In particular, a -axis orientation is accurately recovered in much of the mantle wedge, as shown by the yellow and green regions in Figure 14b, which indicate angular misfits of less than 35° between input and retrieved model a -axis orientation. Angular misfits include both azimuth and dip variations of a -axes. In some regions angular misfit is large ($>50^\circ$, red blocks in Fig. 14b). These zones occur where crossing ray coverage is sparser, for example at the edges of the study region and in the middle of the array far from the two arc-normal lines of stations. The misfit in strength of anisotropy is less than 10% over most of the model (Fig. 14c).

The second recovery test is identical to the first except for the use of a uniform starting model in which a -axes are horizontal and oriented N-S. The retrieved model from this recovery test is shown in Figure 13b, and the starting orientation is plotted for reference in the eastern corner of the map view cross-section. As with the average starting model, retrieval of the input model is less accurate in more poorly sampled regions (e.g., the edges of the array and in between the arc-normal lines). The

distribution and magnitude of angular misfit in this test (not shown) differs somewhat from the recovery test with the average starting model, particularly in poorly sampled regions, due to the fact that the average starting model is initially closer to the input structure. However, where sampling of the model space is good, the inversion with the uniform starting model also converges toward the input structure with no apparent tendency towards the starting orientation (Fig. 13b). This inversion behavior indicates significant sensitivity of shear-wave splitting to the model parameters, despite the non-linear aspects of the forward calculation.

At broad scales (hundreds of kilometers) within the better-sampled regions, the results of the two recovery tests (Figs. 13,14) and the known structure (Figs. 10,11) agree fairly well. Most notably, the arc-parallel and sub-arc-parallel orientations of a -axes beneath the arc and in the central portion of the back-arc mantle wedge beneath 75 km are consistently imaged. In addition, the pocket of roughly arc-normal a -axes in northwest Nicaragua between 25 and 75 km depth is a stable feature in the models (Fig. 13). The inversion with the average starting model retrieves the input structure somewhat more accurately because its initial deviations from the input structure are less extreme, a finding that agrees with the resolution testing conducted by *Abt and Fischer* [2008].

Azimuths of olivine a -axes are in general better resolved than a -axis dips. For a single shear-wave on a vertical path, similar splitting parameters may be predicted by a steeply dipping a -axis with a large anisotropic strength and a horizontal a -axis and a smaller strength. However, where phases sample a block over a wide range of back-azimuths and incidence angles, this non-uniqueness is at least partially mitigated. *Abt and Fischer* [2008] showed that a -axes with a 30° dip could be distinguished from a

horizontal orientation down to depths of 75 km over broad regions of the fore-arc and arc, and at least 75 km into the back-arc. At greater depths, this accuracy of dip retrieval was limited to blocks beneath the intersection of the arc-parallel and arc-normal lines of stations. In the recovery tests shown in Figures 13 and 14, the accuracy of a -axis dip retrieval varies significantly, but is better than 35° over significant regions of the model space (see Figure 14b, which shows the combined recovery of a -axis dip and azimuth). In the model obtained by real data inversion (inset to Figure 11) and its recovery test using the average starting model (inset to Figure 14a) the majority of a -axis dips are within 20° of horizontal. Therefore, in an overall sense we conclude that anisotropy resolved in the Nicaragua-Costa Rica mantle wedge can be characterized by a roughly horizontal fast symmetry axis.

Because a -axis dips in general are fairly small, we focus our interpretation on broad, well-resolved azimuthal patterns, in particular, differences between roughly arc-parallel or arc-normal a -axis azimuths in better-sampled model regions. To delineate the portions of the model space which are most robust and therefore interpretable, we chose regions where resolution matrix diagonal element values from our final model are greater than 0.25 (Fig. 10a) and axis misfit between this model and the model from the recovery test with the average starting model is less than 35° (Fig. 14). Even with this strict filter, roughly arc-parallel a -axes are still apparent in the mantle wedge beneath the arc and back-arc to depths of at least 125 km (Fig. 10b). The pocket of arc-normal a -axes beneath northwest Nicaragua between 25 and 75 km depth observed in the data inversion is replicated in the recovery tests with an azimuth misfit of 35° , but because of a modest discrepancy in dip, a -axes in this pocket have total angular misfits of $\sim 45^\circ$ with the input

structure and are therefore left out of Figure 10b. Additionally, there is a strong similarity between these large-scale structures (dominantly arc-parallel a -axes and the arc-normal pocket) in the model and the pattern of observed splitting parameters (Figs. 3,6).

Some studies [e.g., *Faccenda et al.*, 2008] have proposed that the subducting slab is strongly anisotropic and that slab anisotropy dominates SKS splitting in subduction zones. In the tomographic models obtained in this study from local-S splitting, slab anisotropy would be folded into the model blocks that lie below or contain the slab surface inferred from relocated hypocenters [*Syracuse et al.*, 2008]. Anisotropy within the slab is not in general clearly constrained, as demonstrated by the lack of “slab” blocks with well-resolved model parameters in Figures 10 and 14. This result is explained by the fact that most local-S paths in the slab are fairly short, but it also indicates that the strength and geometry of slab anisotropy are such that they do not make a large overall contribution to the local-S splitting. This finding echoes the correlation of splitting time with wedge path length (Fig. 5), which in an average sense also shows that the local-S splitting is dominated by wedge anisotropy.

5. Discussion

5.1. Interpreting Upper Mantle Deformation from Anisotropy Models

The primary goal of this study is to place constraints on the pattern of flow in the mantle wedge beneath Nicaragua and Costa Rica using the model of anisotropy obtained

by inverting local-S shear-wave splitting observations. The main features of this model are roughly horizontal and arc-parallel olivine a -axes in the mantle wedge, with the suggestion of a zone of arc-normal a -axes beneath northwest Nicaragua (Fig. 10). The direction of shear can only be inferred from modeled a -axis orientations if a relationship between the two is assumed (e.g., A-type slip, where olivine a -axes align parallel to shear), but this relationship may not necessarily remain constant as conditions (e.g., pressure, temperature, melt, water) in the mantle wedge change. Along with the standard shear-parallel a -axis interpretation [e.g., *Zhang and Karato*, 1995; *Ismail and Mainprice*, 1998; *Bystricky et al.*, 2000] we consider both B-type fabric [*Jung and Karato*, 2001; *Jung et al.*, 2006] and melt band formation [*Holtzman et al.*, 2003] as possible deviations from the simple A-type fabric (Fig. 15).

The conditions under which B-type fabric is expected to dominate (i.e., high stress, moderate-to-high water content) are restricted to the extreme wedge corner where isolation of the region may occur due to a reduction in viscous coupling of the convecting mantle with the down-going plate [*Kneller et al.*, 2005; *Kneller et al.*, 2007]. Although *Kneller et al.* [2007] demonstrate that the exact across-arc location of this transition will depend on several parameters (e.g., slab dip, slab age, slab hydration, wedge rheology), B-type fabric is not expected to exist in the higher temperature (lower stress) mantle wedge beneath the arc and back-arc (Fig. 15a,b, right sides of each diagram). If B-type fabric does exist, then a transition in LPO should occur within the fore-arc wedge along with a corresponding transition in observed fast direction orientations at the surface. For 2-D corner flow, such a transition would predict arc-parallel fast directions in the fore-arc changing to plate motion parallel in the arc and back-arc, a pattern that is clearly

observed in Honshu and Ryukyu [Nakajima and Hasegawa, 2004; Long and van der Hilst, 2005; Nakajima *et al.*, 2006].

The presence of melt in deforming olivine samples has also been shown to influence the orientation of fabric relative to shear. When additional “pinning” phases (e.g., chromite or FeS) are present in peridotite, a reduction in permeability occurs and results in the segregation of melt into bands separated by relatively melt-free crystal lenses [Holtzman *et al.*, 2003]. As the material (solid+melt) deforms, strain is partitioned into the melt bands, causing compaction of the crystal lenses and subsequent strain oriented approximately perpendicular to the large-scale shear direction. Thus, the primary implication of oriented melt bands, if the a -axes of olivine in the crystal lenses align sub-parallel to shear, is that the fast seismic direction under these conditions will be perpendicular to overall flow. It is unclear, however, how long these effects on LPO will remain because the melt bands tend to migrate through a region, leaving the crystal lenses behind to have their LPO reset to that imparted by the dominant flow field. Nonetheless, even if melt bands are transient features in the wedge, anisotropy produced by the bands themselves (i.e., SPO) could be significant, and could also produce a fast seismic polarization normal to overall flow, if the melt bands are oriented sub-vertically. The most likely location for the high melt fractions required for melt band formation, and thus deviations in a -axis orientations from flow-parallel, is beneath the volcanic arc (i.e., arc-parallel fast directions close to the arc and arc-normal fast directions in the fore-arc and back-arc; see Fig. 13 in Abt and Fischer [2008] for predicted splitting parameters from such a fabric geometry).

5.2. Proposed Arc-Parallel Mantle Wedge Flow Beneath Costa Rica and Nicaragua

The observed pattern of shear-wave splitting fast directions and tomographically imaged anisotropic structure strongly suggest that an arc-parallel-fast fabric exists over broad regions of the Central American mantle wedge. The most straightforward explanation for this fabric is that it represents olivine *a*-axes aligned with roughly arc-parallel flow. Both arc-parallel fast directions (Figs. 3,6) and model *a*-axes (Figs. 10,11) extend into the back-arc, making 2-D corner flow an unlikely scenario, even with B-type fabric. In addition, the subset of splitting measurements in the northwest Nicaraguan fore-arc display fast directions which rotate from arc-normal closer to the trench to arc-parallel beneath the arc (Fig. 6a), opposite of the observed trend in Japan [Long and van der Hilst, 2005; Nakajima *et al.*, 2006]. This pattern would be expected for along-arc flow and B-type fabric in the extreme wedge corner (Fig. 15b). However, while this pattern in the data is reflected in the inversion (Fig. 10a), it is not robustly retrieved (Fig. 10b), largely because this region lies at the edge of the model. Nor is this pattern as clearly observed in other along-arc segments. Along-arc flow is schematically depicted relative to the anisotropic model in Fig. 10b. Mehl *et al.* [2003] observed a horizontal foliation plane concurrent with arc-parallel *a*-axes in the accreted Talkeetna arc in Alaska, but our current data in Central America cannot distinguish between a horizontal and vertical shear plane; therefore, we illustrate both possibilities (Fig. 10b).

Could 2-D corner flow still be possible in the mantle wedge if melt effects are considered, such as viscosity variations associated with concentrated melt bands that re-orient olivine LPO so that *a*-axes are arc-parallel [e.g., Holtzman *et al.*, 2003] (Fig. 15a, left side), and/or the SPO anisotropy of the melt pockets/bands themselves [Zimmerman

et al., 1999; *Holtzman et al.*, 2003]? Although this type of model cannot be ruled out on seismological grounds alone, it seems unlikely that sufficiently large melt fractions exist throughout the entire mantle wedge. In particular, high V_P/V_S ratios, thought to represent melt rising to the arc from near the top of the slab in Nicaragua [*Syracuse et al.*, 2008] are confined to a tabular volume beneath the arc, and do not extend into the back-arc wedge where arc-parallel-fast anisotropy is also observed (Figs. 10 and 11). This interpretation is also consistent with the concentration of present-day magmatism at a volcanic front that is narrow in the arc-normal direction, although small amounts of magmatism do occur further into the back-arc [e.g., *Walker et al.*, 1995].

The implications of the three pockets of arc-normal a -axes at mantle wedge depths in our model (Figs. 10,11) are not entirely clear. These features are reflected in the shear-wave splitting measurements (Figs. 3 and 6), but only one of these zones is retrieved with total angular misfits of less than 35° in the recovery tests (Fig. 10b), and its location at the very northern edge of the model still makes it somewhat suspect. If these arc-normal zones are real, they may represent a small-scale variation in the flow pattern. For example, upwellings (e.g., buoyant diapirs [*Hall and Kincaid*, 2001; *Behn et al.*, 2007] or downwellings (e.g., foundering arc lithosphere [*Behn et al.*, 2007]) could perturb an otherwise predominantly arc-parallel flow field, as could variable degrees of mantle entrainment with the subducting plate. Another possibility is that these arc-normal volumes represent a localized zone with particularly high melt content. However, none of them are spatially correlated with the volumes of highest V_P/V_S imaged by *Syracuse et al.* [2008].

5.3. Geochemical Evidence for Arc-Parallel Flow

Geochemical data provide key evidence that along-arc mantle wedge flow is the correct interpretation for the observed arc-parallel-fast anisotropy. Northwest arc-parallel flow has been invoked to explain a variety of along-arc geochemical variations [Herrstrom *et al.*, 1995; Abratis and Wörner, 2001; Hoernle *et al.*, 2008], although, as described in the introduction, alternative models have also been proposed [Feigenson *et al.*, 2004; Goss and Kay, 2006].

Hoernle *et al.* [2008] trace the source of high $^{208}\text{Pb}/^{204}\text{Pb}$ ratios and low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in recent (< 6 My) arc volcanics in central Costa Rica to the Galapagos Seamounts that lie on the subducting Cocos Plate (Fig. 1). The $^{208}\text{Pb}/^{204}\text{Pb}$ ratios decline and the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios increase from central Costa Rica to the northwest, until they reach MORB-like values in northwest Nicaragua that match the composition of the locally subducting Cocos Plate crust. These results indicate northwest flow of mantle bearing the seamount signature and its progressive dilution as it mixes with more typical mantle wedge peridotite. Constraints on the time at which the seamount signature was first introduced into the wedge suggest mantle wedge flow at a minimum velocity of 63-190 mm/year, comparable to the subduction rate of the Cocos Plate (~85 mm/yr, DeMets [2001]) [Hoernle *et al.*, 2008]. This model could also help to explain other geochemical trends observed along the arc, for example the northwestward increase in Ba/La, U/Th, and Ba/Th and decrease in La/Yb [e.g., Herrstrom *et al.*, 1995; Patino *et al.*, 2000; Carr *et al.*, 2003], if the Galapagos seamount component had melt-like properties when it entered the wedge [Hoernle *et al.*, 2008]. However, these ratios could also reflect variations in sediment signatures entering the wedge locally with slab fluids (Ba/La,

U/Th, and Ba/Th) or the local degree of melting (La/Yb), as has been previously proposed [Herrstrom *et al.*, 1995; Patino *et al.*, 2000; Carr *et al.*, 2003].

5.4. Drivers for Arc-Parallel Flow and Implications for Mantle Depletion

Flow along the strike of the subducting Cocos Plate could be driven by interactions with the slab edge, slab roll-back, and oblique subduction (Fig. 10b). However, it is also possible that global mantle convection currents [e.g., Conrad and Lithgow-Bertelloni, 2004] contribute to flow that does not conform to the predictions of 2-D corner flow driven by local plate motion. Each of these sources of along-arc flow may not individually be strong enough to overcome the expected plate motion-parallel flow generated by entrainment with the subducting slab, but in concert they may constitute the necessary driving force, particularly if they are aided by weak slab-mantle coupling or other rheological factors.

Tectonic variations within the Nicaragua-Costa Rica subduction zone (Section 2) provide a basis for inferring possible local drivers for along-arc flow. Viscous flow modeling has shown that along-arc flow may be possible if slab roll-back occurs in the vicinity of a slab edge [Kincaid and Griffiths, 2003; Funiciello *et al.*, 2006], and this scenario has been proposed in other subduction zones (e.g., Kamchatka [Peyton *et al.*, 2001] and Tonga [Smith *et al.*, 2001]). The decrease in deep subduction-related seismicity [Protti *et al.*, 1994; Syracuse *et al.*, 2008] and the termination of active arc volcanism in southeast Costa Rica [Carr *et al.*, 2003] indicate the existence of the southeast edge of the Cocos Plate. Because trench motion rates depend on global plate motion reference frame, it is difficult to infer whether the trench in Nicaragua and Costa

Rica is in retreat or advance in an absolute sense [*Lallemand et al.*, 2008; *Schellart et al.*, 2008]. However, for models in which significant along-arc variations in trench migration rate exist [*Schellart et al.*, 2008], the trench in northern Nicaragua is 1-2 cm/yr in retreat relative to the trench in Costa Rica. This latter tendency in present-day trench motion correlates with the geologic record. The modern arc in northern Nicaragua is offset by more than 125 km trenchward relative to arc position in the early Miocene, a much greater shift than is found in Costa Rica [*Weinberg*, 1992; *Plank et al.*, 2002; *Alvarado et al.*, 2007]. Greater slab and trench roll-back in Nicaragua could draw wedge material to the northwest. In addition, whereas present-day extension occurs at an average rate of 5 mm/yr in northwest Nicaragua, arc-normal deformation shifts along the arc to compression in southeast Nicaragua [*Turner et al.*, 2007]. A slight obliquity of subduction [*DeMets*, 2001] may influence entrained flow within the wedge, perhaps by generating a toroidal component of motion [*Kneller and van Keken*, 2008], and northwest transport of the fore-arc sliver of the upper plate [*Turner et al.*, 2007] could also affect wedge flow.

Large changes in slab dip, for example as observed in the Nazca plate subducting beneath South America, can induce significant along-arc flow [*Kneller and van Keken*, 2007 and 2008]. However, this mechanism does not appear to be a dominant factor beneath Nicaragua and Costa Rica. First, changes in slab dip inferred from earthquake hypocenters and velocity tomography [*Syracuse et al.*, 2008] are less than about 15-20°, and initial numerical modeling incorporating this slab shape does not show significant deviations from 2-D corner flow [*P.E. van Keken, pers. comm.*, 2008]. Second, the modest decrease in slab dip occurs to the southeast, which would predict southeastern

along-arc flow, if any along-arc flow were produced. This prediction would contradict the northwest flow indicated by geochemical data [Herrstrom *et al.*, 1995; Abratis and Wörner, 2001; Hoernle *et al.*, 2008].

If along-arc flow is generated by interactions between slab shape and position, plate motions, and larger mantle flow patterns, the rheology of the mantle wedge and slab-wedge interface could help to sustain it with distance along the arc and in time. Decoupling of the mantle wedge from the subducting Cocos Plate may be caused by a low-viscosity layer along the slab interface, perhaps as a result of hydrous phases. Seismic evidence for low velocity layers in the subducting oceanic crust [Abers *et al.*, 2003; Abers, 2005] and mantle [Syracuse *et al.*, 2008] suggest that water may be carried to depths of 150 km or more. When released, these fluids may cause serpentinization or the creation of chlorite in the overlying mantle, as is observed in the fore-arc [DeShon and Schwartz, 2004], but also perhaps at greater depths. Although many subduction zone thermal models predict serpentine to be stable only in the shallow mantle wedge corner above the subducting plate [e.g., Bostock *et al.*, 2002; Hyndman and Peacock, 2003; Abers *et al.*, 2006; Arcay *et al.*, 2007; Cagnioncle *et al.*, 2007], and not above the slab interface at greater depths where we observe arc-parallel *a*-axes (down to 150 km, Fig. 10), the chlorite stability field may persist more deeply in a thin layer just above the slab [Cagnioncle *et al.*, 2007], thereby providing a possible means of decoupling the slab from most of the mantle wedge. In addition, these thermal models assume 2-D corner flow, and one of their basic characteristics is the upwelling of relatively hot back-arc wedge material into the wedge corner before it is dragged down with the plate. With along-arc flow near the slab, mantle wedge temperatures above the slab surface may be

significantly lower, thereby allowing both serpentine and chlorite to exist and decouple the slab from the wedge at greater depths. Broader low viscosity zones likely also exist further into the mantle wedge, produced by the hydrated wedge peridotite and partial melt seen in velocity and attenuation tomography [*Syracuse et al.*, 2008; *Rychert et al.*, 2008]. This type of low viscosity channel could help to focus along-arc flow, thus enhancing the alignment of olivine *a*-axes parallel to the arc [*Conder and Wiens*, 2007].

Along-arc flow in the mantle wedge would have a variety of implications for subduction zone melting models. It would alter slab and wedge temperature structure, as has already been demonstrated in a limited number of 3-D flow scenarios [e.g. *Kincaid and Griffiths*, 2004]. In addition, it could affect the degree of depletion of the mantle wedge material flowing into the fluid-fluxed region of potential melting beneath a given segment of the arc. In 2-D corner flow models, upwelling material from the back-arc wedge is likely to be less depleted and would therefore be more easily melted [e.g., *Cagnioncle et al.*, 2007]. However, if along-arc flow occurs in the mantle wedge where arc magmas originate, then a progressive depletion in the mantle source would be expected. Progressive depletion from Costa Rica to Nicaragua is consistent with geochemical trends [*Herrstrom et al.*, 1995], although it is not required given the possibility of along-arc variations in the degree of new melting or local input from the slab.

6. Conclusions

Tomographic inversion of shear-wave splitting observations reveals anisotropic structure that is not consistent with 2-D corner flow in the Central American mantle wedge. Well-resolved arc-parallel olivine *a*-axes are found in the mantle beneath the arc and back-arc, well outside the shallow (cold) wedge corner where B-type olivine fabrics may exist and align olivine *a*-axes normal to flow. The arc-parallel-fast fabric also occupies too broad a volume to be easily explained by the effects of partial melt. We therefore conclude that the wedge anisotropy is best explained by olivine *a*-axes aligned parallel to along-arc flow. This interpretation is corroborated by $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ trends that indicate northwest transport of material bearing the signature of the Galapagos seamounts at a minimum velocity of 63-190 mm/yr [Hoernle *et al.*, 2008]. Flow along the strike of the slab in the mantle wedge may be driven by the combined effects of slab roll-back, the slab edge in Costa Rica, and along-strike components of plate motion, although broader patterns of mantle convection could also play a role.

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8. List of Figures and Tables

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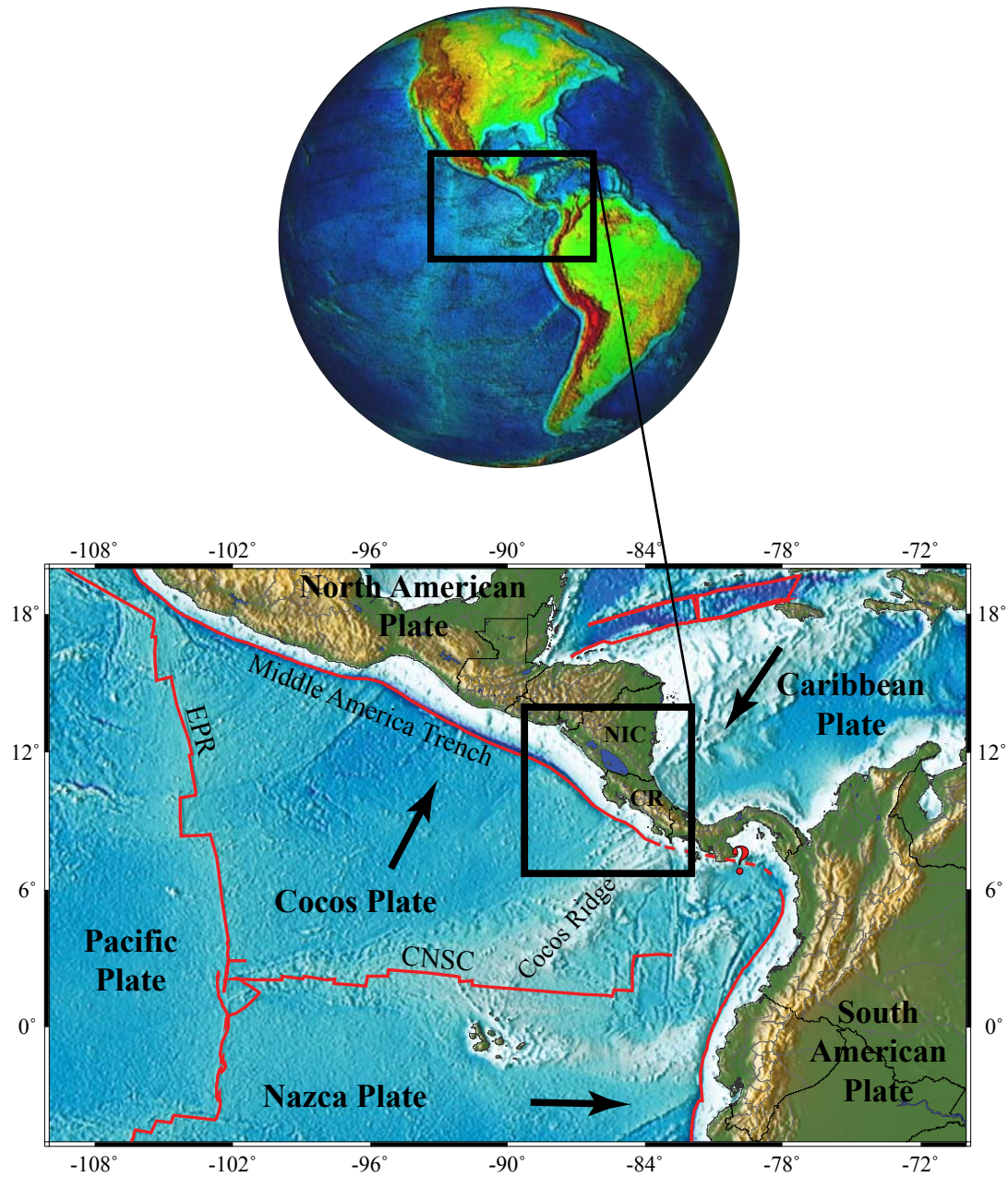


Figure 1a. The TUCAN study region is located at the southeast end of the Central American subduction zone. The Cocos Plate forms at both the East Pacific Rise (EPR) and the Cocos-Nazca Spreading Center (CNSC) and then subducts beneath the North American and Caribbean Plates at the Middle America Trench. The region outlined with the black rectangle (bottom) is shown in Figure 1b on the following page.

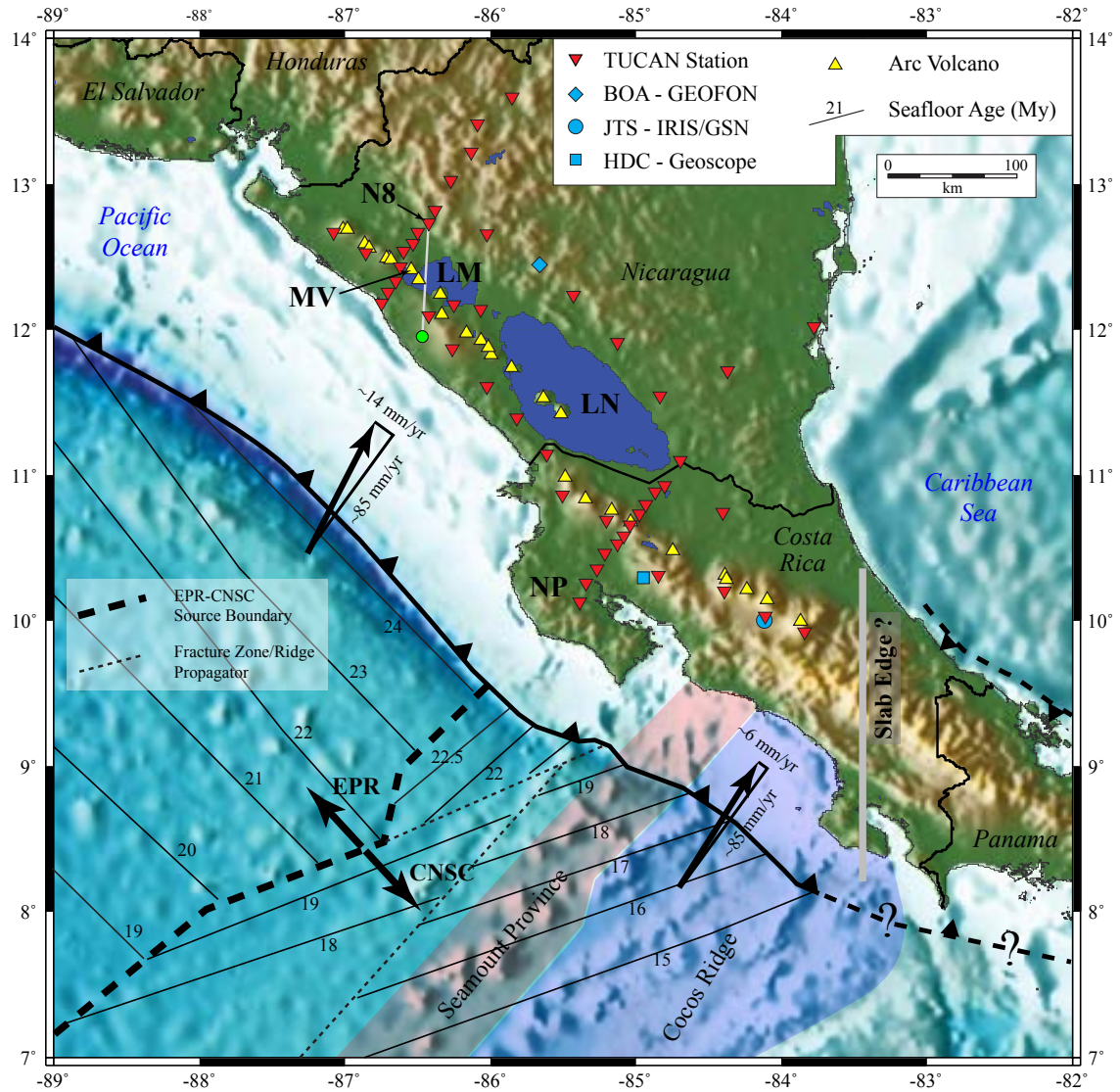


Figure 1b. The enlarged area here (see Figure 1a on previous page) shows the seismic stations used in the splitting analysis, as well as some of the important tectonic features of the region. These include the difference in Cocos Plate age and source ridge across the trace of the triple junction (thick dashed black line) [Barckhausen *et al.*, 2001], the termination of deep, subduction related seismicity in southern Costa Rica [Protti *et al.*, 1994], an apparent bend in the slab beneath northern Costa Rica [Protti *et al.*, 1994; Syracuse *et al.*, 2008], and the change in obliquity of subduction along strike of the trench [DeMets, 2001]. Each of these features may kinematically influence the pattern of flow in the mantle wedge and, therefore, must be considered when interpreting flow from anisotropy. The subducting seamount province and Cocos Ridge, discussed in Section 2 and in Hoernle *et al.* [2008], are highlighted in red and blue, respectively. To maintain legibility, we label only station N8 in the Nicaraguan cross-arc line for reference to the splitting example in Figure 2 (green circle is event location and grey line is ray path for the example). LM – Lake Managua; LN – Lake Nicaragua; MV – Momotombo Volcano; NP – Nicoya Peninsula.

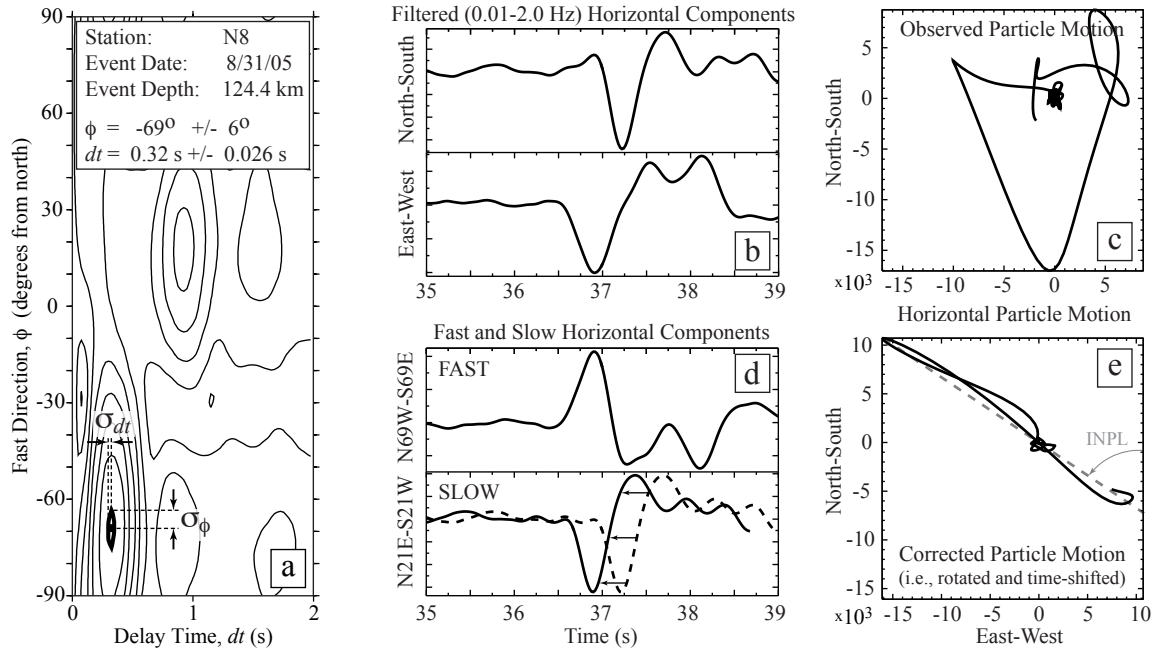


Figure 2. Example splitting measurement at TUCAN station N8 (See Figure 1b). **(a)** Contoured values of the smaller eigenvalue in the cross-correlation matrix (λ_2) around the best-fitting pair of splitting parameters (black dot, $\lambda_{2\min}$) – see Section 3.1. The thick black ellipse encloses the 95% confidence region, and the maximum distances of this contour from the best-fitting ϕ and dt values represent the respective uncertainties for these parameters (σ_ϕ and σ_{dt}). **(b)** Observed N and E components after applying a band-pass filter of 0.01-2.0 Hz. The offset of the phase on the two components is already visible. **(c)** Characteristic elliptical particle motion of the two horizontal components. **(d)** The solid upper and dashed lower waveforms are the result after rotating both components by ϕ (69° west of north), and then by time-shifting the dashed wave by dt (0.32 s), the corrected particle motion in **(e)** is obtained. The angle, INPL, of approximately linear particle motion is used as the initial polarization angle in the forward modeling discussed in Section 4. Note that although we show only a 4 second window in **(b)** and **(c)**, the actual window used in the calculation begins earlier.

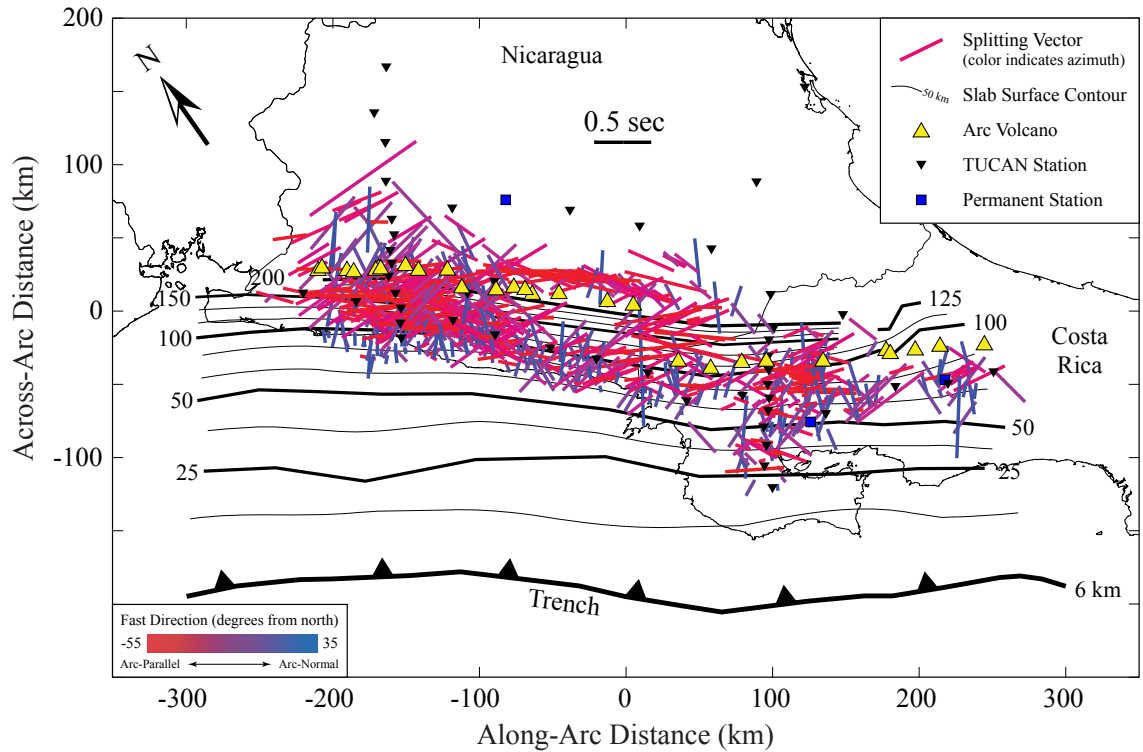


Figure 3. Shear-wave splitting observations (colored vectors) from local events recorded at stations from the temporary TUCAN Broadband Seismic Experiment (inverted black triangles) and three permanent stations (blue squares). All 791 local-S splits are plotted here at ray path midpoints, and because each is plotted at a different depth, many are obscured by overlying vectors. Length is scaled to splitting time (dt), and a reference vector for 0.5 seconds is shown (horizontal black vector in central Nicaragua). Vector orientation is parallel to fast direction (ϕ), with color grading between red (arc-parallel, N55W) and blue (arc-normal, N35E). This color scheme is used to help distinguish patterns so that regions showing significant coherence will stand out relative to those with many different orientations. For example, arc-parallel fast directions dominate along much of the Nicaraguan arc, as well as in the central Nicaraguan back-arc, and this can be seen by the large proportion of red splitting vectors. Also note the relatively arc-normal orientation of most vectors beneath the Nicaraguan coast, where waves are predominantly sampling the wedge corner. However, there seem to be anomalies throughout the entire region, making it hard to interpret these measurements individually - thus the usefulness of a tomographic inversion becomes apparent. The coordinate system is rotated 35° east of north such that the $\sim$ strike of the arc/trench is the x_1 direction and x_2 is arc-normal. Distances (used in the inversion) are in kilometers from $(-85.5^\circ, 11.5^\circ)$, a rough approximation of the center of the array.

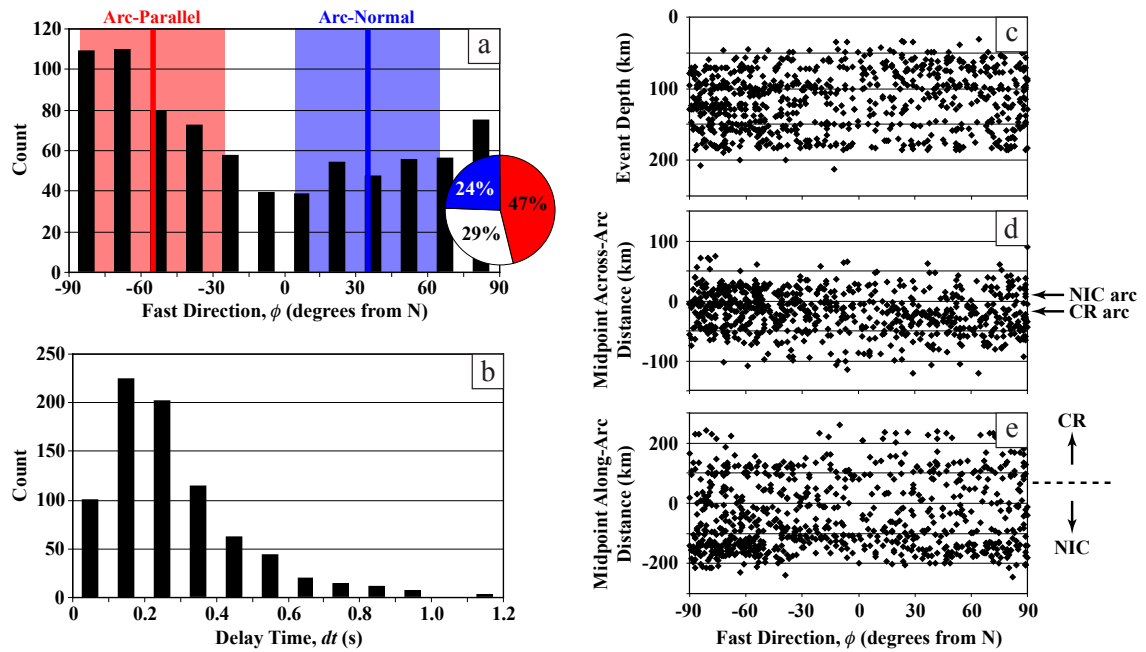


Figure 4. Shear-wave splitting parameter statistics. **(a)** Histogram of ϕ in bins of 15° . The red and blue boxes denote a range of fast directions $\pm 30^\circ$ from arc-parallel and arc-normal, respectively. There is a predominance of approximately arc-parallel fast directions, but this data set is not bimodal (Note pie chart, with blue=arc-normal, red=arc-parallel, white=intermediate), as is seen in Japan [Nakajima *et al.*, 2006] and would be predicted with 2-D corner flow and B-type fabric in the wedge corner [e.g., Kneller *et al.*, 2005]. **(b)** Histogram of dt in bins of 0.1 s. **(c)** Fast direction vs. event depth. **(d)** Fast direction plotted against the across-arc distance of the midpoint of each ray. **(e)** Fast direction vs. along-arc distance of each midpoint. (c)-(e) demonstrate no clear correlation exists for the entire dataset between location and fast direction.

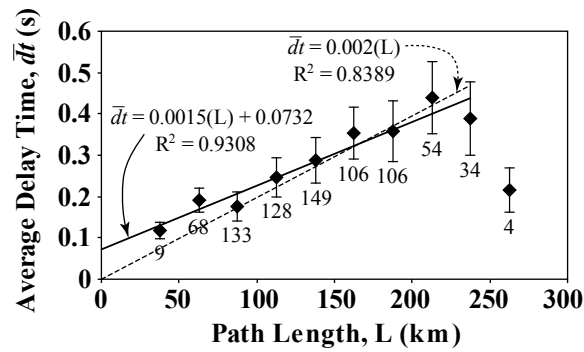


Figure 5. Average delay time for 25 km bins of straight-line path length. Error bars show the average 1σ for the average dt in each bin, and the numbers below each data point indicate the number of splits in each bin. The very coherent trend of increasing birefringence with increasing path length indicates that the mantle wedge is the dominant source of anisotropy. The solid line shows the best-fit linear trend of the data (excluding the 4 longest paths), while the dashed line gives the trend assuming that delay time goes to zero as path length goes to zero. The disparity between these two trends may indicate the presence of anisotropy in the overriding lithosphere or crust that is strong relative to that in the mantle wedge, but which occurs over such a small fraction of total path lengths that the mantle wedge anisotropy dominates.

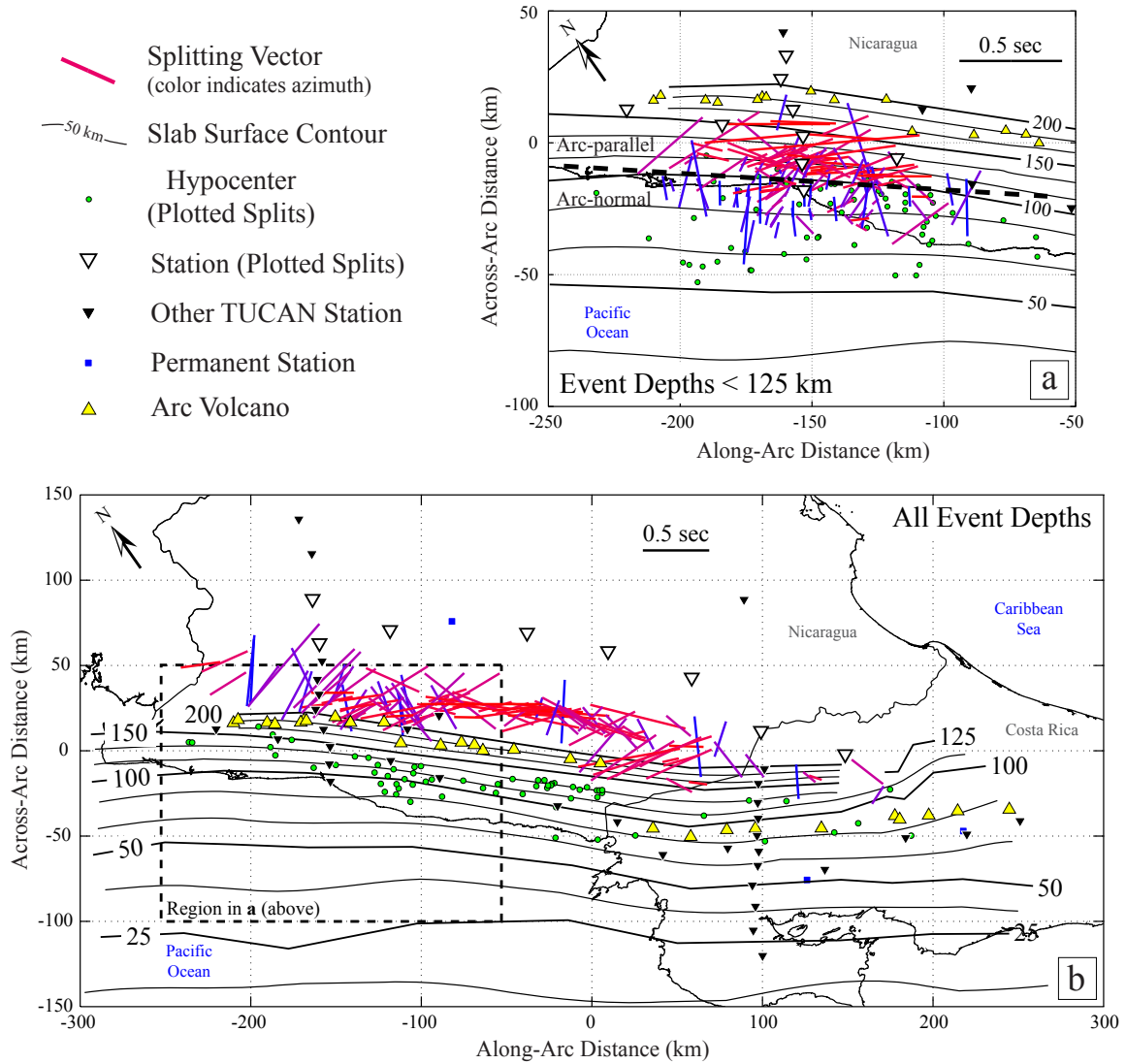


Figure 6. Subsets of the shear-wave splitting measurements show greater coherence than the entire dataset (Fig. 3). The larger inverted empty black triangles represent the stations at which the splits shown were measured. **(a)** In NW Nicaragua, for splits from shallow events (<125 km), a relatively sharp transition (thick dashed line) occurs between arc-normal fast directions closer to the trench and arc-parallel fast directions towards the arc. This pattern is predicted for along-arc flow and B-type fabric in the cold wedge corner [Jung and Karato, 2001; Kneller *et al.*, 2005], but is the inverse of what would be expected for 2-D corner flow (i.e., arc-parallel fast directions close to the trench and changing to arc-normal in the arc and back-arc). **(b)** Across the central Nicaraguan back-arc region, fast directions are strongly arc-parallel, while more oblique fast directions are observed to the NW and SE.

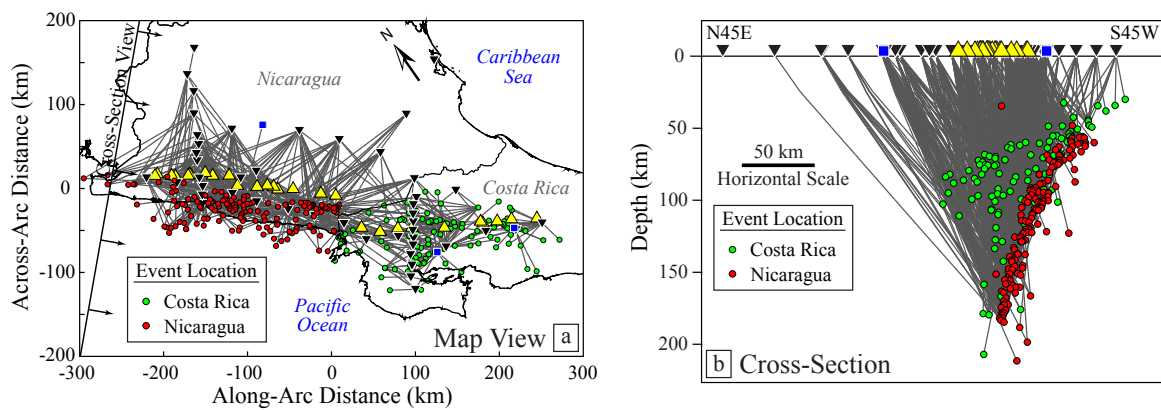


Figure 7. Event distribution and ray coverage in map view (a) and cross-section (b) for the shear-wave splitting data-set used in the inversion. Note that although the model is parameterized as blocks in an along-arc (N55W) and across-arc (N35E) direction, the cross-section here (b) is shown oblique to this orientation (N45E) because the strike of the slab changes slightly along the arc. Yellow triangles are volcanoes, black inverted triangles are TUCAN stations, and blue squares are permanent stations.

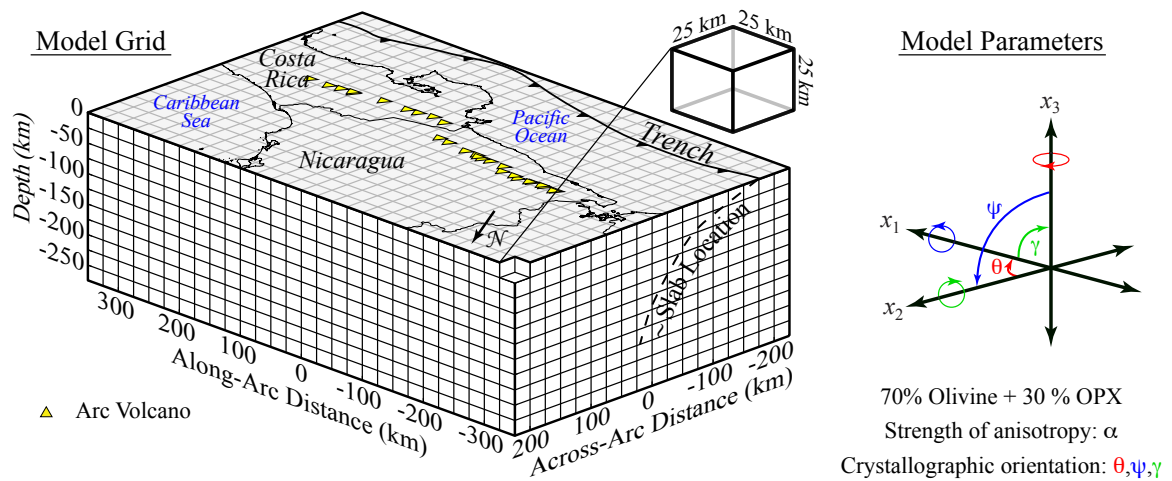


Figure 8. Model discretization and parameterization. The elastic properties of olivine [Anderson and Isaak, 1995; Abramson *et al.*, 1997] and orthopyroxene [Frisillo and Barsch, 1972] are assigned to each model block (left), with their orientation defined by rotation angles (right). For the hexagonal symmetry approximation, only two angles (θ and ψ) are needed. The strength of fabric alignment is determined by the scalar parameter α . The grid of model blocks is shown relative to the volcanic arc (yellow triangles), Nicaragua-Costa Rica coastline, trench, and approximate location of the subducting Cocos Plate at depth beneath north-west Nicaragua.

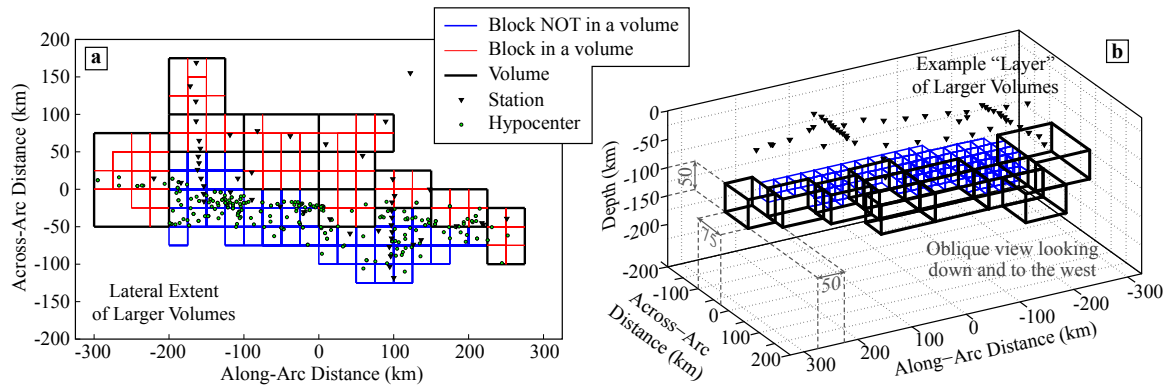


Figure 9. Larger volumes generated to increase resolution of individual parameters and expand the interpretable region of the model space. Blue blocks are those that remain independent after volumes (thick black lines) are formed. (a) Lateral extent of volumes. The red blocks are those within the volume that are actually touched by a ray and used in the inversion. (b) Example “layer” of larger volumes and independent blocks. The lateral extent of each larger volume does not change with depth, and each layer is 50 km thick, except for the surface layer, which is only 25 km. For clarity, the red blocks within the volumes are not shown here.

Inversion Results

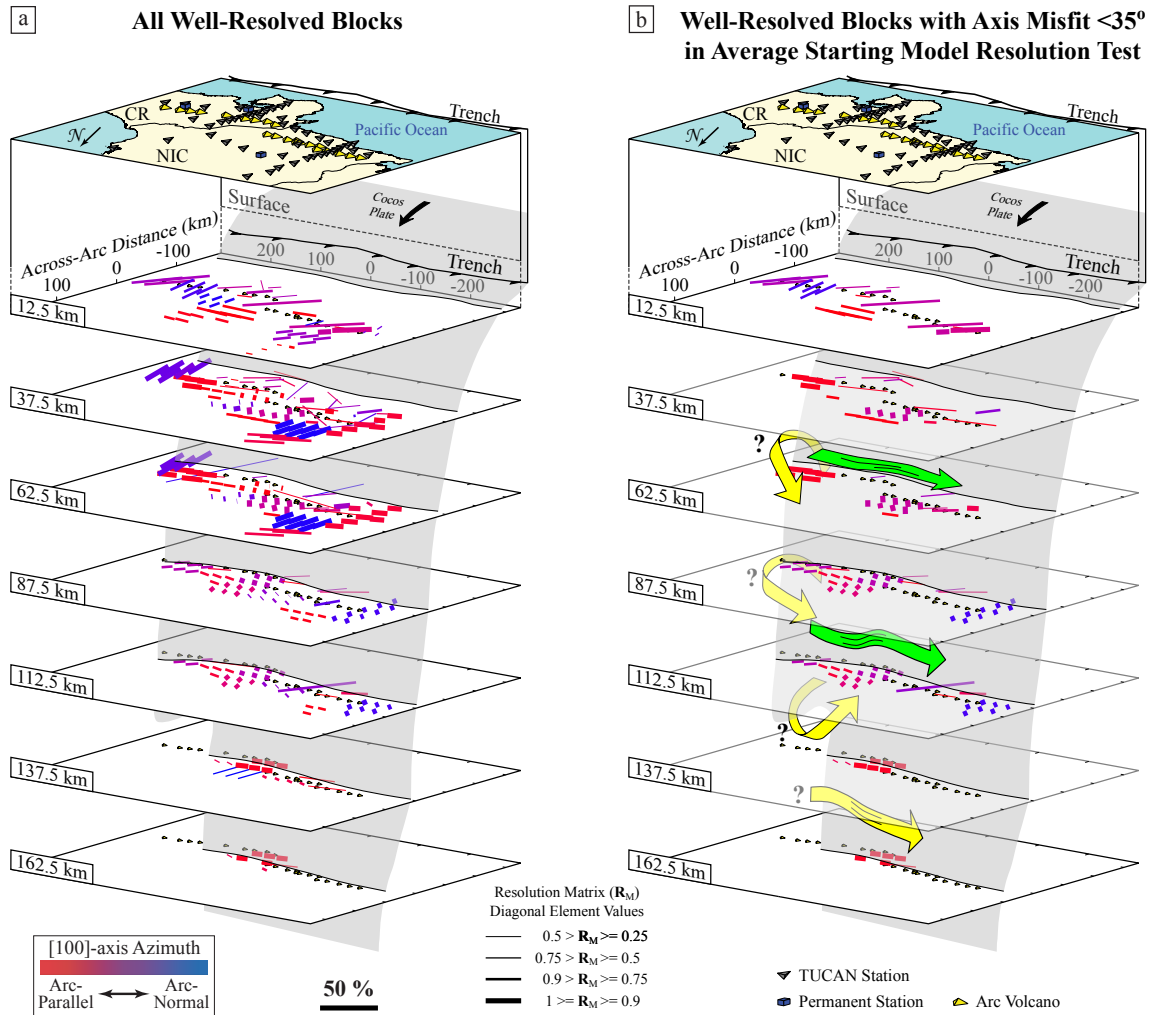


Figure 10. Model of anisotropy in the mantle wedge beneath Costa Rica and Nicaragua. Vectors represent a -axes of olivine, and although dip is solved for, the axes have been projected onto horizontal planes to emphasize their azimuthal orientation. Color represents azimuth, and varies from arc-parallel (red) to arc-normal (blue). The thickness of model vectors (here and in Figs. 11, 13, and 14a) is scaled by the resolution matrix diagonal element values; qualitatively, thicker vectors are better resolved and thinner vectors are less well-resolved. The intersection of the Cocos Plate (grey surface) with each cross-section is shown with a black line (i.e., slab contours), and the volcanic arc is shown for reference on each cross-section as well (small yellow triangles). **(a)** Model blocks deemed “well-resolved” from the resolution matrix, with the thickness of each vector corresponding to the associated resolution matrix diagonal element value (e.g., thinnest lines ≥ 0.25 , thickest lines ≥ 0.90). **(b)** Model blocks from (a) that also have a -axis angular misfits $\leq 35^\circ$ between the input and retrieved structure in the recovery test using the average starting model (Fig. 14b). These remaining vectors are those, which we most confidently interpret. Owing to their location in the mantle wedge (see Section 5.2), we interpret the roughly arc-parallel orientation of most a -axes to indicate the presence of arc-parallel directed shear (i.e., flow) in the mantle wedge. Yellow arrows around the edge of the slab indicated possible flow paths, while horizontal green arrows show the dominant orientation of flow suggested by both shear-wave splitting tomography and geochemical trends (Section 5.3). Our current data and inversion do not distinguish between flow with a horizontal shear plane and that with a vertical shear plane (i.e., that observed in the exposed Talkeetna arc in Alaska [Mehl *et al.*, 2003]), and therefore we illustrate both as possibilities.

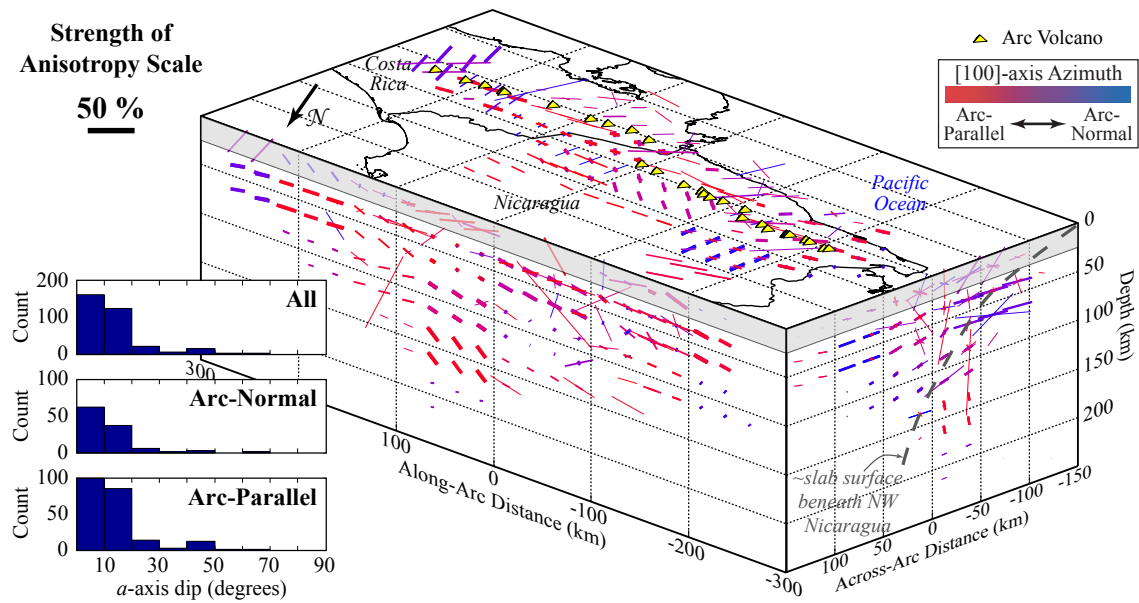


Figure 11. Pseudo 3-D view of the model of anisotropy shown in Figure 10. All well-resolved model a -axes (as determined by the resolution matrix) are projected onto an along-arc, an across-arc, and a horizontal plane. The dip of arc-parallel (arc-normal) a -axes can be seen in the along-arc (across-arc) cross-section. The most prominent deviations in a -axes from horizontal are found in the fore-arc wedge beneath Nicaragua between -50 and -200 km in the along-arc direction. However, histograms of a -axis dip are shown to the left, and nearly all well-resolved a -axes have dips less than 20° . The grey masking of the top layer is meant as a reminder that the mineralogy, and thus anisotropy, at crustal depths is likely not directly interpretable.

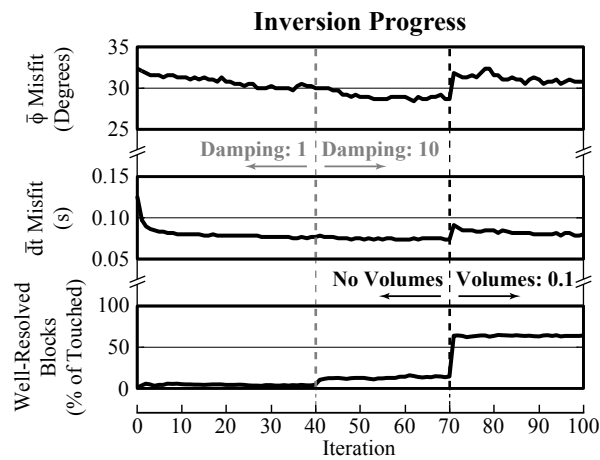


Figure 12. Progression of data misfit during the inversion. Average fast direction misfit and average delay time misfit are weighted by errors in both the splitting observations and the calculation of synthetic splitting parameters. The relaxation in damping at iteration 40 does not result in any observable instability in the model (i.e., a dramatic change in the model leading to a commensurate increase or decrease in data misfit). The resolution matrix takes into account the level of damping, and therefore, increasing *a priori* variance (i.e., reducing the penalty function constraints) leads to a slight increase in the number of “well-resolved” blocks. When blocks are grouped into volumes at iteration 70, the model parameters in each block within a volume are averaged, resulting in a spike in misfit. However, acting as one larger block allows for a greater number of ray paths to constrain the best-fitting LPO, and therefore a much larger portion of the model space becomes interpretable.

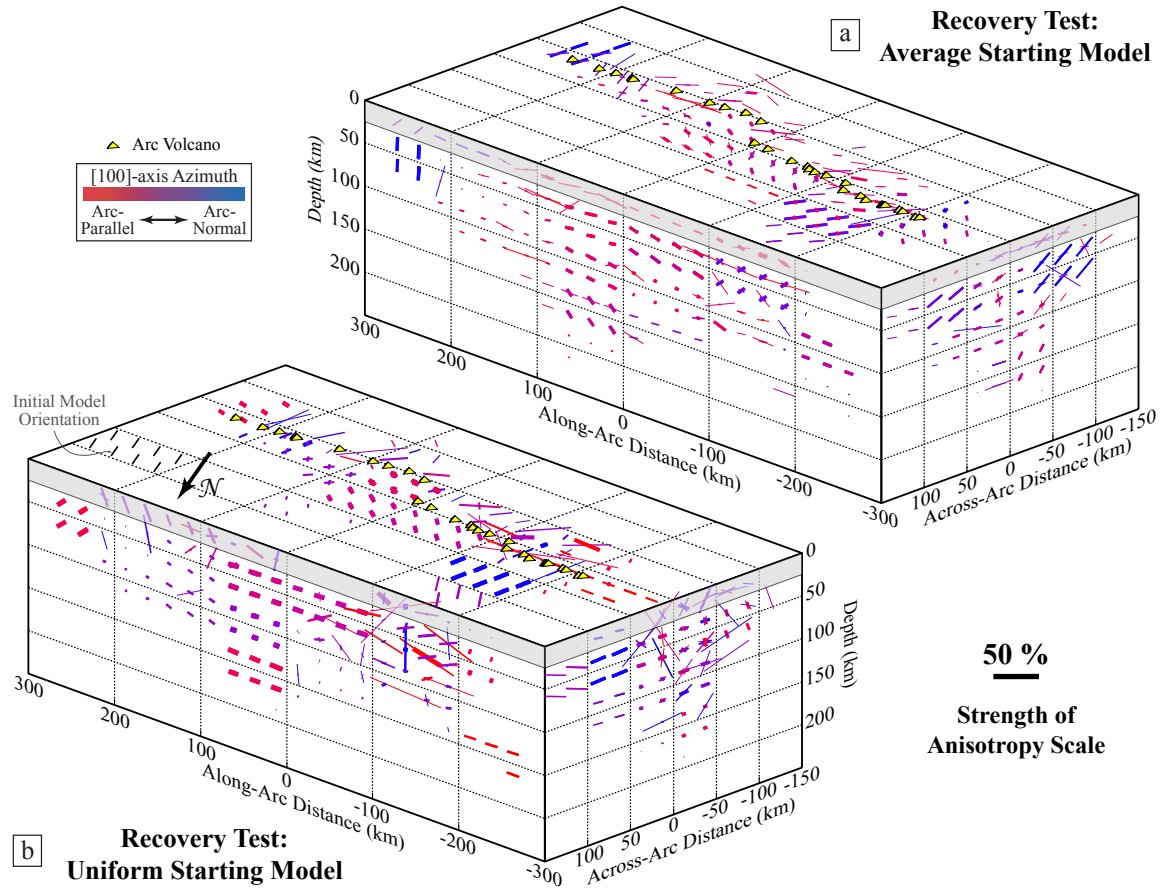


Figure 13. Recovery test results. All blocks shown are those considered well-resolved by the resolution matrix from the final iteration of the inversion. **(a)** Final model from the recovery test using the average starting model. **(b)** Final model from the recovery test using the uniform starting model. The uniform starting model contained horizontal a -axes oriented N-S with a strength of 25% (see eastern corner of the map view cross-section). Note that because the resolution matrix is model dependent, some model blocks that are considered well-resolved in the real inversion (Fig. 11) are not well-resolved in the recovery test and, thus, vectors plotted in (a) and (b) may not be plotted in Figure 11, and vice versa. Despite the dramatic difference in starting model, the large scale structure in both recovery tests shows strong agreement with the input structure (i.e., predominantly arc-parallel a -axes, with pockets of roughly arc-normal a -axes beneath northwest Nicaragua and central Costa Rica).

Average Starting Model Recovery Test Results

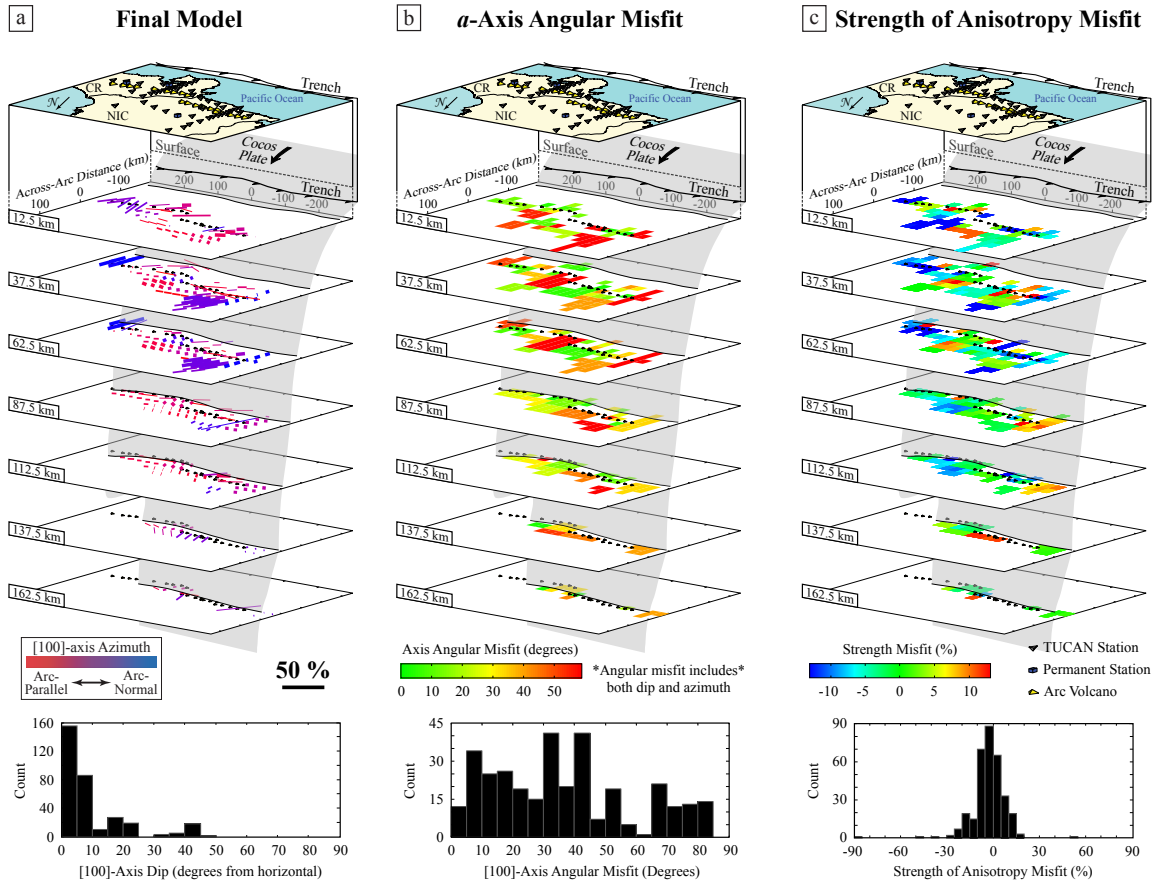


Figure 14. Recovery test results using an average starting model. (a) Same model as in Fig. 13b, except with *a*-axes projected onto horizontal cross-sections (for comparison with Figure 10a). As in Figure 13, only *a*-axes in well-resolved model blocks are shown. (b) Angular misfit between *a*-axes in (a) and those in Figure 10a. Angular misfit includes both dip and azimuth of the *a*-axis. Azimuthal misfits are only slightly lower than angular misfits given the relatively horizontal orientation of most *a*-axes (histogram below (a) and Fig. 12). The color scale has been saturated at one standard deviation from the mean in order to allow for variation at smaller misfits ($<35^\circ$) to be distinguished; as shown by the histogram of axis misfits, this saturation affects ~ 50 blocks. (c) Strength of anisotropy misfit (negative values indicate a lower recovered strength). The color scale is again saturated at one standard deviation from the mean, but here the distribution is much narrower and fewer blocks fall outside this value.

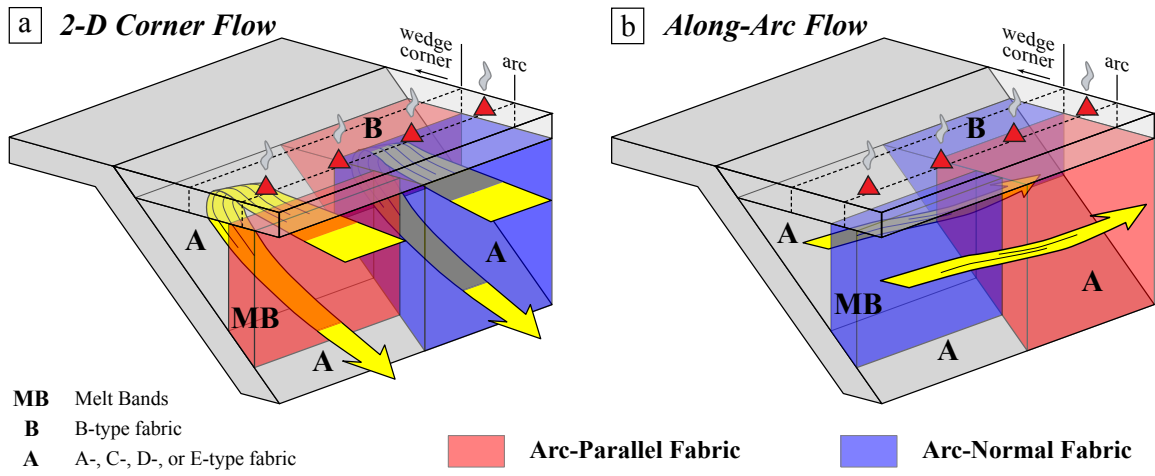


Figure 15. Interpreting the orientation of fast S-wave polarization directions in terms of fabric generated by dislocation creep in olivine during simple shear in the upper mantle. Numerous studies of naturally and experimentally deformed polycrystalline peridotite (and olivine) show that the easiest (i.e., dominant) slip system in olivine will change as stress and water content (H/Si) increase. Despite the various combinations of crystallographic axis orientations relative to the shear plane and shear plane-normal, only the B-type fabric will result in the fast shear-wave polarization direction (for a vertically propagating wave) being normal to the shear direction. Although the dominant slip system observed by *Holtzman et al.* [2003] is A-type, (010)[100], in the solid crystal lenses between melt bands, the fast seismic direction from both LPO and the melt bands themselves is normal to the shear direction. **(a)** Predicted fabric (i.e., fast shear-wave polarization direction) orientation for 2-D corner flow in the presence of melt bands (MB, left) or B-type fabric (B, right). **(b)** Predicted fabric generated by predominantly along-arc flow in the presence of melt bands (MB, left) or B-type fabric (B, right). Note the location of expected deviations from A-type fabric (A) in the wedge relative to the arc and trench (dashed lines). Although predicted splitting parameters (especially ϕ) will vary somewhat across these boundaries, the pattern should mimic the underlying mantle fabric transitions [*Abt and Fischer, 2008*].

This table is too large to be reproduce in print here, but can be found online
with the published version of this chapter [*Abt et al.*, 2009, G³]:

<http://www.agu.org/journals/gc/gc0905/2009GC002375/supplement.shtml>

Table S1. High quality splitting measurements and event and station information. Each event is identified by year, Julian day, and hour (e.g., 2004.237.07). Earthquake origin times and hypocenter locations (after relocation), station locations, and straight-line ray path midpoints are given for each splitting measurement. Locations are given in both a geographic coordinate system (degrees) and in the rotated coordinate system used in the parameterization of the model space and the inversion (kilometers from 11.5N, -85.5W). Delay time (dt) is in seconds and both fast direction (ϕ) and initial polarization (INPL) are in degrees east of north. Errors (σ) are also included; σ_ϕ and σ_{dt} represent the maximum uncertainty in ϕ and dt , respectively (i.e., the maximum distance from the best-fitting pair of splitting parameters to the limits of the 95% confidence region – see Fig. 2a). Initial polarization (INPL) is measured from the corrected particle motion (Fig. 2e). A comma delimited version of this table is also provided.

Chapter 3

Constraints on Upper Mantle Anisotropy Beneath Costa Rica and Nicaragua From SK(K)S Splitting

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Abstract

SKS and SKKS splitting observations are used to constrain the pattern of mantle flow in the Central American subduction zone beneath Costa Rica and Nicaragua. After removing the effects of shallow wedge anisotropy on SK(K)S waveforms, a best-fitting model of anisotropy beneath the Cocos Plate and in the deeper mantle wedge is determined. Fast polarization directions and model symmetry axis orientations in both regions (as well as the shallow wedge) are dominated by roughly arc-parallel azimuths and, therefore, are not consistent with sub-lithospheric mantle flow entrained by the subducting Cocos Plate or simple two-dimensional corner flow in the wedge. In conjunction with geochemical data and local-S splitting tomography, we interpret the SK(K)S splitting observations and models to indicate flow to the WNW within the mantle wedge on the Caribbean side of the Cocos Plate, possibly drawn through a slab window beneath Panama and southern Costa Rica. Flow of Pacific mantle beneath the Cocos Plate is also roughly parallel to the strike of the slab, although its absolute direction is not uniquely constrained.

1. Introduction

The preferential alignment of olivine and orthopyroxene (opx) during deformation is often considered to be the dominant source of seismic anisotropy observed in the upper mantle, and shear-wave splitting is one of the most widely used tools for investigating anisotropy in the mantle beneath subduction zones [e.g., *Russo and Silver, 1994; Fouch and Fischer, 1996; Smith et al., 2001; Anderson et al., 2004; Levin et al., 2004; Long and van der Hilst, 2005; Pozgay et al., 2007; Greve et al., 2008; Long and Silver, 2008*]. By assuming a relationship between the direction of deformation, or flow, and the lattice-preferred orientation (LPO) of these minerals, shear-wave splitting parameters (i.e., fast polarization direction, ϕ , and delay time, dt) can be used to infer the pattern of flow in the upper mantle [e.g., *Silver and Chan, 1991; Mainprice and Silver, 1993; Savage, 1999; Long and Silver, 2008*].

Outside of the mantle wedge region in subduction zones, pressures and temperatures in the sub-lithospheric upper mantle are likely to be such that dislocation creep in olivine is the dominant deformation mechanism, and under these conditions the [100]-axis (or a -axis) of olivine is expected to align roughly parallel to the direction of flow during simple shear (e.g., rigid lithosphere driving deformation in the underlying asthenosphere with A-type slip [*Zhang and Karato, 1995; Ismail and Mainprice, 1998; Mainprice et al., 2000*]). Given these assumptions, the fast shear-wave polarization direction in core phases (e.g., PK(K)S, SK(K)S, etc.) with near vertical incidence (5° - 15°) is expected to be parallel or sub-parallel to the mantle flow direction. In contrast, within the mantle wedge, the presence of additional water [*Jung and Karato, 2001; Jung et al., 2006; Katayama and*

Karato, 2006; Jacobsen et al., 2008] and melt [*Holtzman et al., 2003*] and the likelihood of three-dimensional flow patterns [*Kincaid and Griffiths, 2004; Funiciello et al., 2006; Kneller and van Keken, 2008*] complicate the interpretation of shear-wave splitting.

Anisotropy within the mantle wedge beneath Costa Rica and Nicaragua has previously been investigated from shear-wave splitting in local-S waves [*Abt et al., 2009*]. The highly variable local-S fast directions (Fig. 1) made a direct interpretation in terms of flow difficult, and they were instead utilized in an inversion for anisotropic structure in the mantle wedge [*Abt and Fischer, 2008; Abt et al., 2009*]. The resulting model displayed a predominance of arc-parallel olivine *a*-axes throughout the mantle wedge and not only in the extreme wedge corner or beneath the volcanic arc where B-type fabric [*Jung and Karato, 2001; Kneller et al., 2005*] and melt bands [*Holtzman et al., 2003*], respectively, might affect fabric development and lead to a different relationship between flow direction and olivine *a*-axis orientation. Combined with Pb and Nd isotope data, the best interpretation of the local-S splitting and anisotropy model is that roughly horizontal along-arc flow to the northwest exists in the shallow (<200 km) mantle wedge [*Hoernle et al., 2008; Abt et al., 2009*].

Along-arc flow in the mantle wedge is not consistent with the standard two-dimensional (2-D) corner flow expected to be present here in Central America given the relatively planar shape (5°-15° change in dip over ~400 km along-strike [*Protti et al., 1994; Husen et al., 2003; Syracuse and Abers, 2006*]) and nearly orthogonal convergence (7° in central Costa Rica, 17° in Nicaragua [*Barckhausen et al., 2001; DeMets, 2001*]) of the Cocos Plate beneath the Caribbean Plate. A combination of trench motion, slab edge effects, and possible slab-wedge decoupling was suggested as a means of generating this

arc-parallel flow [Hoernle *et al.*, 2008; Abt *et al.*, 2009]. However, due to the limited depth of local events (<200 km) and lateral extent of the seismic array used (Fig. 1), a first-order question left unanswered by the local-S splitting study is whether or not this counter-intuitive pattern of arc-parallel anisotropy exists beyond the shallow mantle wedge. If it is confined to the shallow wedge and the mantle at greater depth possesses a more arc-normal orientation of anisotropy (i.e., that predicted for 2-D corner flow), then it is likely either that the deeper wedge becomes recoupled to the downgoing slab or that the source of shallow arc-parallel anisotropy is not simply mantle flow (e.g., B-type fabric, aligned melt, crustal anisotropy, or some combination of the these). On the other hand, pervasive arc-parallel anisotropy throughout the upper mantle would support models in which other drivers of flow (e.g., trench roll-back near a slab edge [Buttles and Olsen, 1998; Kincaid and Griffiths, 2003,2004; Funicello *et al.*, 2006] or global convection patterns [e.g., Conrad *et al.*, 2007]) are locally more influential than down-dip motion of the subducting Cocos Plate.

Here we present SKS and SKKS splitting measurements from Costa Rica, Nicaragua and Honduras that complement the local-S data and help constrain the orientation of flow both deeper in the back-arc wedge and beneath the Cocos Plate. In addition, after taking into account the local-S model of anisotropy [Abt *et al.*, 2009], we calculate a best-fitting crystallographic orientation for both the sub-slab and deeper wedge and find that each region likely contains anisotropy characterized by shallowly dipping, roughly arc-parallel olivine *a*-axes (from here on, referred to as the symmetry axis, given that we will assume hexagonal crystallographic symmetry). This result suggests arc-parallel flow is indeed

the most probable source for anisotropy in the shallow wedge imaged by local-S splitting tomography, as well as in the deeper wedge and sub-slab mantle.

2. SK(K)S Splitting

2.1. Calculation of Splitting Parameters

Shear-wave splitting parameters (ϕ, dt) are calculated here from the SV and SH components of motion using an eigenvalue minimization technique [*Silver and Chan, 1991*]. A zero-phase, bandpass filter of 0.05-0.2 Hz (periods of 5-20 sec) is first applied to the raw waveforms, the Z-N-E components are then rotated into Z-R-T, and finally the waveforms are transformed into the P-SV-SH coordinate system by correcting for the effects of the free surface [*Kennett, 1991*] assuming a surface V_P and V_S of 5.9 km/s and 3.41 km/s, respectively. Ray paths are traced through the 1-D reference Earth model AK135 [*Kennett et al., 1995*]. While splitting measurements are typically made on the horizontal components of motion (i.e., N-E or R-T [e.g., *Silver and Chan, 1991; Fouch and Fischer, 1996; Savage, 1999*]), with fast polarization direction measured in degrees east of north, fast directions here are measured clockwise from SV in the SV-SH plane (Fig. 2). Although the difference between the splitting parameters measured in these two coordinate systems is generally minimal (Fig. 3) given the sub-vertical incidence angle of most SK(K)S waves, we choose to use the splitting measurements made from the SV-SH components and not those from the projected N-E, or R-T, components. The window used for analysis is manually chosen and typically includes 20-30 seconds prior to and at

least 1.5-2 periods after the phase arrival. Error bars ($s_{f,dt}$) are the standard splitting parameter errors given by the maximum distance from the best-fitting ϕ and dt to the 95% confidence contour of the l_2 surface (e.g., Fig. 3c,f; also see *Abt and Fischer* [2008]).

2.2. Observations

Our data were recorded by the Tomography Under Costa Rica and Nicaragua (TUCAN) seismic array, a temporary network of 48 broadband seismic stations deployed from July 2004 to March 2006, as well as four permanent stations: BOA (INETER and GEOFON, Program of GFZ Potsdam), HDC (OVSICORI and GEOSCOPE, Institut de Physique du Globe), JTS (OVSICORI and IRIS - Scripps Institute of Oceanography), and TGUH (Caribbean Network, USGS). More than 1900 waveforms from events with magnitude ≥ 6.5 and epicentral distance between 85° and 170° were visually inspected for clear SKS and SKKS arrivals, and of these, more than 900 were analyzed with the method described above in Section 2.1, resulting in 103 “high quality” splitting measurements (Fig. 1, Table 1). We use four basic criteria for assessing the quality of splitting measurements (as in *Abt et al.* [2009]); a “high quality” (i.e., interpretable) measurement will (1) possess a clear phase arrival with a high signal-to-noise ratio, (2) display elliptical or circular particle motion, (3) have a single, elliptical 95% contour, and (4) return roughly linear particle motion with dominantly SV energy when the waveforms are corrected for the best-fitting pair of splitting parameters. Although we do not use strict limits on the size of the 95% confidence region, only 3 of the 103 splits presented here have a fast direction error of $>30^\circ$ and 19 have a delay time error >0.6 sec (Table 1).

In contrast to the relatively complicated pattern of local-S fast directions [Abt *et al.*, 2009], the teleseismic SK(K)S fast directions observed here are more consistently arc-parallel and sub-arc-parallel (Figs. 1,4,5). SK(K)S delay times (1-3 sec) are also generally much larger than those of local-S (0.1-0.6 sec), indicating a significant amount of anisotropy outside the shallow wedge and upper plate volumes sampled by local-S waves. The azimuthal distribution of events from the splitting measurements is neither uniform nor complete (Fig 5). In addition, one set of ray paths primarily sample the mantle beneath the subducting Cocos Plate (i.e., Pacific) and another set primarily samples the wedge-side (i.e., Caribbean) mantle (Figs. 5,6; Table 2); note that here and in the modeling (Section 3) we do not include the ray paths and splits to station TGUH because they fall further outside the local-S model space. We find a relatively small difference in splitting parameters (on average, 5° in ϕ and 0.30 sec, or 15%, in dt) between Caribbean and Pacific rays (Fig. 7); the mean ϕ and dt of the 66 (27) splits from the Caribbean (Pacific) side are -69° (-74°) and 1.95 sec (1.65 sec).

Qualitatively, the consistency in SK(K)S fast directions suggests a similar orientation of anisotropy in both Pacific mantle beneath the Cocos Plate and Caribbean mantle on the wedge side of the Cocos Plate (e.g., $\sim$ arc-parallel symmetry axes). The slight difference in average delay time between waves sampling the two mantle regions could be attributed to differing path lengths through a shallow mantle wedge that possesses a higher degree of anisotropy than the sub-slab or deeper (>200 km) wedge mantle; rays on the Caribbean side generally have a greater fraction of their path within the shallow wedge (Fig. 6). On the other hand, the difference in dt could indicate a difference in the strength and/or orientation of anisotropy outside the mantle wedge between the two regions. For

example, the strength of anisotropy for sub-vertical (10° incidence angle) ray propagation decreases from $\sim 8.5\%$ to $\sim 1\%$ as the symmetry axis rotates from horizontal to vertical (Suppl. Fig. S1). To more quantitatively characterize anisotropy, in the following section, we use the local-S splitting tomography model to remove the effects of anisotropy in the shallow wedge on SK(K)S waveforms and calculate best-fitting models of anisotropy in these two mantle regions.

3. Sub-slab and Deep Back-arc Anisotropy

3.1. Incorporation of Local-S Splitting Tomography Model

Given the model of anisotropy (i.e., crystallographic orientation and strength of anisotropy) in the shallow (<200 km) mantle wedge determined from local-S splitting tomography [Abt *et al.*, 2009] we can correct for the predicted birefringence that would be accrued by SK(K)S phases during passage through this region. To accomplish this, the observed SK(K)S waveforms from all stations (except TGUH, which lies outside the local-S model region) were progressively rotated and time-shifted according to the polarization directions (eigenvectors) and phase velocities (eigenvalues) from the Christoffel matrix (m_{ik}) for each well-resolved block touched in the local-S model

$$(1) \quad m_{ik} = \frac{1}{\rho(z)} \cdot c_{ijkl}(z) \cdot \hat{n}_j \hat{n}_l$$

where ρ is density and a function of depth z (i.e., pressure and temperature), c_{ijkl} is the elastic stiffness tensor of an olivine-opx crystal (also a function of P and T ; Suppl. Table S1), and $\hat{n}$ is ray propagation direction [e.g., Babuska and Cara, 1991]. This method is

nearly identical to the forward calculation of predicted splitting utilized in the local-S inversion [Abt and Fischer, 2008; Abt *et al.*, 2009] except that here we (1) apply the time-shift in the reverse order (i.e., the fast and slow components are shifted backward and forward in time, respectively) and (2) are working with real waveforms and not synthetic wavelets. Abt *et al.* [2009] contains a complete description of the wedge anisotropy model, including block dimensions and model resolution.

Splitting parameters calculated from the resulting “backed-out” waveforms (e.g., Fig. 8) provide us with a means of constraining the orientation and strength of anisotropy in the mantle outside the portion of the wedge sampled by local-S waves (i.e., further and deeper into the back-arc wedge and beneath the subducting slab). An assumption of the maximum depth of anisotropy, somewhere between the core-mantle boundary and the surface, must be made, and here we choose this depth to be the top of the transition zone at 410 km. Comparisons of shear-wave splitting between deep-focus earthquakes and SKS phases show that in general almost no splitting occurs on sub-vertical paths in the transition zone and lower mantle, although transition zone splitting does appear to occur in isolated regions [Fischer and Wiens, 1996; Fouch and Fischer, 1996]. Decreasing (increasing) the anisotropy cut-off depth in our models would simply increase (decrease) the strength of anisotropy inferred from the corrected SK(K)S splitting measurements. We assume the same elastic constants used in the local-S model (i.e., 70% olivine [Anderson and Isaak, 1995; Abramson *et al.*, 1997] and 30% opx [Frisillo and Barsch, 1972] with hexagonal symmetry and the olivine a -axis as the symmetry axis; Suppl. Table S1).

This approach assumes SK(K)S waves, which have dominant frequencies of 0.05-0.1 Hz (wavelengths of ~50-100 km), are affected by mantle wedge anisotropy in the same manner as local-S waves, with dominant frequencies of 0.5-1 Hz (wavelengths of ~5-10 km). If anisotropy in the shallow wedge, as sampled and imaged by local-S waves, is the result of structure (e.g., LPO, aligned melt) that varies with a length scale greater than local-S wavelengths but less than SK(K)S wavelengths, then SK(K)S waves would effectively sense an average of the finer-scale anisotropic structure in the local-S model. The local-S model was parameterized with 25^3 km³ blocks, and individual blocks in the back-arc regions of the model were combined into larger volumes (50 km vertically by up to 100 km laterally). Consequently, SK(K)S waves might be expected to smooth/average the anisotropy in regions of the local-S model beneath the arc and fore-arc. SK(K)S path lengths through the local-S model and subsequent delay times accrued by SK(K)S waves are generally small; on average, the delay times from the backed-out splits differ from those observed at the surface by only 0.12 sec (min=-0.62 sec, max=0.5 sec). Therefore, accounting for finite SK(K)S wavelengths in the local-S model would not likely dramatically change the resulting anisotropy models calculated here.

3.2. Best-Fitting Models of Crystallographic Orientation

The number of SK(K)S splitting measurements (93 to TUCAN stations) is almost an order of magnitude smaller than for local-S (791), and both the distribution of events for the SK(K)S measurements and their near-vertical incidence result in very few crossing ray paths (Fig. 6), unlike with local-S ray paths (see Fig. 7 in *Abt et al.* [2009]). Therefore, we do not attempt to tomographically invert for crystallographic orientation

using the backed-out SK(K)S waveforms. Instead, we simply conduct a grid-search over all possible symmetry axis orientations and strengths of anisotropy (Fig. 9). We search symmetry axis azimuths (q) from -180° to 180° at 10° increments, symmetry axis dips (ν) from 0° to 90° at 10° increments, and anisotropy strengths (a) from 0% to 100% at 10% increments.

For each SK(K)S ray path, a synthetic wavelet with random noise and a signal-to-noise ratio of 10 (typical for our observed SK(K)S waves) is set as the SV component of motion (SH is random noise) at 410 km and propagated up through the model (blocks 25 km on each side) until reaching the local-S model space (thick black line in Figure 6). Splitting of the resulting synthetic wavelet is calculated at this entry point in the local-S model in the same manner as with the real data and then compared with the observed backed-out splitting measurement from the corresponding ray. Because splitting is dependent on the relationship between crystallographic orientation and ray propagation direction (as well as initial polarization), some combinations of ray paths and model parameters will result in unstable or null measurements (characterized by fast direction errors of $>70^\circ$ and delay time errors of $>75\%$ of dt). When this occurs, we do not include the misfit from these measurements in the weighted misfit calculation. The weighted misfits for fast direction (e_f) and delay time (e_{dt}) from N splitting measurements are

$$(2) \quad \mathcal{E}_{\phi, dt} = \frac{\sum_{i=1}^N \left[(\phi, dt_i^{obs} - \phi, dt_i^{pred}) \cdot \left(\frac{\sigma_{\phi, dt_i}^{obs} + \sigma_{\phi, dt_i}^{pred}}{2} \right)^{-1} \right]}{\sum_{i=1}^N \left(\frac{\sigma_{\phi, dt_i}^{obs} + \sigma_{\phi, dt_i}^{pred}}{2} \right)^{-1}},$$

where s^{obs} and s^{pred} are the half-width of the 95% confidence region for the observed backed-out and synthetic splitting parameters, respectively, at the entry point in the local-

S model. The fast direction and splitting time misfits are then normalized by their maximum respective value, and the average of the two yields a combined measure of misfit for a particular set of model parameters (a, q, y) ,

$$(3) \quad \varepsilon^{norm}(\alpha, \theta, \psi) = \frac{|\varepsilon_{\phi} / \varepsilon_{\phi}^{max}| + |\varepsilon_{dt} / \varepsilon_{dt}^{max}|}{2}.$$

The resulting misfit volumes are presented as “slices” through the normalized misfit minimum (Fig. 9g) for both the Caribbean (Fig. 9a-c) and Pacific mantle (Fig. 9d-f), and the individual splitting parameter misfits from the best-fitting model for the backed-out waveforms at the entry point to the local-S model are shown in Figure 9h. If less than 50% of the ray paths from a region for a combination of model parameters result in well-constrained splits (i.e., non-null measurements), then we do not consider those model parameters when determining the model that minimizes misfit. This most often occurs for small strengths of anisotropy (e.g., small values of a and/or larger values of y) and for azimuths roughly parallel or perpendicular to the ray paths. The result of excluding these sets of model parameters is the appearance of “holes” in the misfit volume (gray regions in Fig. 9a-f).

As expected, given the coherent arc-parallel pattern of SK(K)S fast directions (Figs. 1,4,5,7), the best-fitting symmetry axis azimuth for both regions is roughly arc-parallel: $q = -70^\circ$ for the Caribbean mantle, and $q = -80^\circ$ for the Pacific mantle. The dip and strength of anisotropy also show only minor differences between the two regions: $y = 30^\circ$ and $a = 30\%$ for the Caribbean mantle, and $y = 10^\circ$ and $a = 20\%$ for the Pacific mantle. For this best-fitting Caribbean orientation, no measurements were excluded for unstable/null-like behavior, while only 2 were excluded for the Pacific model. These best-fitting symmetry axes are dipping slightly towards the west-northwest, very similar

to the relatively horizontal and dominantly arc-parallel symmetry axes imaged in the shallow wedge by local-S splitting tomography [Abt *et al.*, 2009]. However, symmetry axis dip and strength of anisotropy are not particularly well-resolved by the SK(K)S phases. For example, for the Caribbean mantle, the 10% misfit contour encompasses dip and strength parameters from $\gamma=0^\circ, \alpha=30\%$ to $\gamma=80^\circ, \alpha=90\%$, and shallow and steep dips also provide nearly equivalent fits to the Pacific data (Fig. 9e). This particular trade-off between strength and dip is expected given the decrease of intrinsic anisotropy of our elastic material as the symmetry axis rotates from horizontal to vertical for near-vertical ray propagation (Suppl. Fig. S1). Therefore, although our models are best-characterized by shallowly dipping symmetry axes, we stress that less emphasis should be placed on the interpretation of these dips. In contrast to dip and strength of anisotropy, normalized misfit values of $<10\%$ are confined to a much smaller range of q values in the a and y surfaces (Fig. 9a,c,d,f), providing us with greater confidence interpreting between arc-parallel and arc-normal symmetry axis azimuths (note the 180° symmetry in Fig. 9a,c,d,f that shows similar misfits for WNW and ESE azimuths).

4. Interpretation and Discussion

Assuming the origin of observed anisotropy is A-type dislocation creep in olivine [e.g., Zhang and Karato, 1995], the best-fitting symmetry axis orientations (Fig. 9) suggest roughly horizontal along-arc flow is occurring in both the Caribbean and Pacific mantle beneath southeastern Central America, although models with a more vertical

symmetry axis also fit the splitting data reasonably well. Here we discuss the possible origins of this flow.

Russo and Silver [1994] hypothesized that teleseismic-S and SKS arc-parallel fast polarization directions along the Andean margin are the result of roll-back of the Nazca Plate forcing Pacific mantle to the north and south of where flat-slab subduction is occurring (orange arrows in Figure 10). Subsequent splitting observations in the Caribbean and Scotia Seas [*Russo et al.*, 1996, *Müller et al.*, 2008; *Piñero-Feliciangeli and Kendall*, 2008; *Growdon et al.*, 2009] indicate this flow likely wraps around the edges of the Nazca Plate (empty dashed arrow in Figure 10), a pattern consistent with the Caribbean and Scotia Plates being driven eastward relative to South America as the Pacific mantle shrinks and the Atlantic/Caribbean mantle grows [e.g., *Elsasser*, 1971; *Alvarez*, 1982; *Garfunkel et al.*, 1986]. The tectonic feature allowing mantle to flow north from the Pacific realm into the Caribbean and often invoked to explain geochemical and seismological observations [e.g., *Herrstrom et al.*, 1995; *Johnston and Thorkelson*, 1997; *Sallarès et al.*, 2000; *Abratis and Wörner*, 2001] is generally thought to be a young (<10 Ma) slab window beneath southern Costa Rica, Panama, and northwestern Colombia, separating the Cocos and Nazca plates (Fig. 10b). Global isotropic velocity tomography models [e.g., *Li et al.*, 2008 (V_P); *Ritsema et al.*, 2004 (V_S)] may also indicate the absence of the Cocos Plate in this same region; the models lack clear evidence of high velocity anomalies in the upper mantle beneath southeastern Central America, as would be expected with the presence of a relatively cold slab.

Mantle flow to the north through a slab window and into the mantle wedge beneath Costa Rica and Nicaragua is consistent the local-S splitting tomography model as well as

trends in Pb and Nd isotope ratios that imply a northwest direction of wedge flow [Hoernle *et al.*, 2008; Abt *et al.*, 2009]. We continue to favor this interpretation and note that it could also be used to explain the deeper wedge anisotropy constrained by SK(K)S waves on the Caribbean side of the Cocos Plate (Fig. 6). Beneath the Cocos Plate, the Pacific SK(K)S splitting observations (Figs. 5-7) and corresponding model of anisotropy (Fig. 9d-f) imply flow roughly parallel to slab strike, but the direction of flow (west-northwest versus east-southeast) is not observationally constrained. Although the retreating Nazca Plate may drive regional flow that forces mantle northwest beneath the Cocos slab (as well as through the slab window), motion of the Cocos Plate and Middle America trench may also play a role generating and focusing mantle flow.

Different global plate motion models predict different absolute plate and trench velocities, depending on their reference frame, but the basic relative motions of the Cocos, North American, Caribbean, and Nazca Plates are consistent between them [Schellart *et al.*, 2008]. Plate and trench motions from one of these models (that in the Indo-Atlantic hot spot reference frame [DeMets *et al.*, 1994 O'Neill *et al.*, 2005]) are illustrated in Figure 10 by the black and blue arrows, respectively. In this reference frame, the Middle America trench shows an along-strike increase in southwestward motion from Costa Rica to southern Mexico that suggests a counter-clockwise rotation of the trench about a pole somewhere between southeastern Costa Rica and northwestern Nicaragua. Crustal compression in Costa Rica and extension in Nicaragua, along with the progressive trenchward migration of the volcanic arc in northern Nicaragua are consistent with this plate motion model, and suggest that this type of motion has been relatively long-lived (at least since the early Miocene [Weinberg, 1992]). Such a pattern

of trench retreat and slab roll-back could, as suggested in *Abt et al.* [2009], draw mantle wedge material to the northwest. Similarly, sub-slab mantle under these conditions could be driven to the east-southeast (yellow arrows in Fig. 10) and possibly entrained with the northward flow of mantle from beneath the Nazca Plate and forced through the slab window.

5. Conclusions

The primary objective of this study was to constrain the orientation of anisotropy outside the shallow mantle wedge beneath Costa Rica and Nicaragua. The observed SK(K)S splitting and best-fitting models of crystallographic orientation in both Caribbean (deeper back-arc wedge) and Pacific (sub-slab) mantle indicate the presence of arc-parallel symmetry axes (i.e., olivine *a*-axes). We interpret the anisotropy in the Caribbean mantle to result from along-arc flow to the WNW, originating from the slab window to the southeast beneath southern Costa Rica and Panama, as has been inferred in numerous previous studies [*Russo and Silver*, 1994; *Herrstrom et al.*, 1995; *Johnston and Thorkelson*, 1997; *Abratis and Wörner*, 2001; *Hoernle et al.*, 2008; *Abt et al.*, 2009]. Flow beneath the Cocos Plate also appears to be roughly parallel to the strike of the slab. In contrast to wedge flow, the absolute direction of sub-slab flow is unconstrained by geochemical data, but sub-slab flow could be driven ESE by along-arc gradients in slab rollback as well as regional trench and plate motions.

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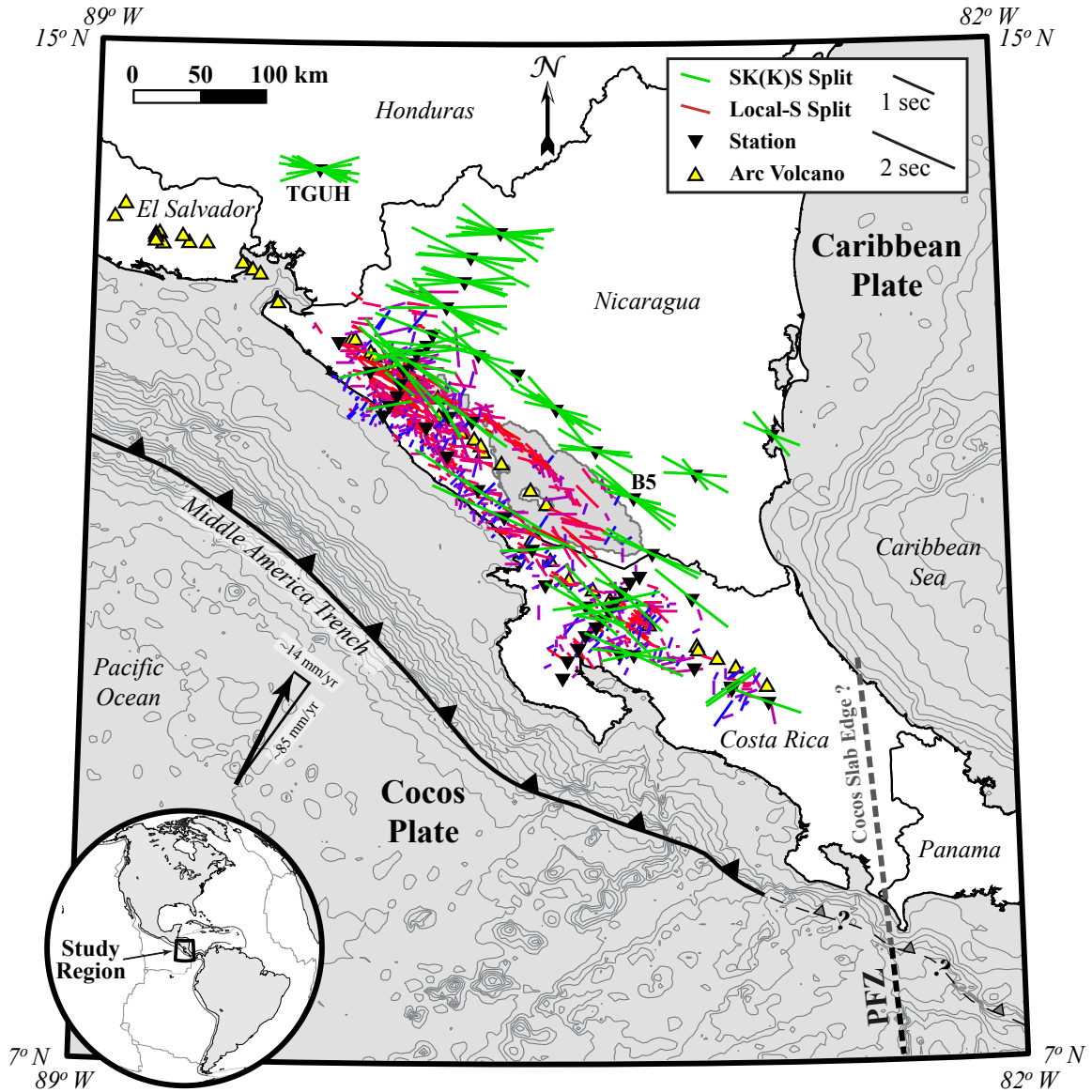


Figure 1. Map of SK(K)S and local-S [Abt *et al.*, 2009] splitting observations. SK(K)S splitting vectors (green) are plotted at the station where they were measured, while local-S splits are plotted at ray path midpoints. Vector orientation is parallel to fast polarization direction and scaled to delay time (reference dt of 1 and 2 sec are shown in the legend). The color of the local-S vectors indicates azimuth as well and grades from red (arc-parallel) to blue (arc-normal). Black triangles are seismic stations and yellow triangles are arc volcanoes; stations TGUH (splits not included in modeling) and B5 (example split shown in Figs. 3,8) are identified. Cocos Plate motion relative to the Caribbean Plate (black arrow) is from DeMets [2001]. The Panama fracture zone (PFZ) and inferred edge of the Cocos slab [e.g., Protti *et al.*, 1994; Herrstrom *et al.*, 1995; Johnston and Thorkelson, 1997] are also shown with the dashed black and gray lines, respectively. The two major differences between the local-S and SK(K)S splitting parameters are readily apparent here: (1) SK(K)S fast directions are much more consistent (arc-parallel) than local-S fast directions, and (2) local-S delay times are generally much smaller than SK(K)S delay times.

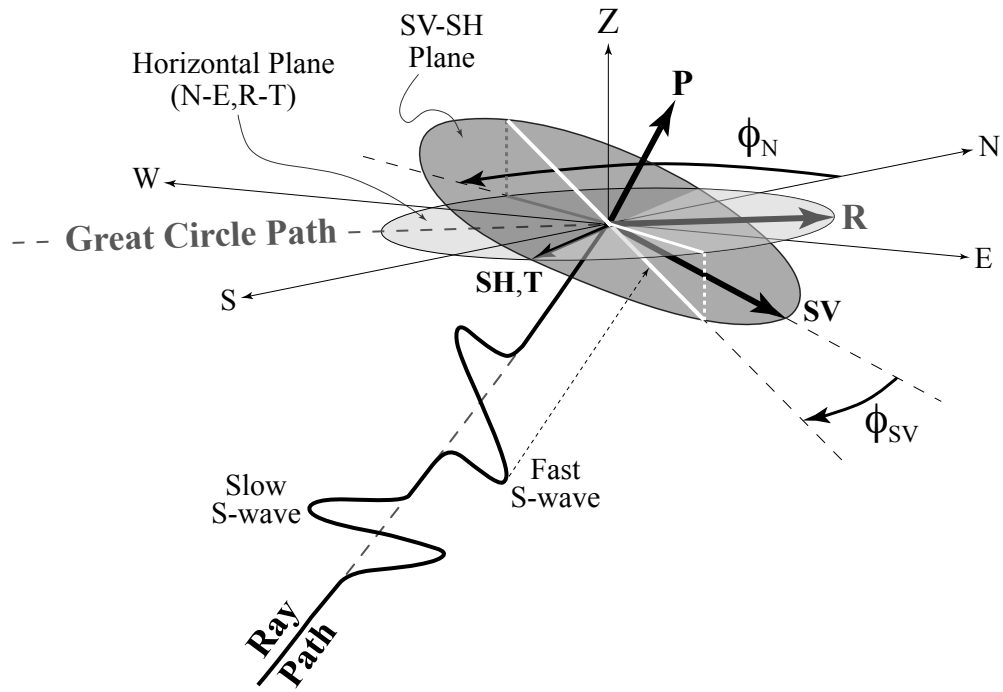


Figure 2. Schematic illustration of the relationship between the Z-R-T and P-SV-SH coordinate systems. The free-surface transform [Kennett, 1991] assuming $V_P = 5.9$ km/sec and $V_S = 3.41$ km/sec is used to obtain the P-SV-SH components. We measure fast polarization direction clockwise from SV in the SV-SH plane (the white line represents the splitting vector). Figure 3 shows an example of the (minor) differences in waveforms and associated splitting parameters between the two coordinate systems.

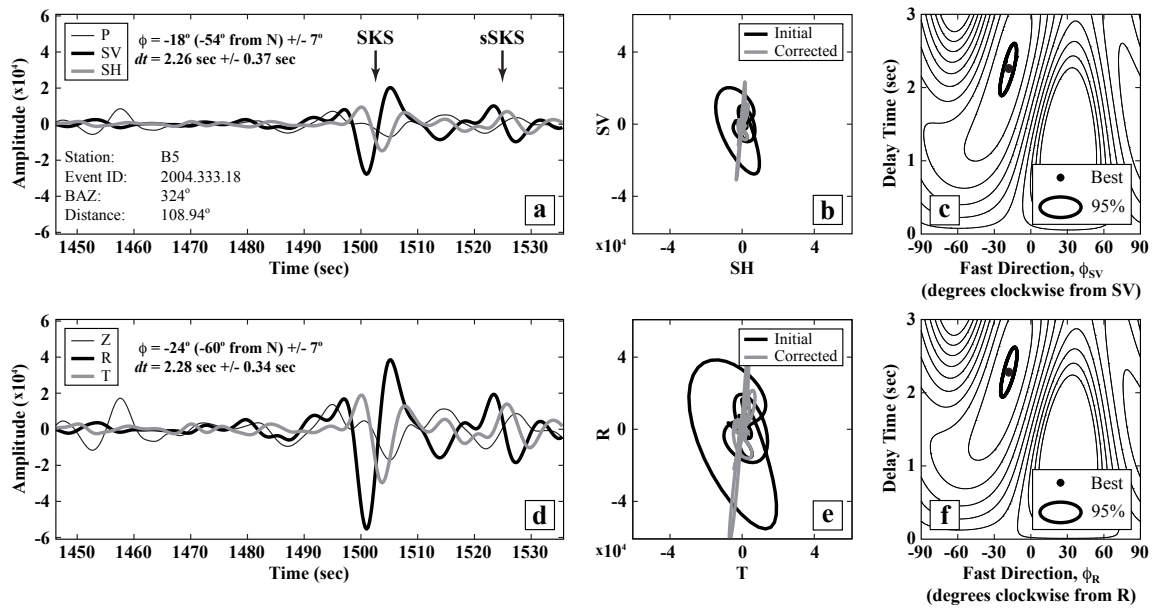


Figure 3. Example splitting parameter calculation in the SV-SH (a-c) and R-T (d-f) planes. The splitting parameters shown in all other figures and referred to throughout the text are those measured in the SV-SH plane. The windowed waveforms used in the calculation are illustrated in (a,d), particle motions are shown in (b,e), and the λ_2 (see Section 2.1) values are contoured around the best-fitting splitting parameter pair (black dot) in (c,f), with the 95% confidence region shown with the thick black line. The event for this particular example was at a depth of 39 km and produced at very clear sSKS arrival, which has been included in the analysis window. This secondary phase can be seen on both the waveforms and in the initial particle motion plots (black line in b and c), with its polarity reversed relative to that of the SKS phase. As expected and demonstrated by the linear corrected particle motion (gray line in b and c), the splitting of both phases is well-characterized by the same pair of ϕ and dt .

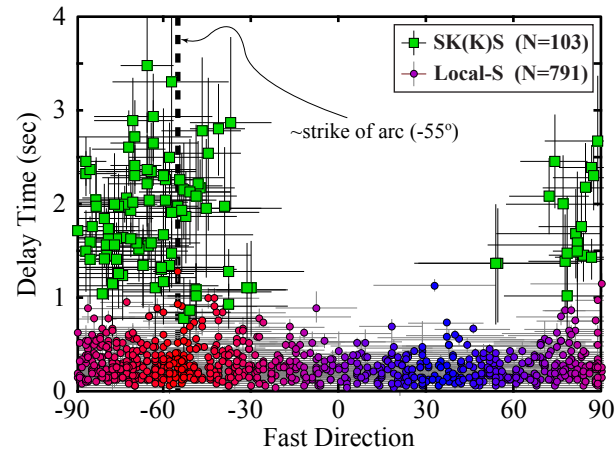


Figure 4. Comparison of SK(K)S and local-S splitting parameters. SK(K)S splits are shown in green and local-S splits grade from red (arc-parallel ϕ) to blue (arc-normal ϕ). Error bars on both are the standard splitting errors (e.g., measured from the 95% confidence contour in Fig. 3c; see Section 2.1). The approximate strike of the volcanic arc (N55°W) is shown with the black dashed line. As can also be seen in map view (Fig. 1), SK(K)S fast directions do not show the same variability as local-S fast directions, and delay times for SK(K)S are much larger.

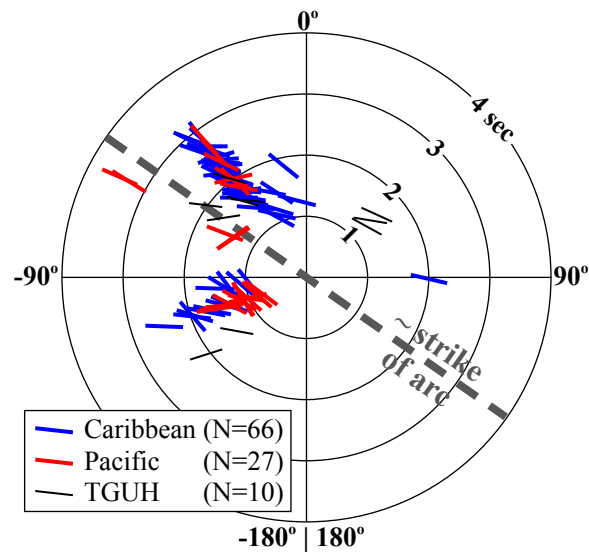


Figure 5. SK(K)S splitting parameters plotted radially as a function of back azimuth and delay time. Vector orientation is parallel to fast direction, with up being north. Splits at TUCAN stations from waves that sample predominantly Caribbean mantle are colored blue and those sampling mostly Pacific mantle are red (see Figure 6 for ray paths). Black vectors are splits at station TGUH. Again, the approximate strike of the arc is shown with the dashed gray line.

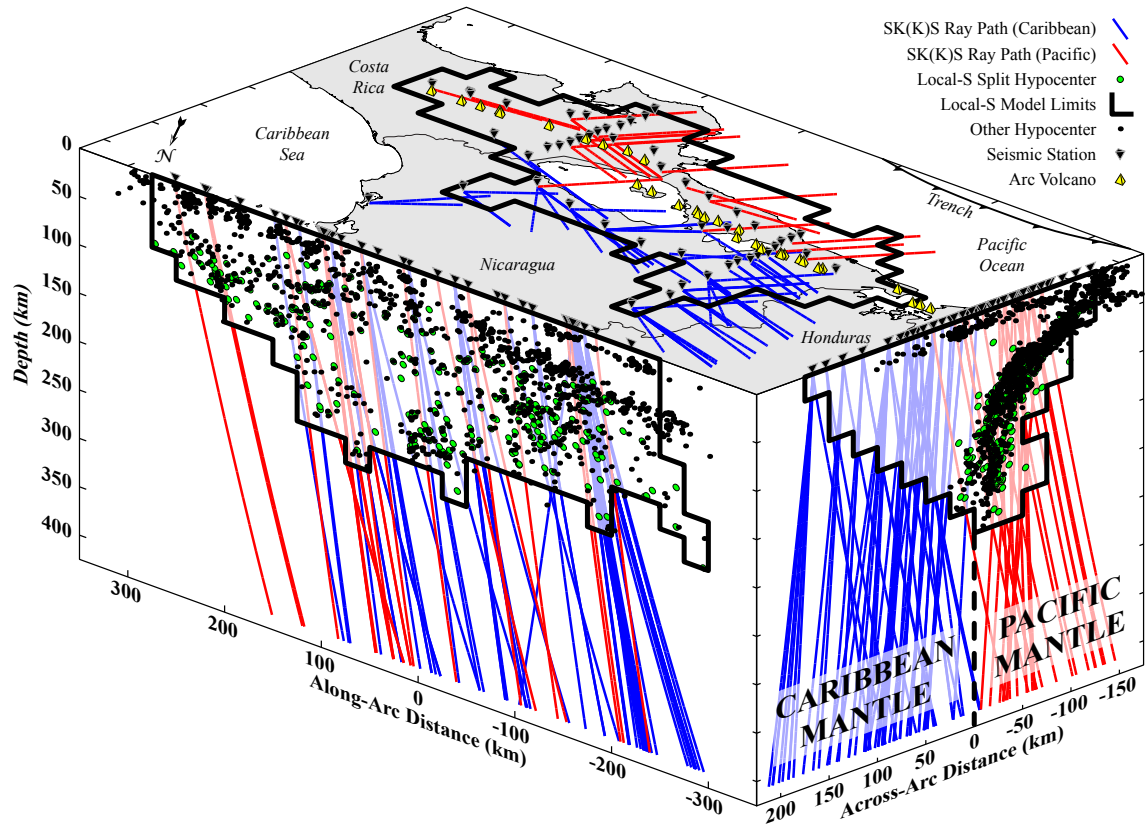


Figure 6. TUCAN SK(K)S ray paths beneath Costa Rica and Nicaragua and through the local-S splitting tomography model space. Rays are traced through AK135 [Kennett *et al.*, 1995]; blue paths are those dominantly traveling through Caribbean upper mantle (i.e., deeper wedge), and red paths are those dominantly in Pacific mantle (i.e., sub-slab). The limits of the well-resolved model space from local-S tomography are shown with the thick black line. Hypocenters from local-S splits are the green circles, and the remainder of the re-located local event catalog [Syracuse *et al.*, 2008] is shown with smaller black circles.

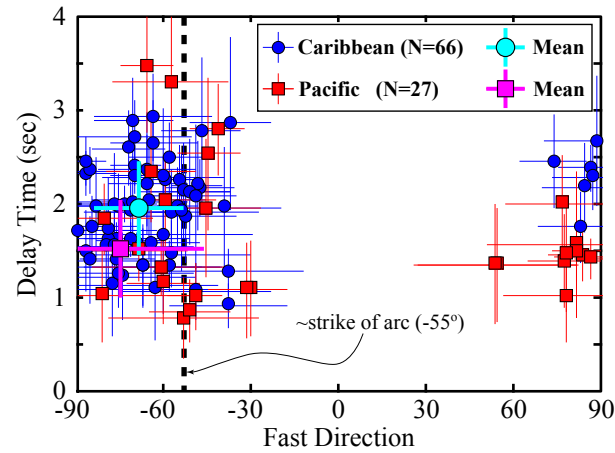


Figure 7. Comparison of splitting parameters from waves sampling Caribbean and Pacific mantle. The mean values for the two regions are shown with the larger cyan circle (Caribbean) and magenta square (Pacific), with the bars representing ± 1 standard deviation. The smaller blue circles and red squares (with standard splitting errors) are the individual splitting parameters for the Caribbean and Pacific paths, respectively. The Pacific splits have a smaller mean dt relative to Caribbean splits (0.30 sec or $\sim 15\%$) and also display less coherence in both ϕ and dt .

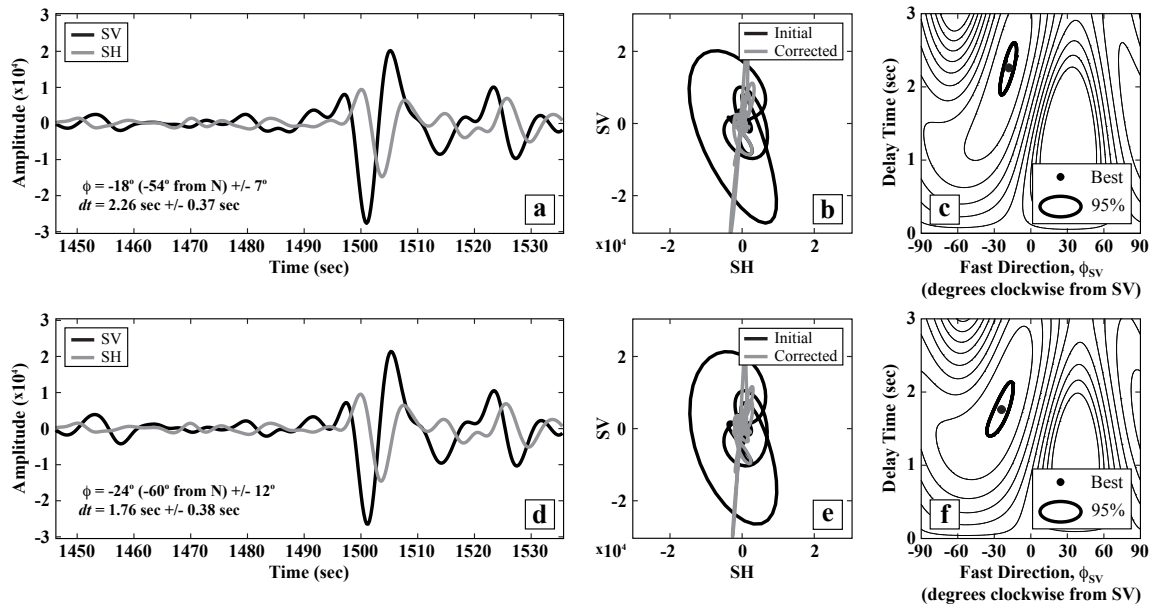


Figure 8. Example of backing out the effects of anisotropy in the local-S splitting tomography model [Abt *et al.*, 2009] on SK(K)S waveforms. The top row is the same as Figure 3a-c, and the bottom row illustrates the waveforms (d), particle motion (e), and best-fitting splitting parameters (f) after accounting for anisotropy in 6 well-resolved local-S blocks over a path length of 127 km (i.e., the expected observation at the entry point in the local-S model). Although the two sets of waveforms, particle motions, and λ_2 surfaces look almost identical to the eye (due to the relatively long analysis window), the delay time for each is modestly different: 1.76 sec at the local-S model entry point compared with 2.26 sec at the surface ($\sim 28\%$ larger at the surface). We use these estimated splitting parameters (bottom row) to establish the orientation of anisotropy outside the shallow mantle wedge.

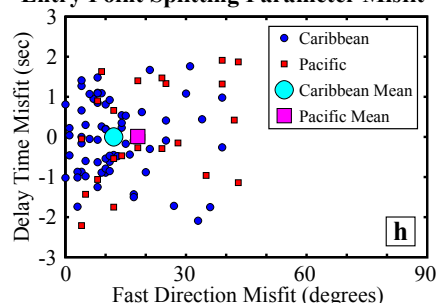
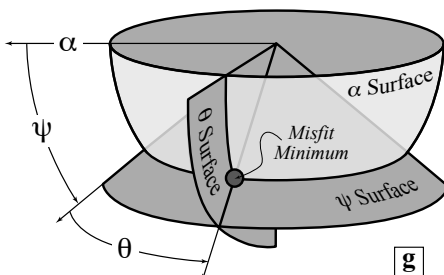
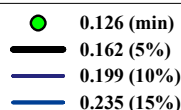
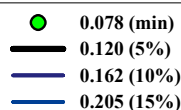
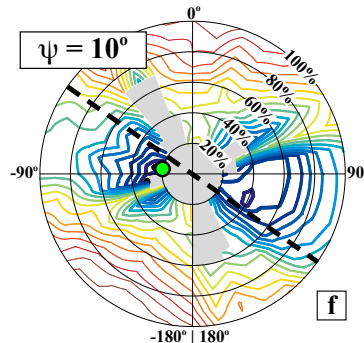
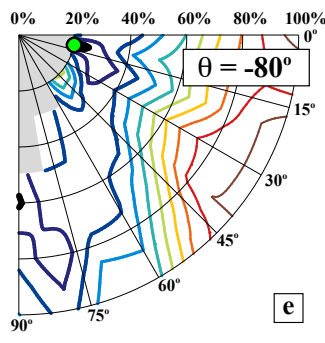
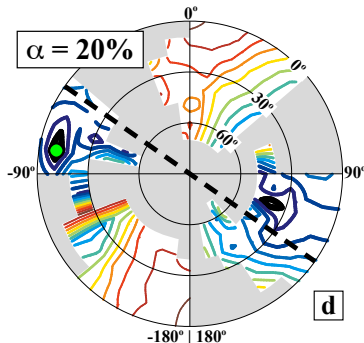
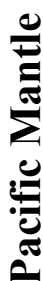
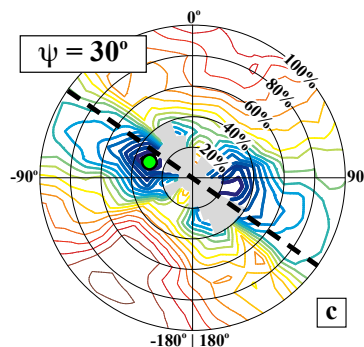
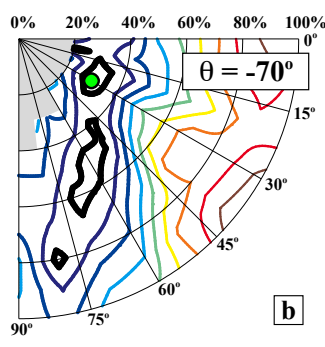
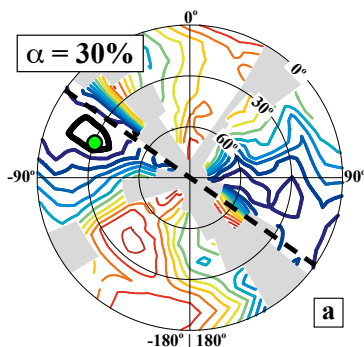


Figure 9. Results of a grid search over symmetry axis azimuth (θ), symmetry axis dip (ψ), and strength of anisotropy (α) for the model of anisotropy that minimizes misfit between predicted splitting parameters and those estimated from waveforms at their entry point in the local-S splitting tomography model (i.e., the shallow mantle wedge). The surfaces in (a-f) are schematically illustrated in (g) relative to the volume of parameter space searched. Note that due to the symmetry of the elastic tensor, we only need to search the lower hemisphere (i.e., dips down from horizontal). The surfaces in (a-c) are contoured at 5% increments between the minimum (0.078) and maximum (0.923) normalized misfit values for the Caribbean mantle splits, and the normalized misfit range for Pacific splits (0.126-0.853) is contoured in (d-f); note that contours between the Caribbean and Pacific plots do not represent exactly the same misfit values. (a,d) show surfaces with a constant strength of anisotropy, (b,e) show surfaces with a constant symmetry axis azimuth, and (c,f) show surfaces with a constant symmetry axis dip. Note the apparent 180° symmetry for θ (i.e., in a,c,d,f); this is a result of the symmetry of the stiffness tensor and is not perfect here because of the non-uniform and non-vertical SK(K)S propagation directions. The gray regions show the combinations of model parameters that resulted in more than 50% of the predicted splitting parameter measurements being unstable or null-like. The dashed black line in (a,c,d,f) shows the approximate strike of the arc. The splitting parameter misfits for these best-fitting models are displayed in (h). As might be expected given the greater spread in splitting parameters from Pacific splits (Fig. 7), the mean weighted misfit for Pacific mantle entry point splits is also greater than that for Caribbean splits (18° vs. 12° for fast direction misfit). For a particular region, delay times are both under predicted (dt misfit >0) and over predicted (dt misfit <0), resulting in mean weighted dt misfits of 0.00 and 0.01 sec for the Caribbean and Pacific splits, respectively; the mean weighted $|dt|$ misfits are 0.72 and 1.00 sec, respectively.

#	Event ID YEAR.JDY.HR	Station	BAZ deg	DIST deg	Phase	ϕ deg	σ_ϕ deg	dt sec	σ_{dt} sec
1	2004.249.10	TISM	319	119	SKKS	-54	26	1.92	0.63
2	2004.250.23	B3	320	119	SKKS	-66	14	2.21	0.53
3	2004.250.23	N12	320	118	SKKS	-64	20	2.92	0.47
4	2004.250.23	N13	320	118	SKKS	-89	27	1.71	0.48
5	2004.250.23	N9	319	118	SKKS	-70	20	2.71	0.39
6	2004.282.08	B3	263	114	SKS	-49	26	1.07	0.30
7	2004.282.08	B4	263	114	SKS	-87	20	1.50	0.42
8	2004.282.08	C12	263	115	SKS	-38	21	0.92	0.23
9	2004.282.08	C13	263	116	SKS	-38	26	1.28	0.25
10	2004.316.17	B3	263	114	SKS	-58	18	1.34	0.27
11	2004.322.21	B3	251	97	SKS	-39	20	1.96	0.65
12	2004.322.21	C13	251	99	SKS	-67	26	1.34	0.44
13	2004.322.21	N10	251	97	SKS	-67	21	1.33	0.40
14	2004.322.21	N11	251	97	SKS	-71	16	1.92	0.32
15	2004.322.21	RITO	251	97	SKS	-61	24	1.32	0.30
16	2004.322.21	TESU	251	97	SKS	82	31	1.57	0.78
17	2004.333.18	B1	323	107	SKS	-48	12	2.17	0.30
18	2004.333.18	B3	324	108	SKS	-56	7	1.97	0.41
19	2004.333.18	B4	324	108	SKS	-57	7	1.91	0.45
20	2004.333.18	B4	324	108	SKKS	-79	28	1.63	0.65
21	2004.333.18	B5	324	109	SKS	-54	7	2.26	0.37
22	2004.333.18	B5	324	109	SKKS	-65	20	2.04	0.61
23	2004.333.18	C12	324	109	SKS	-60	31	1.67	0.83
24	2004.333.18	COLI	323	109	SKS	-46	19	1.95	0.73
25	2004.333.18	COVE	324	110	SKS	-49	14	2.08	0.76
26	2004.333.18	LSOL	323	110	SKKS	-65	15	2.34	0.53
27	2004.333.18	N10	323	107	SKS	-70	9	2.41	0.21
28	2004.333.18	N10	323	107	SKKS	89	19	2.67	0.69
29	2004.333.18	N11	323	107	SKS	-74	11	2.00	0.20
30	2004.333.18	N12	323	107	SKS	-76	15	1.97	0.36
31	2004.333.18	N12	323	107	SKKS	-86	11	2.35	0.25
32	2004.333.18	N13	324	107	SKS	-69	9	1.58	0.12
33	2004.333.18	N13	324	107	SKKS	87	11	2.38	0.25
34	2004.333.18	N5	323	107	SKS	-47	14	2.77	0.78
35	2004.333.18	N6	323	107	SKS	-58	14	2.49	0.38
36	2004.333.18	N7	323	107	SKS	83	14	1.76	0.38
37	2004.333.18	N8	323	107	SKS	-85	16	1.76	0.32
38	2004.333.18	RITO	324	109	SKS	-45	11	2.53	0.80
39	2004.333.18	TABL	324	109	SKKS	-60	21	2.29	0.73
40	2004.333.18	TENO	323	110	SKS	77	20	2.01	0.51
41	2004.333.18	TESU	323	110	SKS	-42	10	2.79	0.49
42	2004.341.14	B1	323	107	SKS	-60	10	2.26	0.27
43	2004.341.14	B3	323	108	SKS	-65	17	1.58	0.48
44	2004.341.14	B5	324	109	SKS	-49	9	2.20	0.48
45	2004.341.14	C12	324	109	SKS	-85	24	1.40	0.46
46	2004.341.14	COLI	323	109	SKS	-60	18	2.04	0.50
47	2004.341.14	LSOL	323	110	SKKS	-81	17	1.84	0.37
48	2004.341.14	N10	323	107	SKS	-70	8	2.30	0.19
49	2004.341.14	N10	323	107	SKKS	-72	13	2.59	0.39
50	2004.341.14	N11	323	107	SKS	-87	11	2.32	0.24
51	2004.341.14	N13	323	107	SKS	-72	12	1.62	0.19
52	2004.341.14	N13	323	107	SKKS	87	14	2.29	0.34

Table 1. Observed SK(K)S splitting parameters. Event information for each split can be obtained by associating Event IDs given here with the year, Julian day, and hour in Table 2. Fast directions are measured east of north. Errors (σ_ϕ and σ_{dt}) are measured on the 95% confidence contour of the λ_2 surface (e.g., Fig. 3c, Section 2.1).

#	Event ID YEAR.JDY.HR	Station	BAZ deg	DIST deg	Phase	ϕ deg	σ_ϕ deg	dt sec	σ_{dt} sec
53	2004.341.14	N5	323	107	SKS	-37	14	2.86	0.92
54	2004.341.14	N6	323	107	SKS	-51	14	2.13	0.56
55	2004.341.14	N7	323	107	SKS	-77	27	1.55	0.44
56	2004.341.14	N8	323	107	SKS	74	12	2.45	0.50
57	2004.341.14	TABL	324	109	SKKS	-66	17	2.36	0.61
58	2005.036.12	F5	298	146	SKKS	-58	20	3.29	0.85
59	2005.036.12	F6	298	147	SKKS	-66	9	3.46	0.70
60	2005.036.12	HDC	298	149	SKKS	55	28	1.35	0.61
61	2005.036.12	HDC5	298	149	SKKS	54	28	1.36	0.64
62	2005.036.12	IRZU	298	149	SKKS	-69	17	1.51	0.34
63	2005.100.10	B3	334	168	SKKS	-63	30	1.09	0.53
64	2005.100.10	B4	335	169	SKKS	-78	37	1.14	0.54
65	2005.100.10	B5	336	169	SKKS	-74	26	1.23	0.46
66	2005.100.10	N13	335	167	SKKS	-80	9	1.55	0.41
67	2005.134.05	N10	341	166	SKKS	-57	27	1.47	0.35
68	2005.323.14	B5	353	166	SKKS	-76	26	1.26	0.37
69	2005.323.14	N10	348	165	SKKS	-53	18	1.86	0.48
70	2005.336.13	N10	321	112	SKS	-70	14	2.88	0.46
71	2005.336.13	N10	321	112	SKKS	-63	15	2.64	0.45
72	2005.336.13	N11	321	112	SKKS	85	15	2.18	0.48
73	2005.336.13	N13	321	112	SKS	-53	16	2.14	0.52
74	2005.336.13	N13	321	112	SKKS	-71	19	2.02	0.39
75	2005.339.12	N6	91	117	SKS	-77	12	2.00	0.54
76	2006.002.22	B1	251	96	SKS	-76	11	1.63	0.19
77	2006.002.22	B3	251	97	SKS	-76	13	1.41	0.14
78	2006.002.22	B5	251	97	SKS	-60	13	1.16	0.45
79	2006.002.22	COLI	251	96	SKS	84	9	1.44	0.26
80	2006.002.22	F2	251	95	SKS	78	13	1.39	0.50
81	2006.002.22	F5	251	96	SKS	-53	26	0.78	0.43
82	2006.002.22	LSOL	251	97	SKS	79	7	1.47	0.61
83	2006.002.22	MGA1	251	96	SKS	-30	12	1.10	0.51
84	2006.002.22	MGA2	251	96	SKS	-31	13	1.09	0.51
85	2006.002.22	N11	251	96	SKS	-79	9	1.72	0.22
86	2006.002.22	N12	251	96	SKS	-87	7	2.45	0.25
87	2006.002.22	N13	251	97	SKS	-83	12	1.97	0.52
88	2006.002.22	N2	251	96	SKS	79	22	1.02	0.50
89	2006.002.22	N4	251	96	SKS	-51	14	0.86	0.33
90	2006.002.22	PALM	251	96	SKS	-81	31	1.03	0.50
91	2006.002.22	SONZ	251	96	SKS	87	6	1.42	0.22
92	2006.002.22	TESU	251	97	SKS	82	8	1.50	0.33
93	2006.002.22	TISM	251	96	SKS	-49	10	1.02	0.35
94	2007.271.13	TGUH	119	306	SKS	-83	10	2.03	0.42
95	2007.271.13	TGUH	119	306	SKKS	82	27	1.67	0.48
96	2007.354.07	TGUH	102	232	SKS	-80	20	1.44	0.29
97	2007.354.07	TGUH	102	232	SKKS	72	14	2.07	0.41
98	2008.045.10	TGUH	96	50	SKS	-61	26	1.29	0.26
99	2008.063.09	TGUH	99	323	SKS	-85	13	1.58	0.21
100	2008.197.03	TGUH	101	48	SKS	-81	6	1.4	0.09
101	2008.197.03	TGUH	101	48	SKKS	-66	14	1.54	0.24
102	2008.201.02	TGUH	111	320	SKKS	-74	17	1.63	0.35
103	2008.205.15	TGUH	110	322	SKKS	-73	15	2.05	0.34

Table 1. (continued) See caption on previous page.

Origin Time YEAR.JDY.HR.MN.SC	Latitude (deg)	Longitude (deg)	Depth (km)	Magnitude (M _w)	Back Azimuth (deg)	Distance (deg)	# of Splits
2004.249.10.07.07	33.07	136.62	14	7.2	320	120	1
2004.250.23.29.35	33.21	137.23	10	6.6	319	120	4
2004.282.08.27.53	-10.95	162.16	36	6.9	263	114	4
2004.316.17.34.52	-11.13	162.21	10	6.7	263	114	1
2004.322.21.09.13	-20.07	-178.71	622	6.6	251	97	6
2004.333.18.32.14	43.01	145.12	39	7.0	323	109	25
2004.341.14.15.11	42.9	145.23	35	6.8	323	109	16
2004.361.04.21.29	6.91	92.96	39	7.5	5	162	0
2005.036.12.23.18	5.29	123.34	525	7.1	299	147	5
2005.100.10.29.11	-1.64	99.61	19	6.7	332	169	4
2005.101.12.20.05	-3.48	145.91	11	6.7	275	128	0
2005.132.11.15.35	-57.38	-139.23	10	6.5	206	82	0
2005.134.05.05.18	0.59	98.46	34	6.8	342	167	1
2005.165.17.10.12	51.24	179.31	17	6.8	321	84	0
2005.323.14.10.13	2.16	96.79	21	6.5	350	166	2
2005.336.13.13.09	38.09	142.12	29	6.5	321	113	5
2005.339.12.19.56	-6.22	29.83	22	7.2	91	116	1
2006.002.06.10.49	-60.96	-21.61	13	7.4	154	88	0
2006.002.22.13.40	-19.93	-178.18	582	7.2	251	96	18
2007.271.13.38.57	22.01	142.67	260	7.5	306	119	2
2007.354.07.55.15	-39.01	178.29	20	6.6	232	102	2
2008.045.10.09.22	36.5	21.67	29	6.9	50	96	1
2008.063.09.31.02	46.41	153.18	10	6.5	323	99	1
2008.197.03.26.35	35.99	27.71	52	6.5	48	101	2
2008.201.02.39.28	37.55	142.21	22	7	320	111	1
2008.205.15.26.20	39.81	141.47	111	6.8	322	110	1

Table 2. SK(K)S event origin times and location information. Back azimuth and distance for all TUCAN splits (#1-93) are measured from a rough estimate of the TUCAN seismic array center: 11.5°N, 85.5°W. For TGUH, the back azimuth and distance are measured from the station.

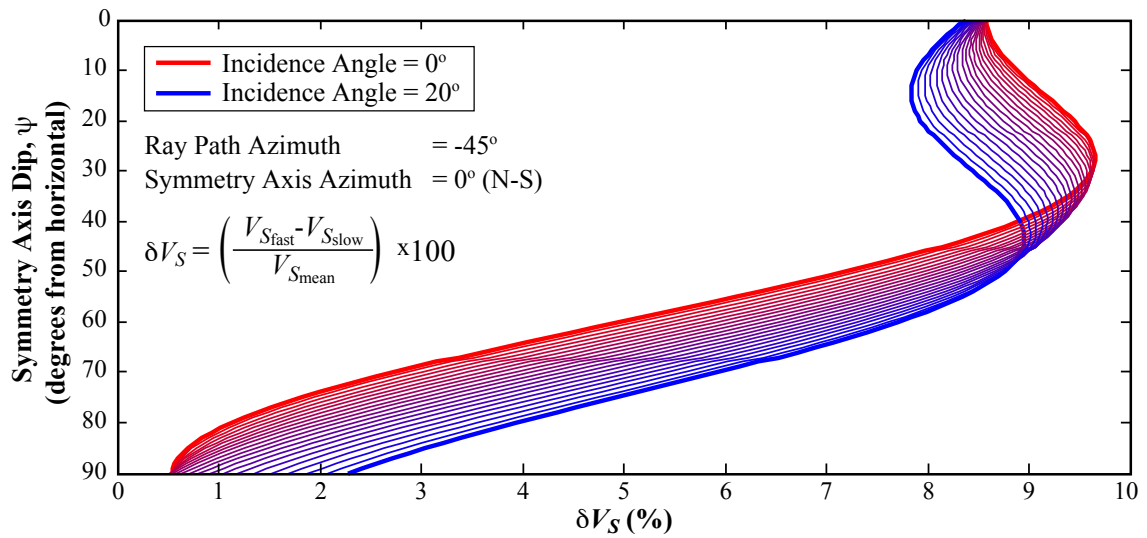


Figure S1. Predicted strength of shear-wave anisotropy for a range of symmetry axis dips and ray path incidence angles. Note that here we are referring to δV_S as a percentage of anisotropy and not α , as in the tomography model parameterization; δV_S is the difference between the fast and slow shear-wave velocities relative to their mean. For this example, initial polarization is 45° from the symmetry axis azimuth (0° or N-S) and the hexagonally averaged elastic properties of 70% olivine and 30% opx (Suppl. Table S1) at 100 km depth are assumed, as well as a temperature profile with a 30 km conductive lid overlying an adiabatic mantle gradient with a temperature of 1350°C just below the lid. Line color grades from red (vertical incidence) to blue (incidence angle of 20°). For a typical SK(K)S path (~10° incidence angle), δV_S starts at ~8.5% for a horizontal symmetry axis, drops to ~8.2% for a 15° symmetry axis dip, peaks at ~9.2% for a 35° dip, and then decreases to ~0.8% for a 90° dip (vertical). These values will not change considerably for different ray path azimuths (i.e., initial polarizations) and symmetry axis azimuths.

ij	Olivine*, Fo ₉₀ (Mg _{1.8} Fe _{0.2} SiO ₄)			OPX**, Bronzite (Mg _{0.8} Fe _{0.2} SiO ₃)		
	C_{ij} (at 1 bar)	$\frac{\delta C_{ij}}{\delta P}$	$\frac{\delta C_{ij}}{\delta T}$	C_{ij} (at 1 bar)	$\frac{\delta C_{ij}}{\delta P}$	$\frac{\delta C_{ij}}{\delta T}$
	units	GPa	-	GPa	-	GPa · K ⁻¹
11	323.7	7.98	-0.034	228.6	11.04	-0.0352
12	66.4	4.74	-0.0105	79.9	6.97	-0.0212
13	71.6	4.48	-0.0094	63.2	9.09	-0.0318
22	197.6	6.37	-0.0285	160.5	9.19	-0.0328
23	75.6	3.76	-0.0051	56.8	8.73	-0.0107
33	235.1	6.38	-0.0286	210.4	16.42	-0.0516
44	64.62	2.17	-0.0128	81.75	2.38	-0.0131
55	78.05	1.64	-0.013	75.48	2.92	-0.0138
66	79.04	2.31	-0.0157	77.6	2.75	-0.0145

*Anderson and Isaak (1995) and Abramson et al. (1997); **Frisillo and Barsch (1972)

$$C_{11} = c_{1111}$$

$$C_{12} = c_{1122} = c_{2211}$$

$$C_{44} = c_{2323} = c_{2332} = c_{3232} = c_{3223}$$

$$C_{22} = c_{2222}$$

$$C_{23} = c_{2233} = c_{3322}$$

$$C_{55} = c_{1313} = c_{1331} = c_{3131} = c_{3113}$$

$$C_{33} = c_{3333}$$

$$C_{13} = c_{1133} = c_{3311}$$

$$C_{66} = c_{1212} = c_{1221} = c_{2121} = c_{2112}$$

Orthorhombic



Hexagonal

$$C_{ij} = \begin{bmatrix} C_{11} & C_{12} & C_{13} & 0 & 0 & 0 \\ C_{21} & C_{22} & C_{23} & 0 & 0 & 0 \\ C_{31} & C_{32} & C_{33} & 0 & 0 & 0 \\ 0 & 0 & 0 & C_{44} & 0 & 0 \\ 0 & 0 & 0 & 0 & C_{55} & 0 \\ 0 & 0 & 0 & 0 & 0 & C_{66} \end{bmatrix} = \begin{bmatrix} A & F & A-2N & 0 & 0 & 0 \\ F & C & F & 0 & 0 & 0 \\ A-2N & F & A & 0 & 0 & 0 \\ 0 & 0 & 0 & L & 0 & 0 \\ 0 & 0 & 0 & 0 & N & 0 \\ 0 & 0 & 0 & 0 & 0 & L \end{bmatrix}$$

$$A = \frac{3}{8}(C_{11} + C_{33}) + \frac{1}{4}C_{13} + \frac{1}{2}C_{55}$$

$$N = \frac{1}{8}(C_{11} + C_{33}) - \frac{1}{4}C_{13} + \frac{1}{2}C_{55}$$

$$C = C_{22}$$

$$F = \frac{1}{2}(C_{12} + C_{23})$$

$$L = \frac{1}{2}(C_{44} + C_{66})$$

Montagner and Anderson [1989]

Table S1. Elastic coefficients assumed in the calculation of predicted splitting parameters (Section 3). The orthorhombic coefficients of olivine (Fo₉₀) and opx (Bronzite) with their pressure and temperature derivatives (**top**) and their transformation to hexagonal coefficients (**bottom**, from Montagner and Anderson [1989]) are given. The *c*-axis of opx is aligned parallel to the *a*-axis of olivine [Mainprice and Silver, 1993]. See Abt et al. [2009] for further details.

Chapter 4

North American Lithospheric Discontinuity Structure Imaged By *Ps* and *Sp* Receiver Functions

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Abstract

Ps and *Sp* converted seismic waves recorded at 93 permanent seismic stations across the continental United States, northwestern Mexico, and southeastern Canada are used to image the discontinuity structure of the upper mantle. Receiver functions are calculated with a frequency-domain deconvolution and migrated with models that account for three-dimensional velocity variations. Strong positive *Ps* phases from the Moho are observed across the study region and agree well with previous crustal thickness estimates. Prominent *Sp* phases from a negative velocity contrast are found at depths of 51-104 km in much of the western U.S., while similar phases with smaller amplitude in the central and eastern U.S. migrate to depths of 57-122 km. To interpret these *Sp* phases as either the lithosphere-asthenosphere boundary (LAB) or a mid-lithospheric discontinuity (MLD), we compare their depth to the transition from the low-velocity zone to the higher velocity lid (i.e., the asthenosphere to the lithosphere) in the shear-velocity model used to migrate the receiver functions.

In the tectonically active western U.S., the depth (51-104 km) and large amplitude of the *Sp* phases are consistent with other lines of evidence (e.g., surface and body wave tomography, heat flow, recent volcanism) that indicate an anomalously hot and shallow asthenosphere. Although the amplitude of these phases are consistent with the presence of melt or high volatile content in the asthenosphere, rapid thermal gradients at the LAB due to localized upwelling cannot be ruled out. In the portions of the eastern U.S. where the *Sp* phases are interpreted as the LAB at depths of 75-111 km, their amplitude also suggests that the discontinuity cannot be explained by purely thermal gradients and that some contribution from partial melt and/or volatiles in the asthenosphere is required,

perhaps in conjunction with vertical variations in anisotropy. In the thicker high velocity lid beneath cratonic North America and parts of the Phanerozoic eastern U.S., where the *Sp* phases are interpreted as the top of a low velocity zone internal to the lithosphere at depths of 59-113 km, they could reflect layering related to cratonic formation or to its subsequent modification by upward percolating melts and metasomatic fluids. The absence of an observable *Sp* phase from a discontinuity at the base of the thickest North American lithosphere indicates the LAB velocity gradient is likely distributed over more than 30 km in depth, and therefore may be explained by temperature alone. Thus, the character of the LAB appears to fundamentally differ between regions of thicker (>150 km) and thinner (<120 km) lithosphere.

1. Introduction

Determining the location and physical properties of the lithosphere-asthenosphere boundary (LAB) is essential to our understanding of plate tectonics and the evolution of the Earth. More specifically, the presence of variations in the depth and character of the LAB across and within different tectonic environments (e.g., cratons, passive margins, extensional provinces, subduction zones) could provide insight into the growth and stability of lithospheric plates as well as the mechanism(s) responsible for their plate-like behavior. Ancient cratonic regions are expected to contain thick lithospheric roots that are rheologically strong with a chemical buoyancy that counteracts their cold temperatures [e.g., *Jordan, 1978; Jordan, 1988; Anderson, 1989; Boyd, 1989; Griffin et al., 1999; Sleep, 2005*], making them stable relative to the surrounding upper mantle [e.g., *Pearson et al., 1995; Carlson et al., 1999; Shapiro et al., 1999*]. A key question that we will address in this paper is whether the properties of the cratonic LAB fundamentally differ from the LAB in regions that have experienced more recent tectonic activity.

The North American lithosphere has a rich tectonic history that includes the progressive accretion of lithotectonic terranes around Archean cratons (e.g., Superior, Wyoming, Hearne, Slave) during the Proterozoic, successful and failed rifting events, multiple episodes of subduction, and current large-scale extension in the Basin and Range province (Fig. 1) [e.g., *Rivers, 1997; Karlstrom and Humphreys, 1998; Chulick and Mooney, 2002; Griffin et al., 2004*]. Numerous global and regional tomographic studies [e.g., *Grand and Helmberger, 1984; Grand, 1994; van der Lee and Nolet, 1997; van der Lee, 2002; Li et al., 2003; Godey et al., 2004; Marone et al., 2007; Xue and Allen, 2007; Nettles and Dziewonski, 2008; Roth et al., 2008*] have interpreted lithospheric thickness

across North America from relatively long wavelength images of velocity heterogeneity in the upper mantle. However, the surface and body waves typically employed in continental-scale tomography are not able to resolve vertical velocity gradients with a precision better than 30-50 km, thus limiting their ability to accurately locate boundaries such as the LAB, which in certain regions appear to be relatively sharp (e.g., 5-10 km [Rychert *et al.*, 2005; Chen *et al.*, 2006]). In contrast, converted phases such as *P*-to-*s* (*Ps*) and *S*-to-*p* (*Sp*) are well suited for interrogating such upper mantle discontinuities (e.g., the *Moho*, LAB, 410-km and 660-km), and in recent years these phases have been widely used to image the LAB and discontinuities within the lithosphere [e.g., Sacks *et al.*, 1979; Bostock, 1998; Li *et al.*, 2000; Collins *et al.*, 2002; Oreshin *et al.*, 2002; Li *et al.*, 2004; Kumar *et al.*, 2005a,b; Rychert *et al.*, 2005, 2007; Vinnik *et al.*, 2005; Chen *et al.*, 2006; Mohsen *et al.*, 2006; Sodoudi *et al.*, 2006a,b; Wolbern *et al.*, 2006; Hansen *et al.*, 2007; Heit *et al.*, 2007; Kumar *et al.*, 2007; Li *et al.*, 2007; Wittlinger and Farra, 2007; Ozacar *et al.*, 2008; Savage and Silver, 2008; Snyder, 2008; Chen, 2009; Ford *et al.*, 2009; Hansen *et al.*, 2009; Kawakatsu *et al.*, 2009; Rychert and Shearer, 2009; Rychert *et al.*, 2009].

Here we present a continental-scale investigation of upper mantle discontinuity structure beneath North America, as imaged by *Ps* and *Sp* receiver functions, with a focus on estimating the thickness of the high velocity seismic lid, which we will refer to as the lithosphere. Our results indicate at least one negative velocity contrast exists at mantle depths beneath most of the continental United States. In the tectonically active western U.S. and portions of the eastern U.S., the largest negative *Sp* phase correlates with surface wave constraints on the LAB. However, in the central U.S. and some eastern

zones, the depth of the inferred discontinuity is much shallower than that estimated for the base of the lithosphere from surface and body wave tomography, suggesting the presence of layering within the lithosphere. In addition, the amplitude of negative Sp phases in the western U.S. is much larger than in the central and eastern U.S., indicating a stronger impedance contrast in the west, likely due to a weaker asthenosphere.

2. Data

Searching the USGS National Earthquake Information Center (NEIC) global event catalog for events with $M_w \geq 5.8$ and epicentral distance between 35° and 80° through the end of March 2009, we have compiled waveform data from the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (DMC) for 121 permanent seismic stations from 18 different networks in the continental United States, Canada, and Mexico (Fig. 1, Table 1). To avoid contamination with $pPPP$, $pPPPP$, and $sPPPP$ [Wilson *et al.*, 2006], we include only events with depths <300 km for Sp analysis. In addition, S waves at distances $<55^\circ$ are post-critical (i.e., vertical incidence angles are too large and no converted Sp phase will be produced) and at distances $>75^\circ$ SKS and $SKSp$ phases will potentially interfere with the Sp phase [Wilson *et al.*, 2006]. Therefore, we further restrict our Sp analysis to events from epicentral distances of 55° - 75° . Station operations times range from 1 to 20 years, starting as early as 1988, and the resulting data set contains more than 52,000 waveforms.

3. Receiver Function Methodology

A key feature of the receiver function approach is to deconvolve parent component waveforms (P for P_s and SV for S_p) from daughter components (SV for P_s and P for S_p) in order to normalize variations in earthquake source-time functions and to enhance the signal of energy converted at any impedance contrasts beneath a particular station. Although techniques for generating P_s and S_p receiver functions [Langston, 1977; Vinnik, 1977; Ammon, 1991; Bostock, 1998; Farra and Vinnik, 2000] have become relatively standard [e.g., Oreshin *et al.*, 2002; Vinnik *et al.*, 2004, 2005; Kumar *et al.*, 2005a, 2005b, 2006; Rychert *et al.*, 2005, 2007; Yuan *et al.*, 2006; Li *et al.*, 2007; Rychert and Shearer, 2009], there are several choices in the analysis process that can affect the resulting receiver functions. Furthermore, due to the large number of waveforms employed in this study (and even larger volumes likely to be used in the future application of our approach to data from the EarthScope Transportable Array) we have developed a semi-automated waveform analysis procedure. Therefore, we provide details on several key steps of our methodology in order to minimize ambiguity and to allow for our work to be more easily replicated.

3.1. Phase Picking

Direct P and S phase arrival times are determined with an automated picking algorithm based on the short-term-average (STA) to long-term-average (LTA) moving window method of Earle and Shearer [1994]. For a point in time (e.g., t_i) along the envelope function of a seismogram, the ratio of the mean of the amplitudes in a short window following t_i (i.e., a “signal” window or the STA) to the mean of the amplitudes in

a long window preceding t_i (i.e., a “noise” window or the LTA) is taken. The envelope is calculated by taking the absolute value of the Hilbert transform of the filtered waveform (0.5-1.5 Hz for P and 0.1-0.75 Hz for S ; note that a slightly different bandpass is used in the deconvolution – Section 3.4). We use a signal window length of 5 and 10 seconds for the P and S phases, respectively, with a noise window length four times that of the signal window (Fig. 2e,f). We calculate the signal-to-noise ratio (S2N) of the envelope function for a window ± 25 sec around the arrival time predicted by the 1-D reference Earth model AK135 [Kennett *et al.*, 1995] (Fig 2a,d). The S2N function is smoothed by convolving it with a Hanning window of the same length as the signal window (Fig. 2g,h). Here, the peak value in this smoothed S2N function is taken as the phase arrival time, rather than the inflection point preceding the peak used in Earle and Shearer [1994]. In cases where the maximum S2N value falls below a value of 2, we rely on the predicted phase arrival time from AK135.

3.2. Determining P and SV Components

To accurately recover the parent and daughter waveforms, the Z-N-E components are rotated into Z-R-T and then transformed into P-SV-SH using the free-surface transfer matrix [Kennett, 1991]

$$(1) \quad \begin{bmatrix} P \\ SV \\ SH \end{bmatrix} = \begin{bmatrix} \frac{(\beta^2 p^2 - 1/2)}{\alpha q_\alpha} & \frac{p\beta^2}{\alpha} & 0 \\ p\beta & \frac{(1/2 - \beta^2 p^2)}{\beta q_\beta} & 0 \\ 0 & 0 & \frac{1}{2} \end{bmatrix} \begin{bmatrix} Z \\ R \\ T \end{bmatrix}$$

where α and β are the near surface compressional and shear velocities, respectively, $q_\alpha = (\alpha^2 - p^2)^{0.5}$, $q_\beta = (\beta^2 - p^2)^{0.5}$, and p is the ray parameter of the direct phase. The free-surface velocities (α, β) are determined for each waveform at a particular station by finding the V_P (between 3.5 and 7 km/sec) and V_P/V_S (between 1.6 and 2.0) values that, when applied in equation (1), minimize the correlation between the P and SV components in a short window around the phase arrival. For P_S , we chose a window from $0.5T_P$ seconds before to T_P seconds after the P arrival, where T_P is the dominant P period. For S_P , the window starts at the S arrival and its length equals the dominant S period. We stack all constrained cross-correlation surfaces for both the P and S arrivals (i.e., those waveforms that have a well-defined minimum correlation in V_P - V_P/V_S space) and define the best free-surface velocities for a particular station as the minimum of this stack. We tested this approach with synthetic seismograms and found that our use of the free-surface transform accurately separates the P and SV components while also ensuring that phase incidence angles were consistent with a single set of free-surface V_P and V_P/V_S values that matched the model used to generate the synthetics. We also experimented with allowing the incident phase ray parameter to vary from its initial value, but this step had a negligible effect on results with real data. Note that our method differs from other studies [e.g., Oreshin *et al.*, 2002; Vinnik *et al.*, 2004; Kumar *et al.*, 2005; Li *et al.*, 2007] in which the P and SV components are assumed to be orthogonal and are determined by rotating the vertical and radial waveforms until the P component amplitude at the direct S arrival time sample is minimized.

3.3. Deconvolution

Because of the relatively small magnitude of scattered wave energy on individual waveforms (e.g., Fig. 2b,c), we enhance the signal of converted waves by combing all waveforms at a particular station. For Ps , the daughter component (SV) window includes 25 sec before and 75 sec after the direct P arrival pick (Fig. 2i); for Sp , the window is 75 sec before and 25 sec after the direct S arrival (Fig. 2j). The parent components are tapered to zero with a 0.5 sec cosine taper before the phase arrival pick. Waveforms are first bandpass filtered (0.03-1 Hz for Ps and 0.03-0.5 Hz for Sp) and then normalized by the maximum amplitude of the parent component and weighted according to the parent phase S2N.

We retrieve a response function (i.e., our receiver function) in time (RF_{time}) by deconvolving the parent phase from the daughter component for a range of possible conversion depths (z_c). We employ a simultaneous deconvolution in the frequency domain (following *Bostock* [1998])

$$(2) \quad RF_{time}(z_c) = F^{-1} \left(\frac{\sum_{i=1}^N DC_i(\omega) PC_i^*(\omega) e^{i\omega\tau_{pi}z_c}}{\sum_i PC_i(\omega) PC_i^*(\omega) + \delta} \right) / (2N\eta)$$

where PC_i and DC_i are the i^{th} waveforms (in the frequency-domain, ω) of the parent (P for Ps and SV for Sp) and daughter (SV for Ps and P for Sp) components, respectively; the * indicates the complex conjugate, F^{-1} is the inverse Fourier transform, N is the number of waveforms, and η is the length (in samples) of the waveforms. To correct for the different move-out of phases with different ray parameters, a time shift (τ) is applied to each waveform to align it with a reference ray parameter (p_{ref}); we choose a p_{ref}

corresponding to an epicentral distance of 60° for Ps and 65° for Sp . τ is a function of ray parameter (p_i) and conversion depth (z_c). δ is a regularization parameter (water level) and effectively acts as a low-pass filter. δ is chosen objectively such that the general cross validation function of the parent and daughter waveforms is minimized [Golub *et al.*, 1979]. In other words, we find the water level that, when applied in equation (2), produces a receiver function that when convolved with the parent component best-replicates the daughter component (see equations 4-6 in Bostock [1998]). We have also calculated receiver functions with an iterative time-domain deconvolution [Ligorria and Ammon, 1999] to determine if artifacts from spectral division in the frequency-domain deconvolution (i.e., side-lobes) contaminate the phases observed on the receiver functions presented in Section 4, and we find that such effects have negligible impact on the phases interpreted in this study.

The critical ray parameter for a converted phase changes with discontinuity depth such that Sp conversions at mantle depths are sometimes not predicted for smaller epicentral distances in the assumed mantle velocity models. Therefore, only waves with pre-critical ray parameters for a particular discontinuity depth are included in the deconvolution when calculating the RF_{time} for that particular depth. Pre-critical Ps conversions are predicted at all depths within the epicentral distance range investigated here (35° - 80°).

3.4. Migration

The receiver functions in time from equation (2) are mapped to depth by extracting amplitudes from each $RF_{time}(z_c)$ according to the predicted delay time (Ps - P or Sp - S)

versus depth function for a particular upper mantle velocity model for the reference ray. We have calculated and migrated receiver functions assuming two different velocity models. Crustal parameters in both models are either calculated by H- k stacking of P_s waveforms [Zhu and Kanamori, 2000] or taken from Crust2.0 [Bassin *et al.*, 2000; available online at <http://mahi.ucsd.edu/Gabi/crust2.html>] when the P_s receiver functions display too much complexity at crustal depths to permit simple interpretation of H- k stacks (Table 2). The first mantle model is simply the 1-D reference Earth model AK135 [Kennett *et al.*, 1995]. The second is a three-dimensionally varying model, in which 1-D velocity models for each station are constructed using mantle V_P from Burdick *et al.* [2008] and mantle V_S from Yuan *et al.* [in preparation]. The absolute V_P and V_S profiles for each station are azimuthal averages of the respective model at a distance of 125 km from station. This distance approximates S_p conversion point distances from the station for a discontinuity at 100 km depth.

The mantle V_S model used here is the isotropic part of the UC-Berkeley 3-D tomographic shear wave velocity structure of the North American continent [Yuan *et al.*, in preparation]. In the framework of normal mode asymptotic coupling theory [Li and Romanowicz, 1995], which includes finite frequency kernels, we simultaneously invert long period surface waveforms for isotropic S-velocity structure and a parameter that describes V_{SH} versus V_{SV} anisotropy (ξ) [Marone *et al.*, 2007]. The surface-wave dataset consists of over 150,000 3-component fundamental and higher mode waveforms recorded from available permanent and temporary broadband seismic stations across the continent, including the USArray Transportable Array and Permanent Backbone stations and the Canadian National Seismograph Network, which greatly improve the path and azimuth

coverage. The model is constructed starting from a new global 1D reference Earth model, based on a combination of seismic and mineral physics data [Cammarano and Romanowicz, 2007], which is smooth at 220 km, and a crustal correction approach that takes into account the non-linearities involved [Lekic *et al.*, 2009].

The choice of velocity model, in particular the V_P - V_S ratio, can have a impact on the depth to which the receiver function phases migrate, but we find very little difference in the receiver functions migrated with AK135 and those from the 3-D model. Although we present depths from both models in Table 2, for simplicity, all receiver functions shown in this paper were migrated with the 1-D station profiles extracted from the 3-D mantle velocity model.

3.5. Error Estimation

Errors are estimated for individual receiver functions by bootstrapping. For both P_s and S_p at each station, 20% of the waveforms are replaced with another random 20% of the waveforms, and the receiver functions are recalculated. This process is repeated 100 times and both the mean and standard deviation (σ) of the resulting set of receiver functions are calculated. We consider the well-resolved portions of the receiver functions to be those still significant at 2σ from the mean receiver function (i.e., the positive region below the mean- 2σ and the negative region above the mean+ 2σ). Errors in depth (amplitude) estimates are calculated from the maximum distance in depth (amplitude) from the pick to the mean $\pm 2\sigma$ (Fig. 3). Although the receiver functions we present here were generated by stacking all available data, we have also tried culling the data by both S2N and by the misfit between the direct phase arrival time predicted by AK135 and that

calculated from the S2N analysis. When excluding P_s and S_p waveforms with S2N less than 4 and 2, respectively, as well as phase arrival time discrepancies of more than 15 sec, we find no appreciable difference between the culled receiver functions and those calculated using all available data.

4. Results and Observations

Of the 121 stations in our data set (Fig. 1), 93 are found to have interpretable P_s (87) and/or S_p (73) receiver functions (Table 2, Suppl. Fig. S1, Suppl. Table S1), and we present these in several formats. First, five receiver function examples at individual stations from different tectonic regions are given to illustrate the phases that we interpret (Figs. 4-8), and results for all stations are summarized in Table 2. Second, to display the receiver function interpretations as a whole, the depths and amplitudes of the most prominent “negative” phase observed on the S_p receiver functions are plotted in map view (Fig. 9). The depths of the first and largest positive phase in the P_s receiver functions (the phase from the Moho) are shown in Supplemental Figure S2. As in most receiver function studies, we have reversed the amplitude of the S_p receiver functions to allow for a direct comparison with the P_s receiver functions; with this convention, a negative phase corresponds to a drop in velocity with depth. Lastly, four cross-sections through the study region demonstrate the variations of both the Moho and negative S_p phase depth estimates with respect to the S_p receiver functions (Fig. 10).

4.1. Rationale for Discontinuity Depth Estimates from Receiver Functions

Moho depths are inferred from the first major positive phase in the P_s receiver functions at reasonable depths (i.e., 20-60 km), but this phase is typically present in the S_p receiver functions as well (e.g., Fig. 4a-d). Crustal multiples from the P phase conversion at the Moho (e.g., $Ppps$, $Psps$, $Ppss$) will typically migrate to depths of 100-200 km (i.e., within the predicted LAB depth range for much of the central and eastern U.S.); crustal reverberations migrate to greater depth at larger epicentral distances (i.e., move-out) and are often prominent and easily distinguishable from direct conversions, which do not move out (e.g., Fig. 4b). These reverberations can mask other direct phases (e.g., the LAB), and to avoid this interference, potential “LAB” depths are picked from, in general, the largest negative phase in the S_p receiver function below the Moho phase. In making these depth estimates, we assess both the individual receiver functions (panels a and c in Figs. 4-8) and phase consistency across receiver functions binned by epicentral distance (panels b and d in Figs. 4-8). When multiple negative phases exist in the S_p receiver functions, we select the phase with the largest (most negative) amplitude, but emphasize that this choice of a single depth for inclusion in Table 2 obviously does not preclude the interpretation of the other phases as indicating significant upper mantle structure.

To guide our interpretation of the negative phases picked on the S_p receiver functions as either the LAB or a mid-lithospheric discontinuity (MLD), we compare the depth of these phases with absolute V_S in the 3-D shear velocity model used in the migration [Yuan *et al.*, in preparation]. We estimate the most likely LAB depth range from the V_S model to be from the minimum V_S below the Moho up to the next local peak in V_S (i.e.,

the transition from the most prominent low velocity zone, LVZ, to the overriding high velocity lid). In the western U.S. (e.g., station VTV, Fig. 4) and within the cratonic core of the continent (e.g., station ULM, Fig. 5) interpretation of the potential LAB depth range from the V_S model is straightforward. However, over a significant fraction of the central U.S. and the Phanerozoic eastern U.S., the absolute V_S profiles are fast with respect to the reference Earth model AK135 and contain two local minima, one typically near 100 km depth and a second near 200-250 km depth (Figs. 6e-8e). In these circumstances, potential LAB depth ranges are defined only when the true V_S minimum below the Moho is more than 0.5% slower relative to any other local minimum. For example, the two V_S minima at station HRV are 4.5633 km/s (100 km) and 4.5979 km/s (220 km), yielding a difference of 0.7570% (Fig. 6e). Where this criterion is not met, no potential LAB depth range is defined and the interpretation of the negative Sp phase is designated as “ambiguous”. This situation occurs, for example, at station FVM where the two V_S minima are 4.5973 km/s and 4.5920 km/s, or a difference of only 0.1166% (Fig. 7e) and at station SDMD where the difference in V_S between the two minima is 0.3467% (Fig. 8e).

Although the depth of the maximum negative velocity gradient has also been used to define the LAB (see *Eaton et al.* [2009] for a review of seismological definitions of the LAB), we find that in the model used here, the most negative dV_S/dz value in our estimated LAB depth range is often not the maximum negative dV_S/dz (the absolute minimum in dV_S/dz). However, for comparison, we illustrate both our estimated LAB depth range (gray shaded regions in Figures 4-6) and all minima in dV_S/dz points (gray circles in Figures 4-8; the larger circle denotes the absolute minimum in dV_S/dz).

4.2. Selected Receiver Function Examples

4.2.1. VTV (CI): San Andreas Fault (Southern California, U.S.)

Station VTV is located on the northern side of the southern San Andreas Fault near the southwestern edge of the Basin and Range (Fig. 1) where upper mantle velocities are considerably slower than beneath the central and eastern U.S. [e.g., *van der Lee and Nolet*, 1997; *Grand*, 2002; *Marone et al.*, 2007; *Burdick et al.*, 2008, *Nettles and Dziewonski*, 2008]. Crustal thickness estimates from H-*k* stacking and from the migrated *Ps* receiver functions (Fig. 4a) are 25 km and 25±2 km, respectively (Table 2), and are only slightly smaller than those from Crust2.0 [*Bassin et al.*, 2000] (28 km) and *Chulick and Mooney* [2002] (~30 km).

The *Sp* receiver function displays a single large positive Moho phase (at a depth similar to the *Ps* Moho) and a single large negative phase at 68±10 km (Fig. 4c). The depth of the negative *Sp* phase (horizontal magenta line) falls within the transition from the LVZ to the shallow higher velocity lid (gray region in Figure 4). Therefore, this phase is interpreted as the LAB. The inferred depth of the LAB at VTV is comparable to values found in the *Sp* receiver function study of *Li et al.* [2007]. The amplitude of the negative *Sp* phase at VTV, as well as at other stations in the western U.S. in general, is much larger than at most stations in the central and eastern U.S. The *Ps* receiver function does contain a negative phase at a similar depth as the *Sp* “LAB” (Fig. 4a,c), but this phase cannot be as unambiguously attributed to a direct phase given the proximity of crustal reverberations (Fig. 4b).

4.2.2. ULM (CN): Superior Craton (Manitoba, Canada)

Station ULM is located within the Archean Superior craton (Fig. 1) where seismic studies [e.g., *van der Lee and Nolet*, 1997; *Grand*, 2002; *Godey et al.*, 2004; *Ritsema et al.*, 2004; *Darbyshire et al.*, 2007; *Marone et al.*, 2007; *Burdick et al.*, 2008] indicate high velocities extend to nearly 250 km depth (i.e., cratonic North American lithosphere). The Moho phase is relatively sharp in both the P_s and S_p receiver functions (Fig. 5a,c), and clear crustal reverberations can be seen in P_s from 125-175 km (Fig. 5b). Crustal thickness estimates from H- k stacking and from the migrated P_s receiver functions (Fig. 5a) are 34 km and 36 ± 2 km, respectively (Table 2); these values are in good agreement with those from *Chulick and Mooney* [2002] (~ 35 km) and Crust2.0 [*Bassin et al.*, 2000] (39 km).

The S_p receiver function contains a broad region of negative energy with its maximum amplitude at a depth of 101 ± 14 km. In this example, the high velocity lithosphere is particularly well-defined in the V_S model, with the maximum negative dV_S/dz occurring at depths close to the minimum V_S (200-220 km, Fig. 5e). Because the negative S_p phase lies above the potential LAB depth range (gray region in Figure 5) we interpret this phase as a discontinuity within the lithosphere (i.e., a MLD). No significant S_p energy is found within the likely LAB depth range, suggesting that any velocity gradient at the base of the lithosphere is too gradual to produce significant converted phases. The P_s receiver functions binned by epicentral distance do contain negative energy at depths of 150-175 km, but because of a negative Moho reverberation at this depth interval, these signals are difficult to interpret.

4.2.3. HRV (IU): Passive Atlantic Continental Margin (Massachusetts, U.S.)

HRV is located in eastern Massachusetts along the passive Atlantic margin of North America and just to the east of a pronounced decrease in the thickness of the high velocity lid (from ~ 200 km to ~ 90 km) as imaged by surface-wave tomography [van der Lee, 2002; Li *et al.*, 2003]. At HRV, H-*k* stacking yields a crustal thickness of 31 km, and the peak of the Moho *Ps* phase migrates to a depth of 33 ± 2 km. These estimates of crustal thickness are consistent with previous analysis of *Ps* phases at this station [Li *et al.*, 2002; Ramesh *et al.*, 2002; Rychert *et al.*, 2005, 2007], as well as with those from Crust2.0 [Bassin *et al.*, 2000] (34 km) and Chulick and Mooney [2001] (~ 35 km).

The largest negative *Sp* phase at HRV migrates to a depth of 111 ± 7 km (Figure 6c). In the V_S model, the minimum near 100 km is sufficiently small relative to the minimum near 220 km (a difference of 0.787%) that a potential LAB depth range is definable. The negative *Sp* phase falls slightly below the V_S minimum, but close enough that it is interpreted as the LAB. Rychert *et al.* [2007] also observed a similar negative *Sp* phase at HRV, but the LAB depth inferred from *Sp* in this earlier work (~ 90 km for a V_P/V_S of 1.8 and 96-99 km for a V_P/V_S of 1.7) is slightly shallower than that found here (111 ± 7 km for an average V_P/V_S of 1.75). Although the receiver function methodology used here is nearly identical to that of Rychert *et al.* [2005, 2007], the number of waveforms included and the bandpass filter used are different, which may account for the differences in the receiver functions. The *Sp* receiver functions here (Fig. 6) were generated with 421 waveforms (Table 2), versus 56 waveforms used in Rychert *et al.* [2007], and the bandpass filter here is 0.03-0.5 Hz, as opposed to 0.01-0.175 Hz in Rychert *et al.* [2007]. In addition, the *Ps* receiver functions here employed 635 waveforms and a high

frequency corner in the bandpass filter of 1 Hz, versus 123 waveforms and 2 Hz in *Rychert et al.* [2005, 2007], with the result that a negative P_s phase at roughly 100 km is less clearly separated from Moho reverberations in this study.

By modeling P_s and Sp receiver functions, *Rychert et al.* [2005, 2007] found that a relatively sharp LAB discontinuity (5-6 km thick) with a negative velocity contrast of 5.3-7.4% is required beneath HRV. Given that the amplitude and width of the HRV Sp LAB phase obtained here do not differ dramatically from those in *Rychert et al.* [2005,2007], it is reasonable to infer that our data also indicate a similar velocity contrast and discontinuity sharpness.

4.2.4. FVM (NM): New Madrid Seismic Zone (Missouri, U.S.)

Although still characterized by relatively high velocities ($>3\%$) down to 200 km in some models [e.g. *van der Lee*, 2002; the 3-D V_S model used here], the New Madrid Seismic Zone and northern Mississippi Embayment are located near the margin of Proterozoic North America (what was Laurentia) where local seismic studies [*Zhang et al.*, 2009] and geochemical data [e.g., *Griffin et al.*, 2004 and references therein] suggest lower velocity lithospheric mantle may exist between 100 and 200 km. At FVM, as well as the other New Madrid (NM) seismic stations (Table 2), we use the crustal model from Crust2.0 [*Bassin et al.*, 2000] because the P_s receiver function at crustal depths contains both a strong shallow (<10 km) conversion and multiple phases between 40 and 60 km (Fig. 7a,b), and a reasonable H- k stacking result was not obtained. Nonetheless, the depth of the positive P_s receiver function phase (46 ± 3 km) interpreted as the Moho (also on the Sp receiver function) is in good agreement with the Crust2.0 Moho depth (42 km).

It is also comparable to the Moho depth of roughly 45 km at nearby station CCM obtained from H-*k* stacking in this study (Table 2) and by *French et al.* [2009].

The *Sp* receiver function shows a significant negative phase at 101 ± 11 km depth (Fig. 7c). Because the V_S minima at 100 km and 220 km are nearly the same (4.5973 km/s and 4.5920 km/s; 0.1166% difference), no potential LAB depth range was defined for this station. Therefore, the discontinuity that produced the negative *Sp* phase is not interpreted as the LAB or a MLD, and is designated as “ambiguous”. The shape of the absolute V_S profile at FVM is similar to V_S structure obtained in cratonic regions globally [Romanowicz, 2009] and the coincidence of the negative *Sp* phase with the V_S minimum near 100 km depth suggests the presence of a MLD. However, in order to maintain a consistent criterion in how the V_S profiles are interpreted (section 4.1) we retained the “ambiguous” designation.

4.2.5. SDMD (LD): Passive Atlantic Continental Margin (Maryland, U.S.)

SDMD is located in central Maryland along the passive Atlantic margin of North America. Crustal thickness at SDMD was determined by H-*k* stacking to be 30 km. Both the *Ps* and *Sp* receiver functions display a strong Moho phase at 32 ± 3 km, consistent with Crust2.0 [Bassin *et al.*, 2000] (34 km) and *Chulick and Mooney* [2001] (~ 30 km), and crustal reverberations are clear in the *Ps* distance bins (Fig. 8b).

We observe two negative phases in the *Sp* receiver function (70 and 105 km; Fig. 8c,d), but because the phase at 105 ± 9 km has a larger magnitude at two standard deviations it was chosen as the phase to interpret. Most other stations along the Atlantic margin to the northeast of SDMD show a similar pattern. The absolute V_S model has

three local minima, and although the V_S at 100 km is the minimum value, it is not more than 0.5% slower than the other local minima. The interpretation of the negative Sp phase at this station is therefore “ambiguous”, rather than either a MLD or the LAB.

4.3. Negative Sp Phase Depths

The depths of the largest negative Sp phase at each station are presented in map view as a smoothed function of Sp conversion point location (Fig. 9a). We first calculate theoretical Sp piercing points at the depth of the chosen negative Sp phase (Table 2) for each station assuming the migration velocity model profile at each station (e.g., Figs. 4e-8e) and then place each piercing point onto a grid of $0.05^\circ \times 0.05^\circ$ elements (discontinuity depths for each station are shown using the raw piercing points in Supplemental Figure S3). Piercing points were calculated only for paths with pre-critical ray parameters at the particular interpreted discontinuity depth, resulting in ~ 11000 Sp piercing points being used. Despite the relatively fine spacing of this grid (~ 5.5 km), piercing points from multiple stations often fall on the same grid element, and a single depth value is calculated for each element by taking the mean of all piercing points that hit it (weighted by the individual depth errors). This gridding and averaging process produces some smoothing, but we introduce additional smoothing by applying a circular filter with a radius of 6 points, which yields an effective grid spacing of 0.3° (~ 33 km), roughly the size of the first Fresnel zone for our phases at 70 km depth (assuming V_S equals 4.5 km/s and a period of 6.5 s).

Broadly speaking, the negative Sp phase depths are deeper beneath the central and eastern U.S. and shallower beneath the western U.S., although considerable variability

exists within each region. For example, phases in western central Nevada are significantly deeper (85-95 km) than the rest of the western U.S. (60-70 km), and southern Wisconsin manifests the shallowest phases in the central and eastern U.S. (59 km).

Although it may be tempting to conceptualize these points as a continuous surface, it is important to remember that the comparison of the receiver functions with absolute V_S leads to the interpretation of some negative Sp phases as the LAB and others as a MLD (black versus white stations in Figure 9), in addition to those which cannot be unambiguously interpreted (gray stations in Figure 9). Within the tectonically active western U.S., the discontinuity lies within the transition to a very slow asthenosphere, and corresponds to the LAB. At stations in Phanerozoic terranes to the south and east of the craton in Texas, Georgia, and most of the northeastern U.S., the discontinuity also appears to be the LAB. Within the cratonic regions of central North America (Fig. 1), the thick high velocity lid (see V_S contours in Figure 9b) and shallow Sp phases, lead to the interpretation of the discontinuity as a MLD. The V_S model in the region separating the LAB and MLD regions contains at least two low velocity zones of comparable magnitude, and therefore, the negative Sp phases there could arguably be attributed to either the LAB or an MLD, but we do not interpret this ambiguous region to represent a smooth transition (i.e., continuous discontinuity) between the LAB and MLD regions.

4.4. Sp Negative Phase Amplitude Variations

Using the same gridding and smoothing technique as with the negative Sp phase depths (Section 4.3), we also present the variation of negative Sp phase amplitude (Fig.

9b). Similar to the depth of the negative Sp phases (Table 2), the difference in amplitude between the receiver functions migrated with the 3-D and 1-D mantle models are virtually indistinguishable, and here only the amplitudes using the 3-D model are shown. Unlike the depths of the prominent negative Sp phase (Fig. 9a), the amplitudes show a clear distinction between larger negative amplitudes (red and green) in the western U.S. and smaller amplitudes (blue) in the central and eastern U.S. A rough correlation exists between depth (Fig. 9a) and amplitude (Fig. 9b); deeper phases tend to have smaller amplitude.

For stations where the negative Sp phase is interpreted as the LAB, the larger amplitude in the western U.S. implies that the velocity contrast between the lithospheric lid and asthenosphere is larger or occurs as a more rapid velocity gradient than in the eastern U.S. The absolute V_S model contains much slower velocities at asthenospheric depths in the western U.S. than in the eastern U.S. (see for example the velocity anomaly contours for a depth of 140 km in Figure 9b, or contrast the VTV and HRV V_S profiles in Figures 4 and 6). This result suggests that the apparently larger LAB velocity contrasts in the western U.S. are due to a much slower asthenosphere.

4.5. Selected Cross-Sections

Presentation of all receiver functions in the format of Figures 4-8 is not practical, but we show four cross-sections (Fig. 10) that illustrate more than half of our picked Sp receiver functions relative to the Moho and negative Sp phase depths (cyan and magenta lines, respectively) and the inferred LAB depth from the V_S model (gray regions). For clarity and in contrast to the examples in Figures 4-8, Figure 10 only shows receiver

function amplitudes within the two standard deviation uncertainties from the bootstrap testing (i.e., only resolved positive (blue) and negative (red) amplitudes are shown).

4.5.1. Cross-Section A-B (ULM-LRAL)

The first cross-section (A-B, Fig. 10a) runs south from station ULM within the Superior craton in southern Manitoba, Canada through the New Madrid Seismic Zone and to station LRAL in Alabama. The Moho (positive P_s phase) shows a slight increase in depth from ~ 40 km near ULM to ~ 50 km near LRAL. The negative Sp phase decreases in depth from 101 ± 14 km at ULM to 59 ± 12 km beneath JFWS before increasing again towards LRAL (91 ± 21 km). This shallowing would be less prominent if the deeper (and smaller) phase observed at JFWS had been picked as the primary arrival. In any case, the depth of the Sp phase at all of these stations lies well above the potential LAB depth ranges from the V_S model where they exist (gray regions in Figure 10a). The phases are interpreted as a MLD, except at SLM, SIUC and UTMT which fall into the “ambiguous” category. Overall, very little converted energy is seen at depths greater than 100 km.

4.5.2. Cross-Section C-D (SCZ-CEH)

Cross-section C-D (Fig. 10b) spans the continental U.S. from central California to North Carolina. The crust is relatively shallow beneath SCZ (25-30 km) and gradually thickens to ~ 55 km across the Basin and Range, through the Great Plains, and into the southern Appalachians before thinning again to ~ 35 km in eastern North Carolina. Although our resolution is lower, these values agree well with those presented in *Chulick*

and Mooney [2002]. In contrast with the Moho, the negative Sp phases show a moderate amount of variability, ranging between 70 and 100 km. West of Colorado (station ISCO), amplitudes of the negative Sp phases are larger and their depths agree with the likely LAB range from absolute V_S (gray regions). The LAB appears to deepen in the western Basin and Range at stations MNV and TPH (also see Fig. 9a), but the easterly extent of this apparently thicker lithosphere is not well-constrained because no piercing points exist east of TPH until ISCO. East of WCI, the V_S model indicates a thicker high velocity lid (~ 200 km) but the negative Sp phases are at depths of ~ 70 -100 km, resulting in their interpretation as a MLD. No eastern stations in this profile have significant negative Sp energy within the highlighted LVZ-to-lid transition region.

4.5.3. Cross-Section E-F (LTX-WVL)

Cross-section E-F (Fig. 10c) crosses from Phanerozoic lithosphere in Texas, through the Proterozoic terranes in the central U.S., to the Paleozoic Appalachians in the northeastern U.S. The negative Sp phases show significant variability but no significant negative Sp energy exists between 150 and 250 km across the line. The V_S model indicates a thin lithosphere (50-100 km) at either end of this line and a thick lithosphere (150-200 km) near its center at BLO and MCWV (gray regions), but the LAB depth is unclear in both the New Madrid Seismic Zone (stations UALR, UTMT, and SIUC) and the central Appalachians (stations SSPA and BINY). Therefore, the negative Sp phases are interpreted as the LAB near the Atlantic margin and as a MLD in a portion of the Proterozoic craton, but are not interpreted in between.

4.5.4. Cross-Section G-H (GNW-NE71)

The last cross-section (G-H, Fig. 10d) differs from the first three in that it follows the tectonically and volcanically active western margin of the U.S. (i.e., Cascadia, the eastern Sierras, and southern California). Along this profile, the negative Sp phases fall within the potential LAB depth range from the V_S model and are interpreted as the LAB, with the caveat that it may be difficult to distinguish the LAB from discontinuities associated with the Juan de Fuca slab in the Pacific Northwest (stations GNW, LTY, PIN, YBH). An obvious feature of these receiver functions is the large amplitude of the negative Sp phases, indicating a larger velocity contrast at the LAB than is typical of the eastern U.S. In addition, more energy appears at greater depths than in other profiles. These deeper phases may represent real structure and are not unexpected given the likelihood of vertical heterogeneity associated with recent and current subduction of the Juan de Fuca/Farallon Plate and processes such as lithospheric delamination beneath the Sierras [e.g., Zandt *et al.*, 2004]. Li *et al.* [2007] also investigated this same general region with Sp receiver functions, and our results here do not differ significantly from theirs; the only major discrepancy appears to be near TPH and MNV, where they image a shallower LAB (60 km versus 80-83 km obtained here; see cross-section C-D, Fig. 10b).

5. Discussion

A growing body of studies that employ Sp and Ps receiver functions have found evidence for discontinuities characterized by a significant negative velocity contrast at depths comparable to those documented here. These studies span oceanic regions [Li *et*

al., 2000; *Collins et al.*, 2002; *Li et al.*, 2004; *Kumar et al.*, 2005b; *Vinnik et al.*, 2005; *Wolbern et al.*, 2006; *Heit et al.*, 2007; *Kumar et al.*, 2007; *Kawakatsu et al.*, 2009; *Rychert and Shearer*, 2009], regions of relatively thin (<120 km) continental lithosphere [*Oreshin et al.*, 2002; *Kumar et al.*, 2005a; *Rychert et al.*, 2005, 2007; *Chen et al.*, 2006; *Mohsen et al.*, 2006; *Sodoudi et al.*, 2006a,b; *Hansen et al.*, 2007; *Heit et al.*, 2007; *Kumar et al.*, 2007; *Li et al.*, 2007; *Ozacar et al.*, 2008; *Rychert and Shearer*, 2009; *Chen*, 2009], and continental regions with thicker cratonic-style lithosphere [*Bostock*, 1998; *Ford et al.*, 2009; *Rychert and Shearer*, 2009]. A widely distributed layer of low velocities at roughly 100 km depth beneath continents has also been argued for on the basis of long-range active source profiles [e.g. *Thybo and Perchuc*, 1997; *Thybo*, 2006] and absolute V_S models from global surface-wave tomography [*Romanowicz*, 2009]. However, in some cratonic regions deeper discontinuities (>150 km) have also been proposed by Sp and Ps studies [*Sacks et al.*, 1979; *Kumar et al.*, 2007; *Wittlinger and Farra*, 2007; *Savage and Silver*, 2008; *Snyder*, 2008; *Hansen et al.*, 2009].

Here we discuss possible physical and chemical mechanisms that could produce a negative velocity contrast and be responsible for the discontinuity that we have imaged beneath the continental U.S. and a small portion of northwest Mexico and southeast Canada. We first explore the origin of this feature in regions where we have interpreted it as the LAB and then in regions of thicker lithosphere where we have interpreted it as a MLD (Table 2).

5.1. Negative Velocity Contrast at the Lithosphere-Asthenosphere Boundary

Where interpreted as the LAB, the depth of the discontinuity imaged in this study varies from 51 km to 111 km (black stations in Figure 9, Table 2). In addition, the amplitude of the *Sp* phase marking the discontinuity is significantly larger in the tectonically active western U.S. than within the Phanerozoic terranes along the southern and eastern margins of Proterozoic North America (Fig. 9b), a result that correlates with lower asthenospheric shear velocities in the west.

Some studies conclude that temperature, or a combination of temperature and grain size, can largely explain the transition from fast lithosphere to slow asthenosphere [e.g. *Faul and Jackson*, 2005; *Stixrude and Lithgow-Bertelloni*, 2005; *Priestley and McKenzie*, 2006]. Other studies emphasize the faster velocities expected in dry, depleted lithosphere, versus more hydrated and fertile asthenosphere [*Hirth and Kohlstedt*, 1996; *Karato and Jung*, 1998; *Gaherty et al.*, 1999]; however, the amount of shear velocity drop that can be attributed to partial melting-induced depletion is the subject of debate [*Lee*, 2003; *Schutt and Lesher*, 2006]. The presence of a small amount of partial melt in the asthenosphere could dramatically reduce its shear velocity, although the magnitude of the effect strongly depends on the melt distribution geometry [e.g. *Gribb and Cooper*, 2000; *Hammond and Humphreys*, 2000; *Takei*, 2002; *Faul et al.*, 2004; *Kawakatsu et al.*, 2009]. Rapid variations in anisotropy with depth may also create discontinuities [e.g. *Bostock*, 1998; *Gaherty et al.*, 1999; *Levin and Park*, 2000; *Gung et al.*, 2003]. Considering these possibilities, what can we postulate about the source of the negative *Sp* phase where we interpret it on our receiver functions as the LAB?

The amplitudes of Sp and Ps phases are largely controlled by the gradient in shear velocity at the discontinuity where they are generated [e.g. *Rychert et al.*, 2007]. To accurately bound the shear velocity gradient, detailed waveform modeling is required [e.g. *Collins et al.*, 2002; *Rychert et al.*, 2005, 2007; *Chen et al.*, 2006; *Li et al.*, 2007; *Kawakatsu et al.*, 2009], and such an effort with the data gathered in this study will be the focus of a future investigation. Nonetheless, existing modeling provides some guidelines that we can apply to the results shown here. For example, even when accounting for anelastic (viscoelastic relaxation or attenuation) effects on shear wave velocity [*Faul and Jackson*, 2005], *Rychert et al.* [2005, 2007] found that the thermal gradient required at the LAB to match their Ps data in the northeastern U.S. (at station HRV; found here in Figure 6 and on cross-section E-F in Figure 10c) is larger than that expected at the base of the lithosphere from geodynamical models in which viscosity depends solely on temperature and pressure [*King and Ritsema*, 2000; *Zarnek et al.*, 2005]. *Chen et al.* [2006] arrived at a similar conclusion from their modeling of Ps receiver function data in eastern China. In both cases [*Rychert et al.*, 2005, 2007; *Chen et al.*, 2006], the additional velocity reduction was explained by the presence of water or melt in the asthenosphere, possibly in combination with greater depletion in the lithosphere or (in the case of *Rychert et al.* [2007]) a small contribution from depth variations in anisotropy.

The amplitude and sharpness of the Sp receiver functions we observe at HRV (Fig. 6) and other eastern seaboard stations are similar to those from *Rychert et al.* [2007] at HRV. In addition, the dominant periods of the S parent phases used in our study are smaller (6.5 ± 1.5 s) than those produced by the filtering employed by *Rychert et al.* [2007] (roughly 11 s), so in the case of gradational boundaries, a stronger velocity

contrast would be required with our data to produce a phase of comparable amplitude between the two studies. Therefore, we speculate that the negative *Sp* phase interpreted as the LAB east of the Cordillera (stations BRNJ, GOGA, HKT, HRV, JCT, LBNH, MVL, PAL, and WVL; see Table 2, Fig. 9, and Suppl. Fig. S1) may require the presence of a small amount of partial melt and/or water in the asthenosphere in addition to thermal gradients.

Could melt be present beneath the continental lithosphere in the eastern and southern U.S.? When *Mierdel et al.* [2007] jointly considered water solubility in Al-saturated enstatite and olivine, they showed that a minimum in water solubility occurs at depths of 100 km (for a typical old oceanic geotherm) to 150 km (for relatively cold continental geotherms), leading them to infer that melting would occur beneath the continental lithosphere in the presence of a few hundred parts per million of water. *Till et al.* (2009) concluded that small percentages of partial melt could exist in the asthenosphere beneath eastern North America (1-2% melting for an asthenosphere with a mantle potential temperature of 1400°C and 150-300 ppm H₂O, or a potential temperature of 1350°C with 450-900 ppm H₂O). In these models, the LAB is inferred to be a “damp” solidus, and therefore may correspond to a permeability barrier that would trap melt within the asthenosphere. In addition, melt that was able to migrate above the LAB would crystallize relatively quickly, assuming melt bodies with reasonable dimensions [*Till et al.*, 2009].

In contrast to eastern U.S. stations, the negative *Sp* phases at western U.S. stations display larger amplitudes (Fig. 9b), likely due to much lower asthenospheric velocities, as seen in the V_S model and other seismological studies [e.g., *Grand*, 1994; *Humphreys and*

Dueker, 1994; van der Lee and Nolet, 1997; van der Lee, 2002; Godey et al., 2004; Marone et al., 2007; Xue and Allen, 2007; Li et al., 2008; Nettles and Dziewonski, 2008; Roth et al., 2008]. Other geophysical and geological observations in the western U.S., such as high heat flow [*Lachenbruch, 1978; Morgan and Gosnald, 1989; Blackwell et al., 1991*], Basin and Range extension [*McKenzie and Bickle, 1988; Wernicke et al., 1988; Bennett et al., 1998*], dynamic topography [*Lowry et al., 2000*], and a plethora of recent and active volcanism (e.g., the Cascade arc, High Lava Plains, Yellowstone hotspot track, Long Valley Caldera), indicate the shallow upper mantle is anomalously hot and, at least in certain regions, likely partially molten. Owing to this well-established view of the western U.S., it is reasonable to interpret the larger amplitude of negative S_p phases in this region to a high temperature, partially molten asthenosphere underlying a relatively melt-free lithosphere. However, very strong LAB temperature gradients due to small-scale convection, plus the effects of asthenospheric water, cannot be ruled out. These interpretations require testing with detailed modeling of the LAB velocity gradient and focused geodynamical studies.

5.2. Absence of a Sharp LAB Beneath Thick Lithosphere

A key result of this study is the apparent absence of a sharp gradient in shear velocity beneath all regions of the North American continent where absolute V_S structure indicates a seismic lid of more than 130 km in thickness. These include the sampled Archean and Proterozoic craton, as well as portions of the Phanerozoic Appalachian orogen in the eastern U.S. (Figs. 1 and 9). This finding contrasts with other cratonic regions where S_p receiver functions have been used to infer an observable LAB discontinuity at depths of

more than 160 km [Sacks *et al.*, 1979; Kumar *et al.*, 2007; Wittlinger and Farra, 2007; Hansen *et al.*, 2009]. Elsewhere in North America, a sharp LAB has been invoked beneath the Slave Craton [Snyder *et al.*, 2008], although this study employed P_s , rather than S_p , receiver functions.

On its own, the absence of significant S_p phases from the regions of thick lithosphere sampled in this study implies that the velocity drop at the LAB is very weak and/or gradual in depth. However, the V_S model at most cratonic stations indicates a significant drop in velocity from the fast lid to the asthenosphere. For example, at ULM, this velocity drop is 4% (Fig. 5e). Therefore, the lack of S_p energy from the LAB likely represents a gradual velocity gradient.

How gradual would this gradient have to be? Assuming that the velocity drop occurs as a step function in depth (a gradient with zero thickness) and that the amplitude of the resulting S_p phase is A, then if the same drop in velocity occurs over a depth range of H, where the wavelength of the seismic phase at the discontinuity divided by H equals 2, the amplitude will be 50-70% of A [Rychert *et al.*, 2007, 2009]. (Figure 7 of Rychert *et al.* [2007] contains a plotting error. A correct version of this figure can be found in Rychert *et al.* [2009]) For a 4% velocity drop, this gradient thickness should be observable. If the wavelength of the phase at the discontinuity equals H, the amplitude will be less than 30-35% of A. Gradient thicknesses that exceed this threshold would almost certainly not be observable for a 4% velocity drop, given typical noise levels in data. For the dominant S periods in our filtered waveforms, the typical wavelength at the base of the high V_S lid is roughly 30 km. Therefore, if we assume that the S_p phase becomes unobservable at 30-35% of A, then the gradient must occur over more than 30 km. We therefore conclude

that the shear velocity gradient beneath the thick North American lithosphere occurs over 30 km or more. The relatively gradual temperature gradients predicted by geodynamical models where viscosity depends only on temperature and pressure [King and Ritsema, 2000; Zaranek *et al.*, 2005] are therefore a viable option in these regions of the continent.

5.3. Negative Velocity Contrast Within the Lithosphere

Given that significant negative discontinuities appear to exist at depths of 59-113 km in the lithosphere of cratonic North America and certain portions of the Phanerozoic eastern U.S., what are the possible sources of the velocity contrasts accompanying this layering? Here we evaluate the mechanisms presented in the previous section and some additional possibilities, in light of likely geotherms for the region. Moreover, because a discontinuity corresponding to the top of a low velocity layer internal to the cratonic lithosphere has been proposed in other cratonic regions [Savage and Silver, 2008; Ford *et al.*, 2009] as well as in cratons globally [Thybo and Perchuc, 1997; Thybo, 2006; Romanowicz, 2009; Rychert and Shearer, 2009; Rychert *et al.*, 2009], the conclusions drawn here have implications beyond North America.

Thybo [2006] argues that a lowering of the peridotite solidus by the presence water and/or carbon dioxide is likely responsible for a mid-lithospheric LVZ at depths of roughly 100 km through the effects of partial melt (if temperatures are above the solidus) or by a decrease in shear modulus and increase in attenuation at sub-solidus temperatures [Sato *et al.*, 1989a,b]. Within the cratonic regions sampled in our study, the geotherms preferred by Griffin *et al.* [2004] based on peridotitic garnet xenocrysts imply temperatures below 800°C at depths of 100 km. These values are below even the lowest

temperature estimates for the water-saturated solidus [e.g., *Katz et al.*, 2003; *Grove et al.*, 2006], making the present-day existence of melt unlikely. Allowing for warmer cratonic geotherms at 100 km, for example the higher temperature end of the 750-900°C range inferred for the craton from seismic velocities [*Goes and van der Lee*, 2002] or 900-1100°C from heat flow [*Artemieva and Mooney*, 2001], melt or near-solidus conditions might exist in the presence of water contents approaching saturation. In addition, the highest of these values (1100°C) just reaches the solidus for natural carbonated peridotite at 100 km [*Dasgupta and Hirschmann*, 2006]. Hydration of the North American lithosphere by flat-slab subduction of the Farallon Plate in the early Cenozoic has been proposed to explain widespread uplift and volcanism in the western U.S. [*Humphreys*, 1995; *Humphreys et al.*, 2003], and the Farallon slab has been proposed as a source of volatile-enrichment in the central and eastern U.S. asthenosphere [*van der Lee et al.*, 2008]. However, these scenarios raise the question of how volatiles would have become highly concentrated in a layer whose top lies at 70-115 km, and how this layer was maintained over time; one possibility is the rapid increase in olivine and Al-saturated enstatite water solubility at roughly 100 km predicted for a cold, cratonic geotherm [*Mierdel et al.*, 2007]. In addition, recent modeling of surface motions in response to glacial isostatic adjustment [*Peltier and Drummond*, 2008] favors a lower-lithospheric layer of reduced viscosity (10^{22} Pa s) from 60-100 km depth that relaxes a large fraction of horizontal stress and reduces misfits to observed horizontal motions. This type of weak zone could be taken as evidence for a layer of melt or highly hydrated mantle.

Another reasonable scenario is that the *Sp* discontinuity represents the sharp top of a now-solid melt cumulate layer, whose lower boundary is more gradual, and therefore

does not produce its own observable phase. At depths comparable to the observed MLD, the cumulate could be pyroxenite, whose velocity would be sufficiently slow with respect to peridotite [Cin-Ty Lee, *pers. comm.*; *Stixrude and Lithgow-Bertelloni*, 2005; *Behn and Kelemen*, 2006]. This melt layer could have been created at an earlier time when the cratonic lithosphere was warmer, allowing water-rich melts from the asthenosphere to percolate up through the lithosphere and collect in a layer whose top would have been the wet pyroxenite solidus. To explain the 4.5% drop in velocity they image at 150 km depth in the Kalahari craton, *Savage and Silver* [2008] propose that infiltration of basaltic melt and metasomatic fluids refertilized and reduced velocity in the originally depleted mantle of the cratonic lithosphere, and that the sharp top of this slow layer corresponds to the basalt liquidus at the time the melts were present. While a mantle refertilization process could have played a role in creating slow velocities internal to the North American lithosphere, the basalt liquidus [*Yasuda et al.*, 1994; *Medard and Grove*, 2006] would likely have been too hot relative to a cratonic geotherm at MLD depths to provide a general mechanism for the MLD observed in this study.

Anisotropic fabric is another plausible mechanism for generating a velocity contrast within the lithosphere. A change in anisotropy, either between two regions with different orientations of anisotropy or between an isotropic and an anisotropic region, should be observable with receiver functions [e.g., *McNamara and Owens*, 1993; *Levin and Park*, 1997; *Nagaya et al.*, 2008] as variations in phase amplitude timing with back-azimuth and including the presence of converted SH phases. Although rigorous analysis of evidence for anisotropy in the receiver functions presented in this study is reserved for future work, a number of teleseismic converted phase studies have argued for anisotropic

layering within thick cratonic lithosphere [Bostock, 1998; Levin and Park, 2000; Saul *et al.*, 2000; Wittlinger and Farra, 2007; Mercier *et al.*, 2008; Snyder, 2008], sometimes associating discontinuities in the 70-115 km depth range with the Hales discontinuity [Bostock, 1998; Levin and Park, 2000; Snyder, 2008]. The original definition of the Hales discontinuity involved an increase in isotropic velocity with depth [Hales, 1969], as opposed to the decreases in velocity at the MLD found in this study. However, if a discontinuity involves anisotropy, the apparent sign of the discontinuity may vary as a function of sampling with back-azimuth.

In the Canadian Shield, north of our study region, *Ps* receiver function imaging, along with xenolith and reflection data, has been used as evidence for significant lithospheric layering that resulted from stacking of oceanic plates during paleo-Proterozoic subduction [Cook *et al.*, 1997; Bostock, 1998; Mercier *et al.*, 2008; Snyder, 2008]. In particular, a thin (10 km) anisotropic layer inferred to be the top of a subducted slab was observed at depths of 30-90 km beneath the Wopmay orogen and the Slave craton [Bostock, 1998; Mercier *et al.*, 2008], overlapping the range of MLD depths found in this study (59-113 km), and deeper anisotropic discontinuities were found further to the west beneath the Slave craton [Snyder, 2008]. Anisotropic layering has also been documented in the lithosphere of the Arabian Shield at 70 km depth [Levin and Park, 2000] and the Indian Shield at 90 km depth [Saul *et al.*, 2000]. However, the existence of such anisotropic layering within the Proterozoic mobile belts of the central and eastern U.S. has not been established by receiver function studies.

The spinel-garnet transition was also evaluated as a possible cause of the MLD observed in our data at 60-110 km. However, while this phase boundary occurs at

approximately the correct depth, and the deeper phase (garnet) does have a slower velocity, the relatively small percent of spinel in a peridotite mantle lithosphere would make the velocity contrast associated with this transition too small to explain the observed MLD [Greg Hirth, *pers. comm.*; *Stixrude and Lithgow-Bertelloni*, 2005].

6. Conclusions

Recent studies propose that a global discontinuity between 80 and 120 km with a negative velocity contrast exists [e.g., *Thybo*, 2006; *Romanowicz*, 2009; *Rychert and Shearer*, 2009]. The *Sp* receiver functions presented here also reveal at least one discontinuity with a negative velocity gradient beneath most of the continental U.S. (and portions of Mexico and Canada) at a similar depth (Fig. 9). However, comparison of this discontinuity to absolute shear-wave velocities leads to its interpretation as the LAB in the western U.S. and portions of the Phanerozoic eastern U.S. and as a MLD in the cratonic core of the continent and certain Phanerozoic regions.

In the tectonically active western U.S. where high heat flow and a shallow, very slow asthenosphere are observed, it is reasonable to infer that the negative *Sp* phases observed at 51 to 104 km (Figs. 4,9,10) do indeed represent the LAB. The large amplitude of the inferred LAB discontinuity in the western U.S. points to the need for an additional mechanism beyond elevated temperatures in the asthenosphere, such as asthenospheric partial melt, unless localized mantle upwelling produces anomalously sharp vertical temperature variations. In the portions of the eastern U.S. where the *Sp* phases are interpreted as the LAB (Figs. 6,9,10), some contribution from volatiles or partial melt in

the asthenosphere may be required. Although the eastern U.S. S_p phases are generally weaker than in the western U.S, numerical models indicate that eastern U.S. thermal gradients will be too gradual to explain the observed receiver functions, unless another factor, such as volatiles or partial melt, sharpens the velocity contrast between lithosphere and asthenosphere. In contrast, the relatively thick (~ 200 km) high velocity lid imaged by surface and body wave tomography in cratonic North America and sections of the eastern U.S. leads us to interpret the negative S_p phases at 59-113km as a MLD, perhaps caused by elevated volatile contents at near solidus temperatures within the lithosphere, the top of a melt cumulate layer, or by anisotropic layering. The absence of a strong negative S_p phase at the base of the thicker North American lithosphere implies that the LAB there is relatively broad (>30 km) and may be purely thermal in origin. Overall, these results indicate that the LAB is much more gradual in regions where the lithosphere is thicker than 150 km, compared to the sharper LAB velocity gradients observed in regions of thinner (<120 km) lithosphere.

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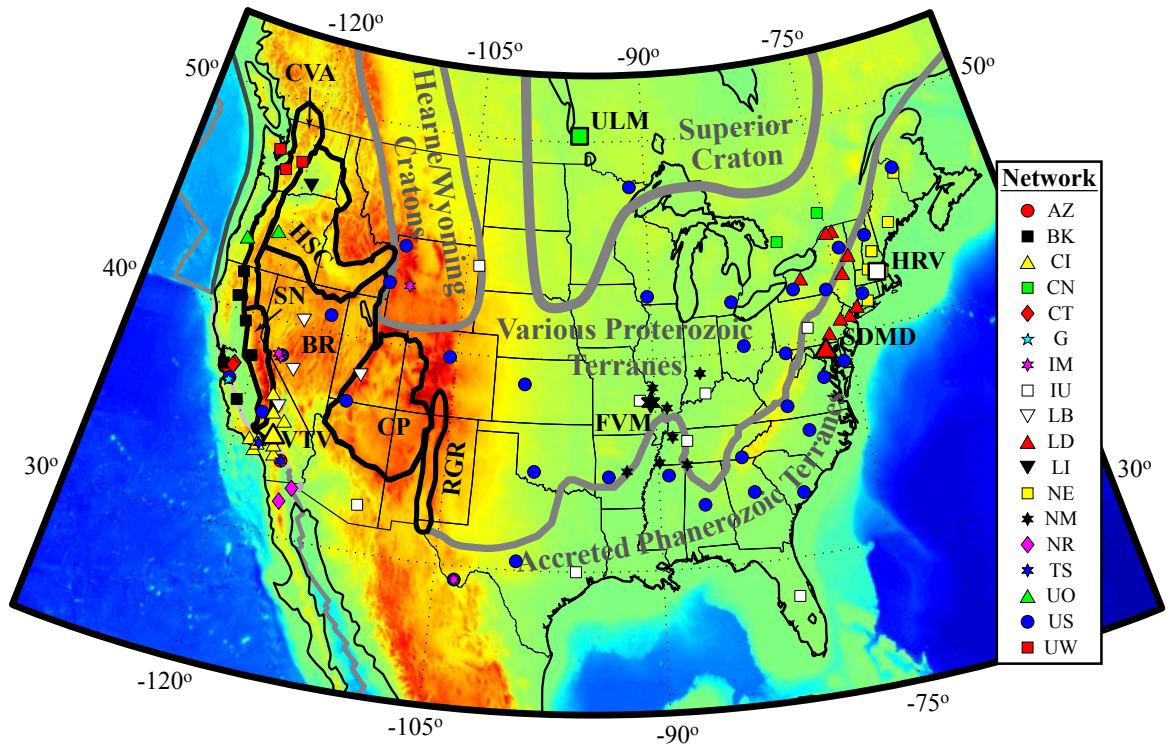


Figure 1. Permanent seismic stations used in this study, topography, and general tectonic setting. The stations are distinguished by network, and the five stations presented as detailed examples in Section 4.2 (VTV, ULM, HRV, FVM, and SDMD) are highlighted. See Table 1 for network information and Supplemental Figure S1 for all station locations. Major boundaries between Archean, Proterozoic, and Phanerozoic lithotectonic terranes are shown with thick gray lines (drawn very schematically for geographic reference only and not detailed interpretations [from *Rivers, 1997; Chulick and Mooney, 2002, Griffin et al., 2004*]), and the current approximate extents of the Rio Grande Rift (RGR), Colorado Plateau (CP), Basin and Range (BR), High Lava Plains/Snake River Plain/Columbia Plateau (HSC), Cascades volcanic arc (CA), and Sierra Nevada batholith (SN) are shown with thick black lines [Freeman and Johnson, 1946]. Topography and bathymetry are from *Smith and Sandwell [1997]*.

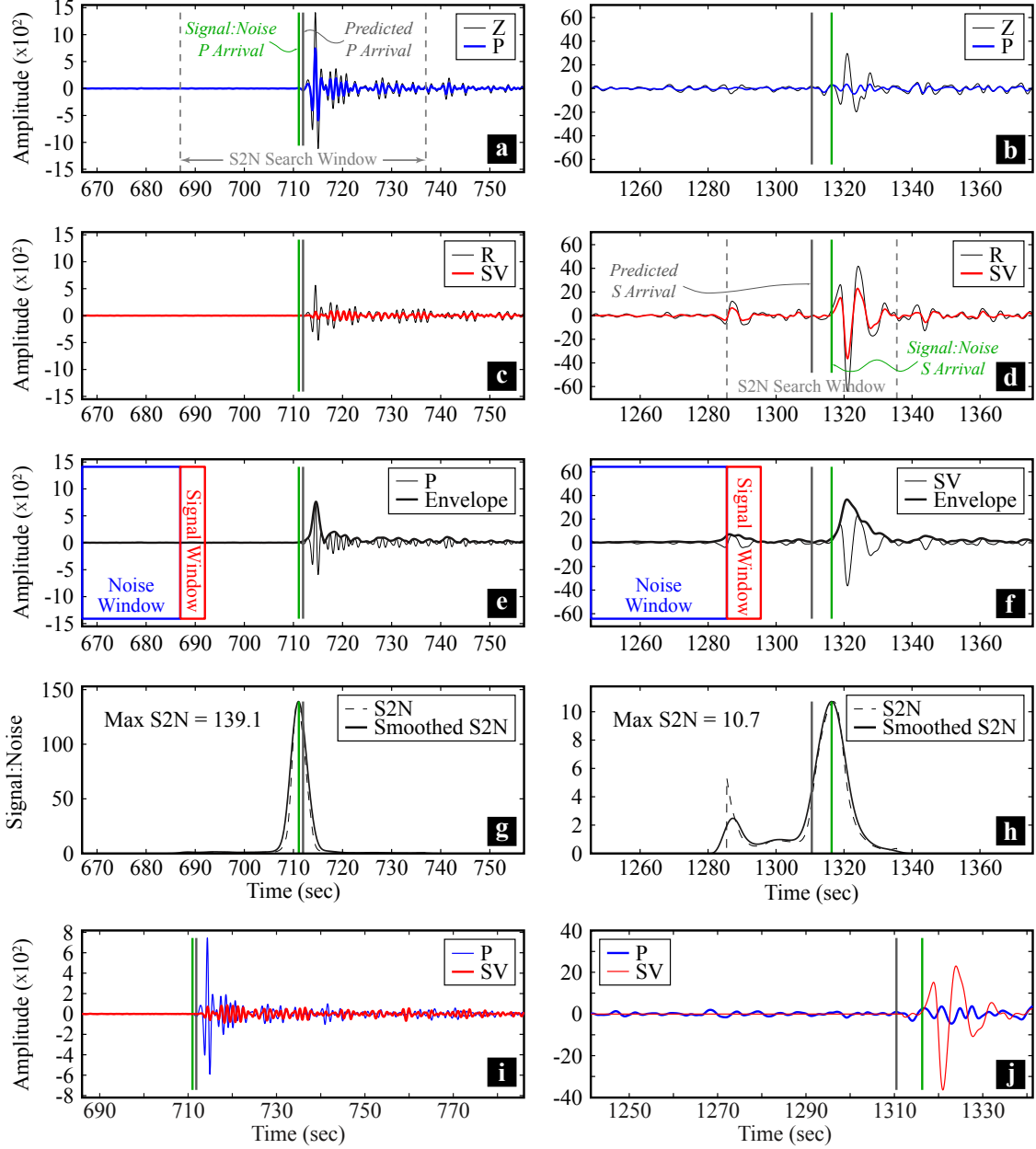


Figure 2. Example of the coordinate system rotation and phase picking routine employed in the pre-processing of waveform data. The left (right) column shows windows around the P (S) arrival time. For this signal:noise (S2N) analysis, waveforms for P s (left column) are bandpass filtered at 0.5-1.5 Hz (0.1-0.75 Hz for S p, right column). The Z and R components (thin black lines in a-d) are transformed into P and SV (thick blue and red lines in a-d, respectively) with the free-surface transform [Kennett, 1991; see Section 3.2]. The large reduction in amplitude on the transformed daughter components (SV for P s and P for S p) relative to the Z-R components is easily observed. The vertical gray line in each panel shows the predicted phase arrival time from the 1-D reference Earth model AK135 [Kennett et al., 1995]. Panels e and f display the envelope of each parent waveform and the signal and noise windows used. The window of time for which S2N is calculated is shown in panels a and d. The resulting S2N functions are in panels g and h, and the peak in S2N is indicated by the vertical green line, which also appears in all other panels. Although the predicted AK135 and calculated S2N arrival times are similar for P , a more accurate pick of the S arrival is made by the S2N analysis. The full windows used in the deconvolution are shown in panels i and j; note the slight difference in time scale in panels a-h.

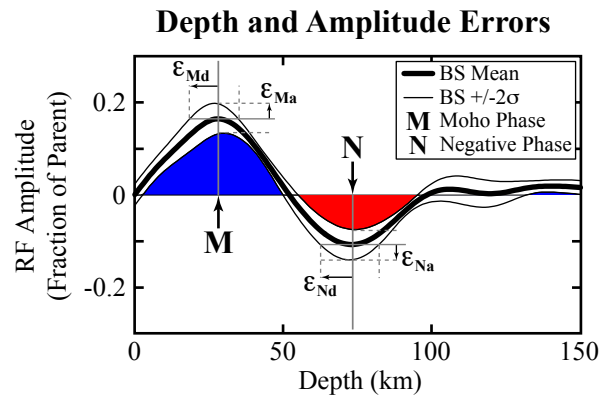


Figure 3. Illustration of error estimates made on the receiver functions. The thick black line is the mean receiver function from the bootstrap testing. The thin black lines on either side of the mean are ± 2 standard deviations (σ) from the bootstrap testing. For the Moho (M) phase pick, the error in depth (ϵ_{Md}) is taken as the maximum distance from the pick on the mean receiver function to the $+2\sigma$ line; for the negative phase pick (N), the -2σ line is used to measure the error (ϵ_{Nd}). Error in amplitude (ϵ_{Ma} or ϵ_{Na}) is simply the 2σ value at the depth of the pick. Although we illustrate these error estimates on the same receiver function here, the values found in Table 2 have been calculated from either Ps for M or Sp for N.

VTV (CI) - San Andreas Fault

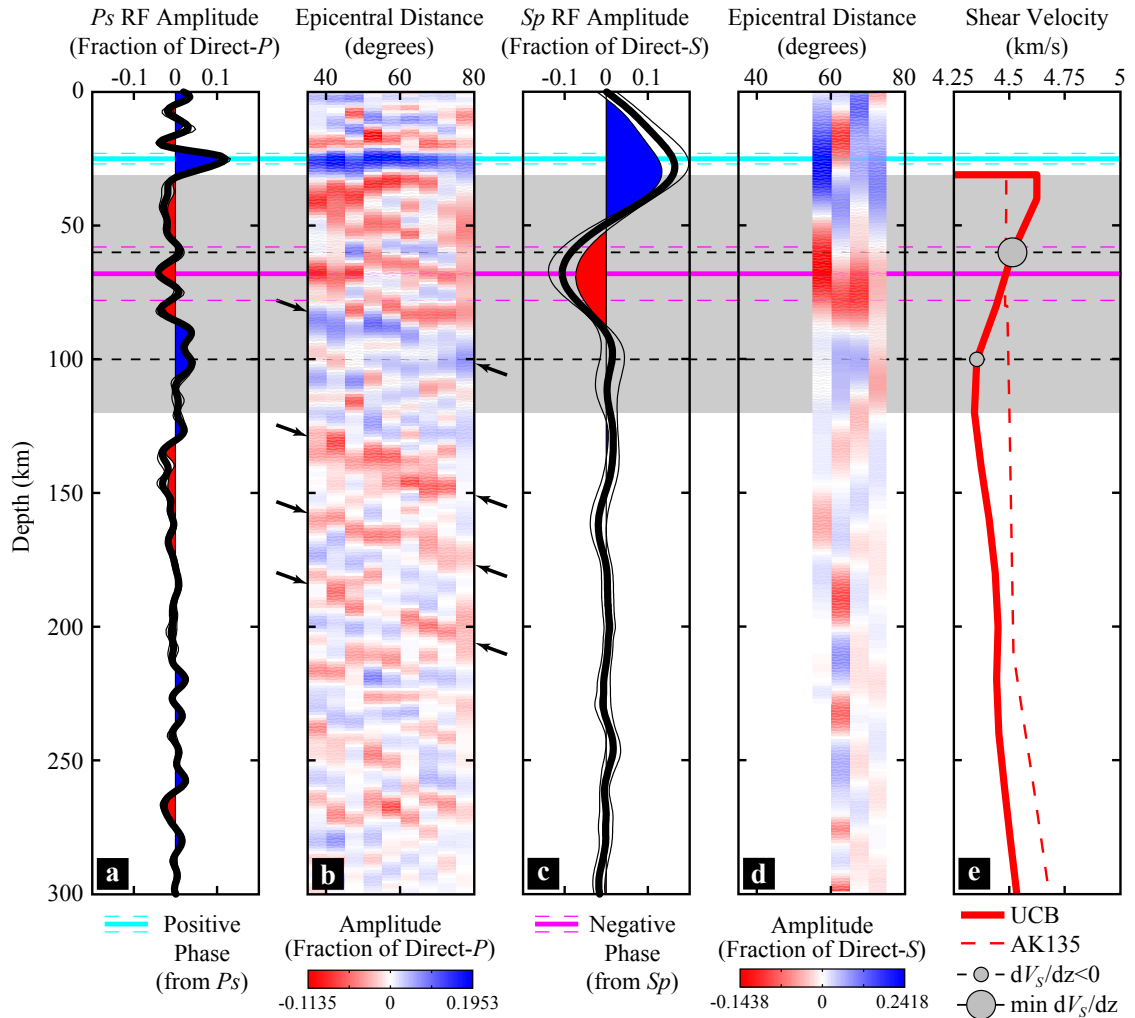


Figure 4. Detailed example of P_s and S_p receiver functions and V_s from the 3-D migration model used at station VTV. As in Figure 3, the mean single receiver functions (a and c) from the bootstrap testing are shown as thick black lines and $\pm 2\sigma$ is given by the thin black lines. The P_s and S_p receiver functions binned by distance are in b and d, respectively, and are not bootstrapped. Blue indicates positive amplitude (i.e., an impedance contrast resulting from a velocity increase with increasing depth) and red indicates negative. In a and c, the portions colored red and blue are those that fall above the mean- 2σ and below the mean+ 2σ , respectively. The Moho phase (large positive phase at reasonable crustal depth) is picked on the P_s receiver function and highlighted with the horizontal cyan line across each panel, and the most prominent negative phase following the Moho is picked on the S_p receiver function and marked with the magenta line. 2σ errors for both the Moho and negative S_p phase are given by the dashed cyan and magenta lines, respectively. Clear crustal reverberations are marked in b with black arrows. The mantle shear velocity profile at VTV extracted from the 3-D model [Yuan *et al.*, in preparation] is shown with the thick red line in e; the dashed red line shows AK135 for reference. Gray circles mark points in the V_s profile where a negative trough in the gradient of V_s exists, with the larger circle being the maximum negative dV_s/dz ; thin dashed black lines extend from these points. The minimum V_s value below the Moho and above 300 km up to the next peak in V_s (i.e., where dV_s/dz becomes negative) is shown by the gray shaded region. We consider this gray region the most likely location for the LAB based on the 3-D V_s model. If the most prominent negative S_p phase falls within this region, as is the case here, we interpret the phase to represent a conversion at the LAB (Table 2). This example is representative of most stations in the western U.S. in that the negative S_p phase has a much larger amplitude and is found at a shallower depth than in the central and eastern U.S.

ULM (CN) - Superior Craton

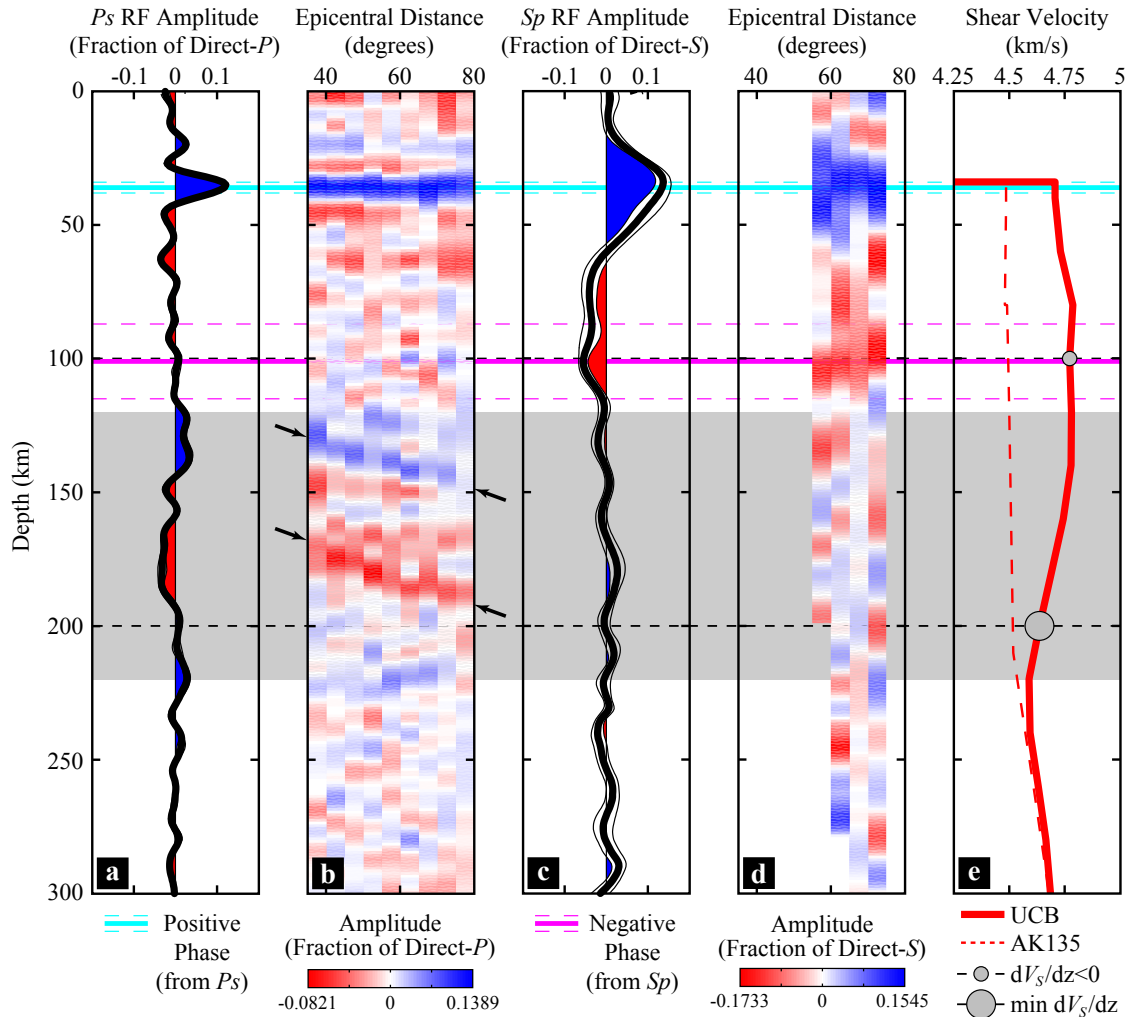


Figure 5. Detailed example of P_s and S_p receiver functions and V_s from the 3-D migration model used at station ULM. Notice that not all S_p distance bins contain receiver function data down to 300 km. The specific velocity model at this station is such that only waves from greater epicentral distances have pre-critical incidence angles at shallower conversion depths (e.g., for the 55°-60° bin, only conversions at depths above 200 km are predicted). In contrast to the previous example (VTV, Fig. 4), the negative S_p phase is shallower than the likely LAB range (gray area), and so we interpret it here as the result of a mid-lithospheric discontinuity (MLD). See the caption from Figure 4 for a full explanation.

HRV (IU) - Atlantic Margin, Massachusetts

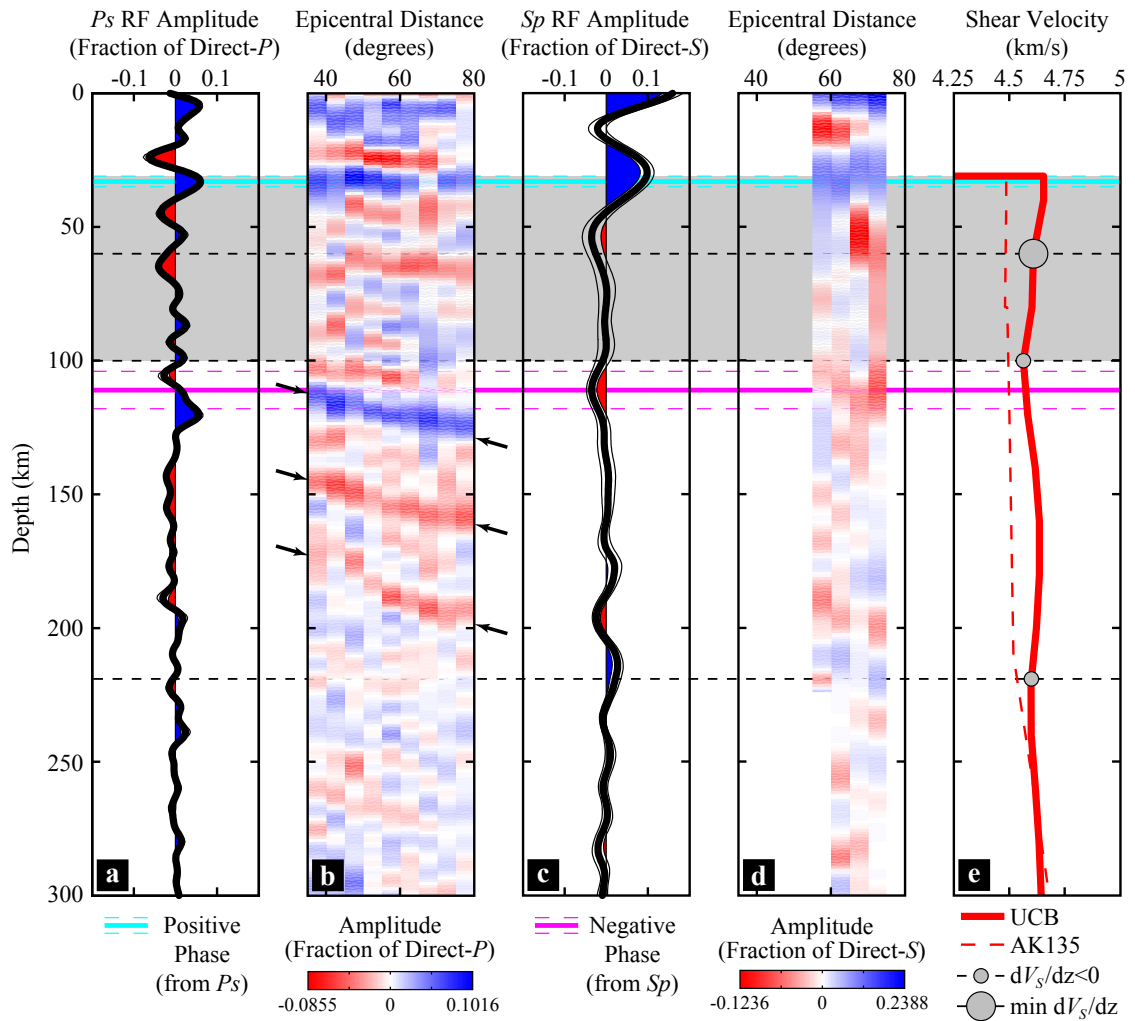


Figure 6. Detailed example of P_s and S_p receiver functions and V_s from the 3-D migration model used at station HRV. In this case, the negative S_p phase picked (111 ± 7 km) is located slightly deeper than the minimum V_s (100 km), and we interpret the phase to be caused by the LAB. See the caption from Figure 4 for a full explanation.

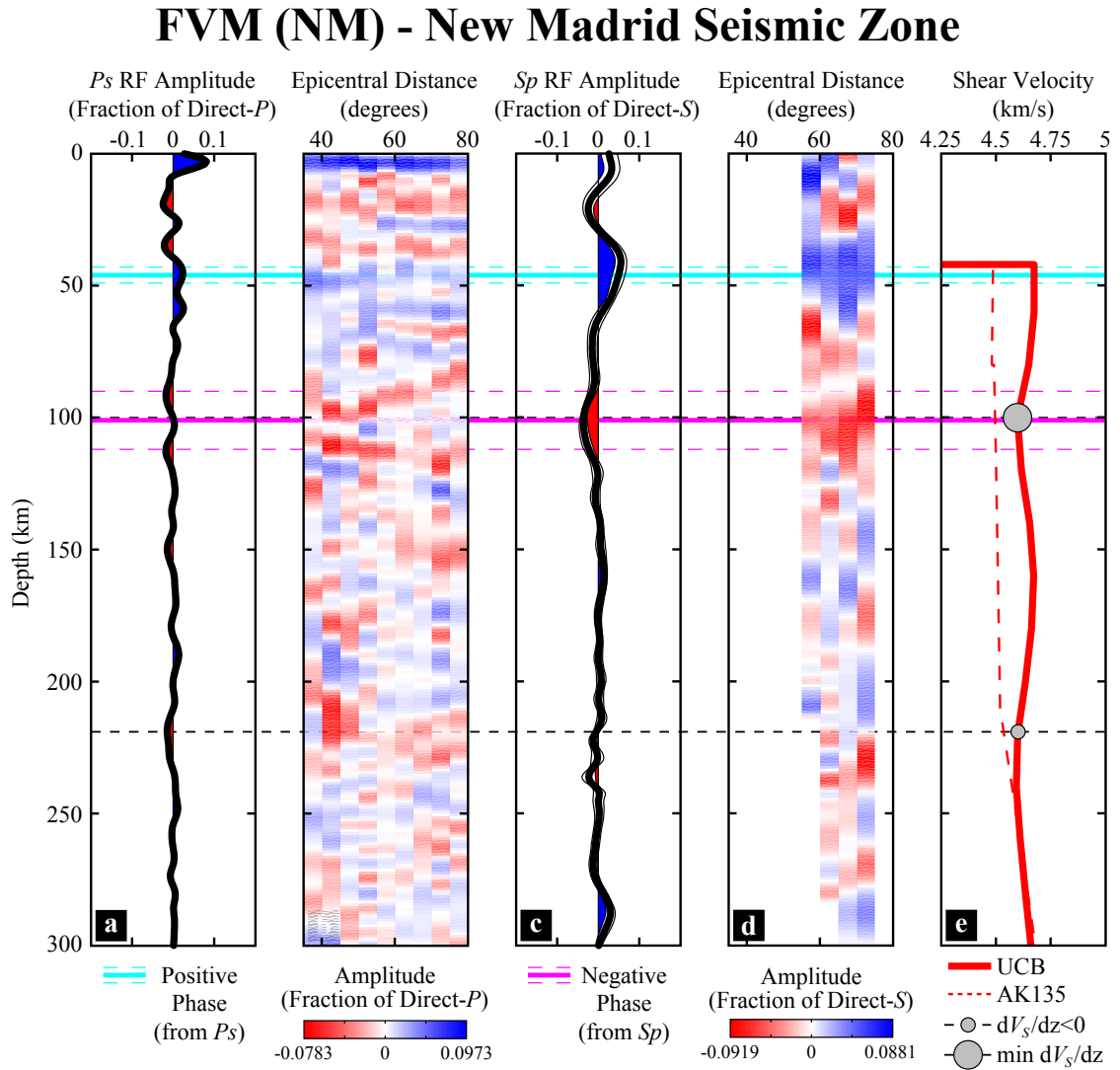


Figure 7. Detailed example of P_s and S_p receiver functions and V_s from the 3-D migration model used at station FVM. The high velocity lid is not as pronounced here as it is at ULM (Fig. 5), and two low velocity zones exist (100 km and 220 km), with only a small difference in absolute velocity between the two (0.1166%; see Section 4.2.4). Given our threshold difference of 0.5% (see Section 4.1), this similarity between the two V_s minima does not allow us to unambiguously define the LAB from the V_s model, and therefore we do not interpret the negative S_p phase here (101 ± 11 km) as either the LAB or a MLD. See the caption from Figure 4 for a full explanation.

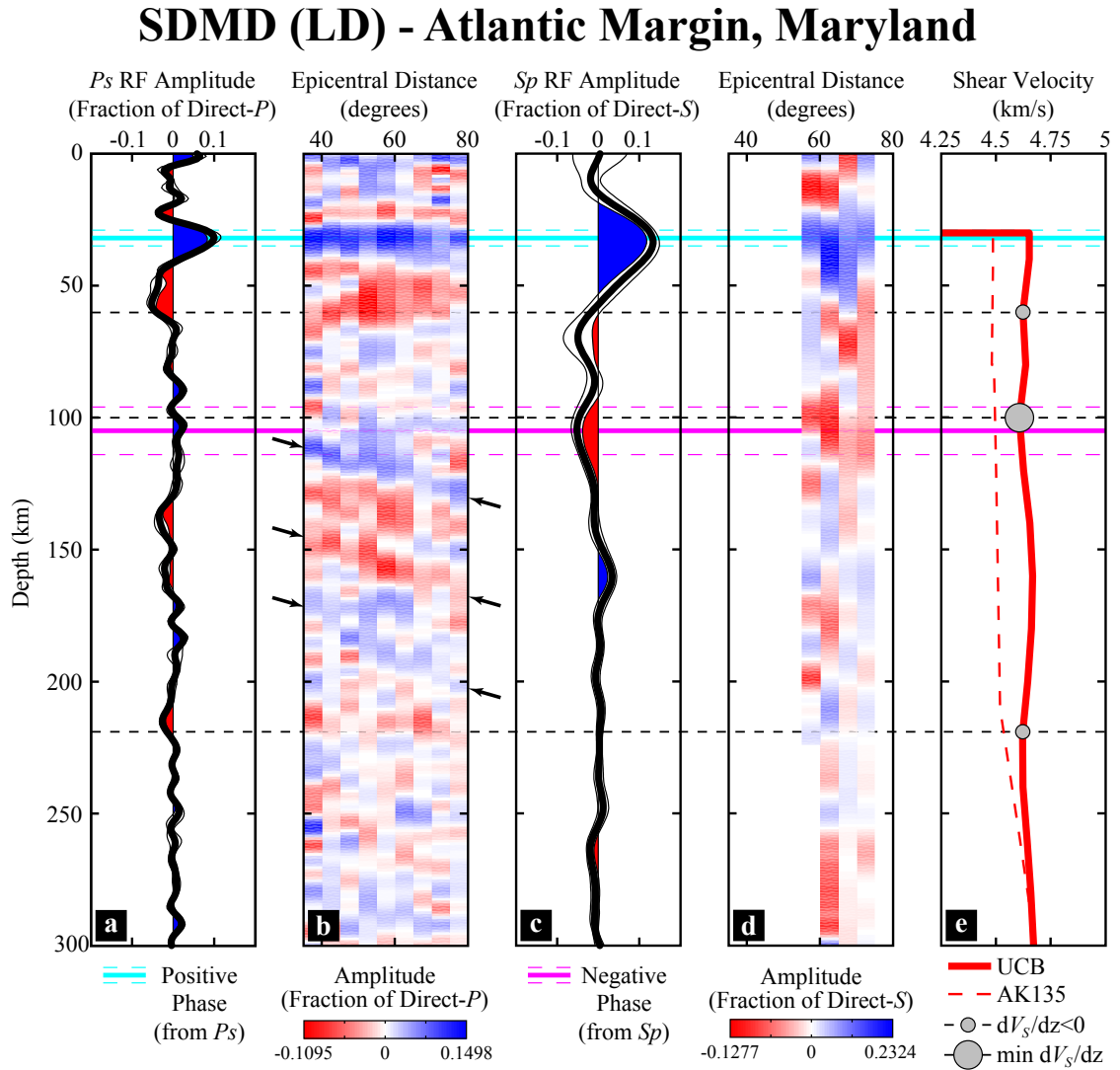


Figure 8. Detailed example of P_s and S_p receiver functions and V_s from the 3-D migration model used at station SDMD. Similar to FVM (Fig. 7), the depth of the LAB from the V_s model is ambiguous, and therefore we cannot interpret the negative S_p phase (105 ± 9 km) as either the LAB or a MLD. See the caption from Figure 4 for a full explanation.

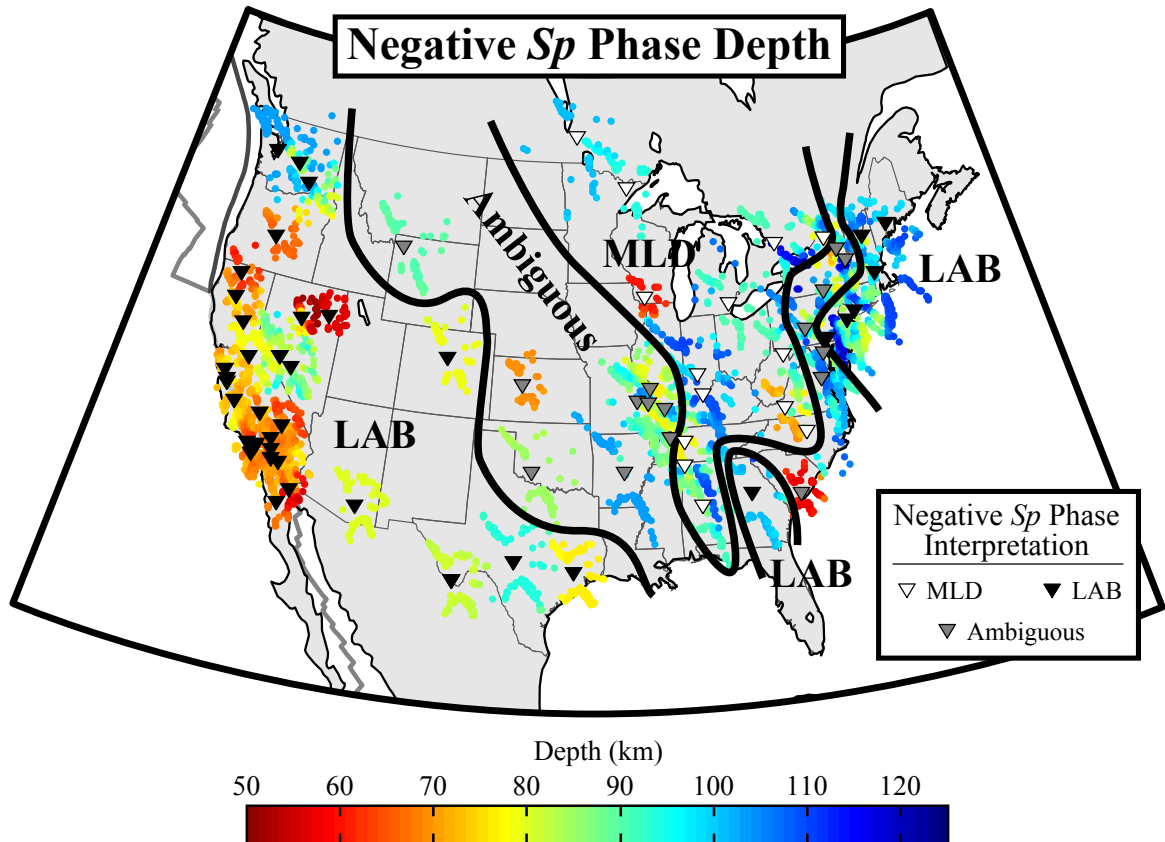


Figure 9a. Depths of the prominent negative S_p phase picked on the receiver functions migrated with velocity profiles extracted from the 3-D mantle model. The dots are colored for depth and represent S_p piercing points that have been placed onto a fine grid and smoothed with a circular filter (see Section 4.3 for details). Black inverted triangles indicate stations where the negative S_p phase is interpreted as the lithosphere-asthenosphere boundary (LAB), white inverted triangles are stations where the phase is interpreted as a mid-lithospheric discontinuity (MLD), and gray stations indicate ambiguity in the depth of the LAB from the V_S model and thus ambiguity in the interpretation of the negative S_p phase; the thick black lines separates these regions. Southern and northern California, northwestern Nevada, Wisconsin, and South Carolina all display shallow depths (50-60 km), while depths greater than 90 km are found in New England, the central U.S., and the Puget Sound. Stations along the Phanerozoic margin (southern and eastern U.S.) have negative S_p phases interpreted as the LAB but with smaller amplitudes and slightly greater depths than in the western U.S. See Figure 9b for additional description.

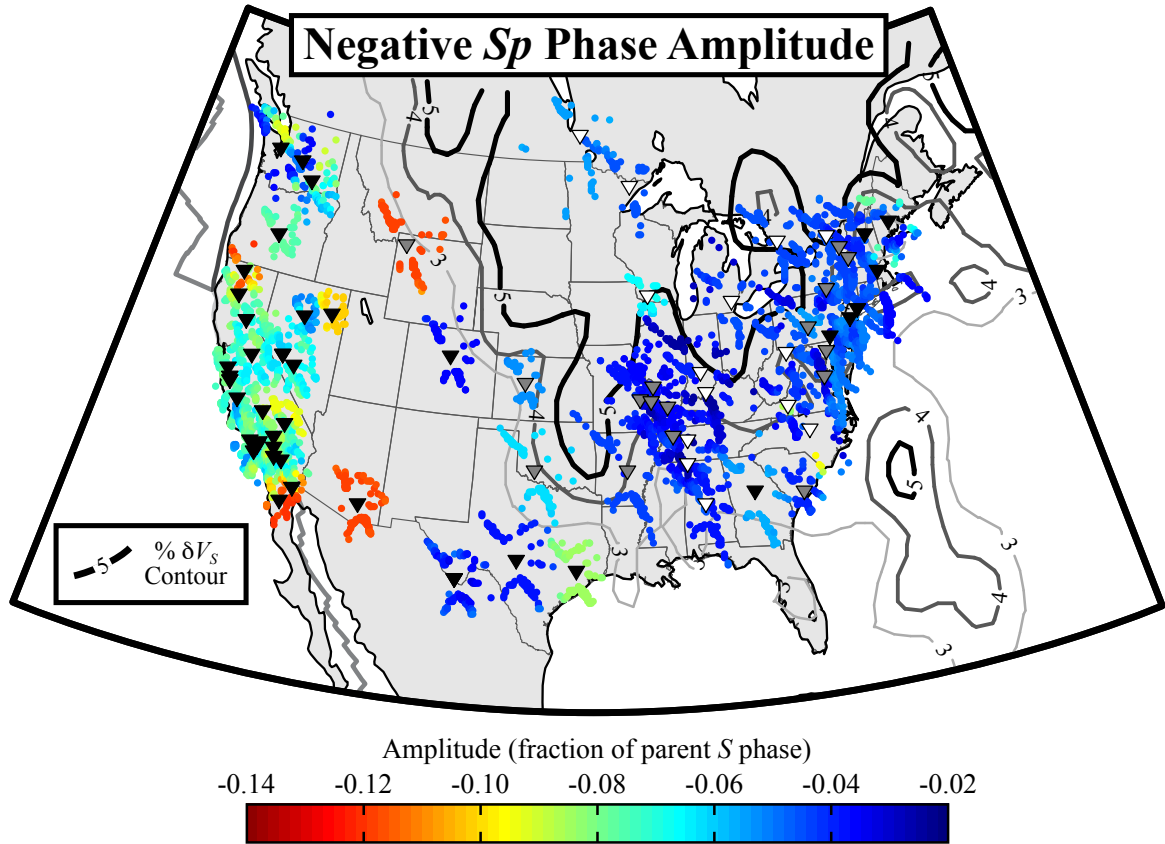


Figure 9b. Amplitudes of the prominent negative Sp phase picked on the receiver functions migrated with velocity profiles extracted from the 3-D mantle model. The dots are colored for amplitude and represent Sp piercing points that have been placed onto a fine grid and smoothed with a circular filter (see Section 4.3 for details). The region of thicker lithosphere in the eastern U.S. is illustrated by the 3%-5% positive velocity anomaly contours from the V_s model at 140 km depth [Yuan *et al.*, in preparation]. Broadly speaking, the thick lithosphere delineated by the high V_s anomaly correlates with negative Sp phases interpreted as a MLD and with lower amplitudes, and the western U.S. is characterized by negative Sp phases with larger amplitudes and interpreted at the LAB. Stations along the Phanerozoic margin (southern and eastern U.S.) have negative Sp phases interpreted as the LAB but with smaller amplitudes and slightly greater depths than in the western U.S. See Figure 9a caption for additional description.

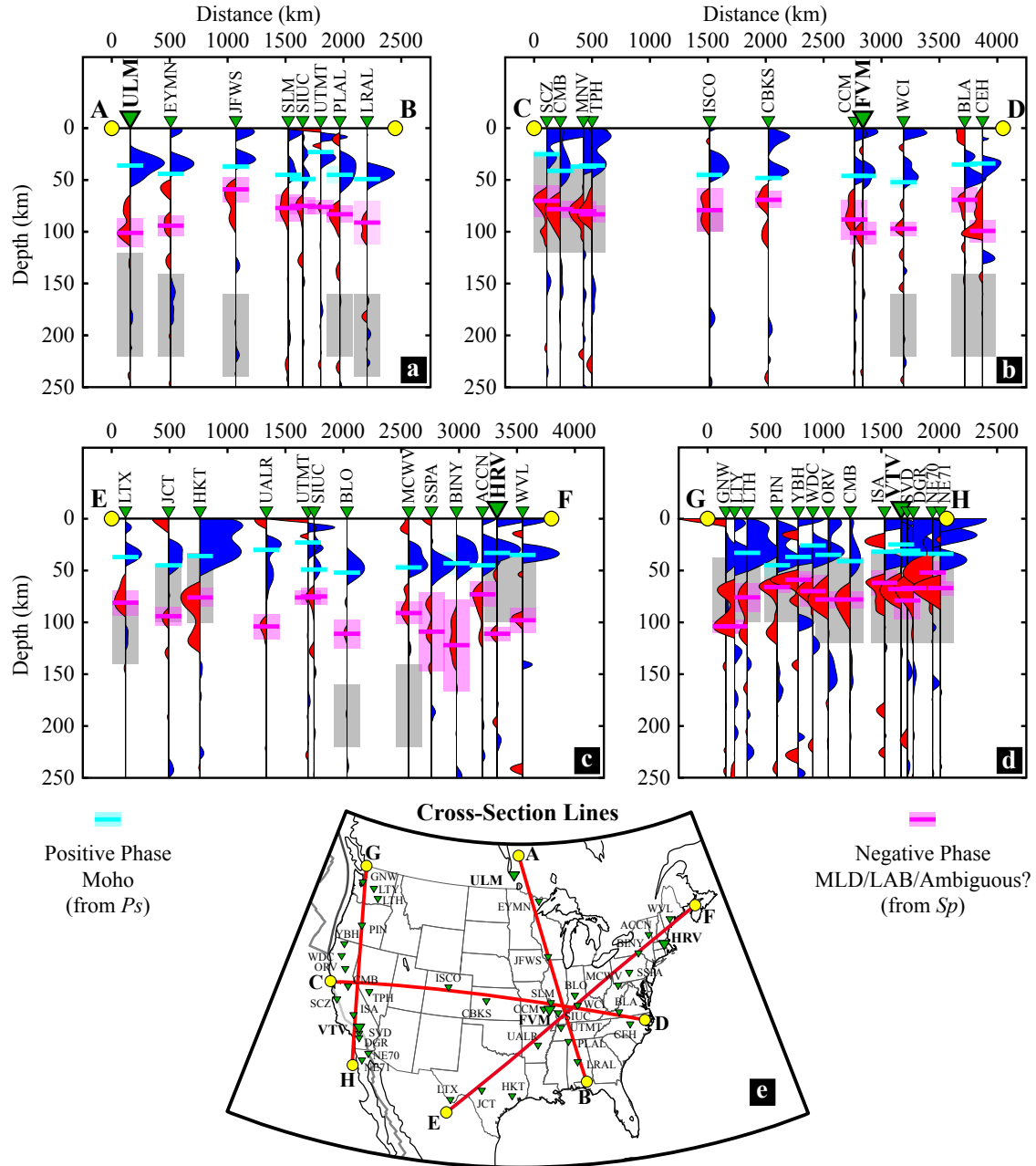


Figure 10. Cross-sections illustrating the relationship between S_p receiver functions and both the Moho (cyan line in **a-d**) and negative S_p phase (magenta line in **a-d**) depths. The LAB depth range inferred from the V_s model, when unambiguous, is shown by the gray shaded regions in **a-d** (see Section 4 and caption to Figure 4 for explanation). In general, the negative S_p phases are located between 60 and 100 km. In the western and northeastern U.S., as well as some stations in the southeast, the inferred transition from the minimum V_s up to the next peak in V_s falls in this same depth range, and we interpret the negative S_p phases in these regions to represent the conversion of energy at the LAB. In contrast, the transition from the LVZ to the higher velocity lid beneath the central U.S. occurs at roughly 200 km, and here the negative S_p phases are interpreted at a MLD. The receiver functions in each cross-section are plotted at the same scale with only resolved positive (blue) and negative (red) amplitudes shown; the larger amplitude of the negative phases along the west coast (Line G-H, **d**) relative to other regions is easily observed. A detailed description of each cross-section can be found in Section 4.4. Only stations shown in **a-d** are plotted on the reference map in **e** with the cross-section lines. Supplemental Figure S4 shows these same receiver functions down to 400 km.

NET	Network Name	Institution/Affiliation
AZ	ANZA Regional Network	UC San Diego
BK	Berkeley Digital Seismograph Network	UC Berkeley
CI	Caltech Regional Seismic Network	California Institute of Technology (Caltech)
CN	Canadian National Seismograph Network	Geological Survey of Canada (GSC)
CT	California Transect Network	UC Santa Cruz
G	GEOSCOPE	IPGP (Institut de Physique du Globe de Paris) CNR/ULP
IM	International Miscellaneous Stations	
IU	Global Seismograph Network (GSN-IRIS/USGS)	Albuquerque Seismological Laboratory (ASL)/USGS
LB	Leo Brady Network	Sandia National Laboratories
LD	Lamont-Doherty Cooperative Seismographic Network	LDEO, Columbia University
LI	Laser Interferometer Gravitational-Wave Experiment (LIGO)	California Institute of Technology (Caltech)
NE	Laser Interferometer Gravitational-Wave Experiment (LIGO)	California Institute of Technology (Caltech)
NM	Cooperative New Madrid Seismic Network	Saint Louis University
NR	NARS-Baja Seismic Network	Utrecht University
TS	TERRAscope (Southern California Seismic Network)	California Institute of Technology (Caltech)
UO	University of Oregon Regional Network	University of Oregon
US	United States National Seismic Network	USGS National Earthquake Information Center (NEIC)
UW	Pacific Northwest Regional Seismic Network	University of Washington

Table 1. Network affiliations for the seismic stations used in this study. All data was acquired from the IRIS DMC.

NET	STA	# <i>Ps</i>	# <i>Sp</i>	Crustal Model				Receiver Function Interpretations				
				Source	Moho	V_P	V_P/V_S	M_{3D} (M_{1D})	$\epsilon_{M3D}(\epsilon_{M1D})$	N_{3D} (N_{1D})	$\epsilon_{N3D}(\epsilon_{N1D})$	LAB?
BK	CMB	724	340	Crust 2.0	38	6.31	1.78	41 (41)	2 (1)	78 (76)	8 (9)	LAB
BK	JRSC	188	106	H- <i>k</i> Stack	33	6.39	1.59	32 (32)	3 (3)	-	-	-
BK	ORV	238	117	H- <i>k</i> Stack	34	6.48	1.85	35 (35)	4 (4)	78 (77)	8 (9)	LAB
BK	PKD	197	101	H- <i>k</i> Stack	28	6.18	1.81	30 (29)	2 (2)	79 (76)	18 (18)	LAB
BK	WDC	274	144	Crust 2.0	37	6.56	1.77	26 (27)	2 (2)	70 (68)	15 (16)	LAB
BK	YBH	108	53	Crust 2.0	34	6.56	1.77	37 (37)	8 (7)	59 (58)	8 (8)	LAB
CI	CWC	482	273	H- <i>k</i> Stack	30	6.27	1.8	32 (30)	32 (30)	-	-	-
CI	DGR	626	326	H- <i>k</i> Stack	33	6.23	1.78	34 (34)	1 (1)	67 (65)	9 (9)	LAB
CI	GSC	607	313	H- <i>k</i> Stack	23	6.25	1.96	26 (26)	2 (2)	61 (59)	6 (6)	LAB
CI	RPV	436	236	H- <i>k</i> Stack	29	6.28	1.6	30 (30)	2 (2)	65 (63)	8 (7)	LAB
CI	SVD	336	173	H- <i>k</i> Stack	31	6.22	1.96	31 (31)	2 (2)	79 (75)	18 (16)	LAB
CI	TOV	411	230	H- <i>k</i> Stack	23	6.25	1.91	36 (35)	2 (2)	75 (72)	24 (23)	LAB
CI	USC	243	227	Crust 2.0	27	6.26	1.78	-	-	58 (55)	13 (12)	LAB
CI	VTV	369	187	H-<i>k</i> Stack	25	6.24	1.92	25 (25)	2 (2)	68 (66)	10 (10)	LAB
CN	SADO	187	120	H- <i>k</i> Stack	38	6.49	1.77	40 (40)	2 (2)	90 (90)	21 (19)	MLD
CN	ULM	205	85	H-<i>k</i> Stack	34	6.42	1.73	36 (36)	2 (2)	101 (96)	14 (14)	MLD
CT	FERN	45	24	Crust 2.0	29	6.38	1.79	24 (25)	16 (14)	66 (63)	8 (7)	LAB
CT	SAVY	10	8	Crust 2.0	31	6.35	1.79	25 (25)	2 (2)	-	-	-
G	SCZ	735	419	H- <i>k</i> Stack	24	6.32	1.77	25 (24)	4 (3)	70 (66)	15 (16)	LAB
IM	NV31	176	94	H- <i>k</i> Stack	36	6.28	1.74	37 (36)	2 (2)	102 (100)	10 (9)	LAB
IM	PD31	340	143	Crust 2.0	45	6.18	1.76	36 (36)	1 (1)	-	-	-
IM	TX31	272	148	H- <i>k</i> Stack	36	6.34	1.8	37 (37)	2 (2)	83 (82)	19 (19)	LAB
IM	TX32	245	138	H- <i>k</i> Stack	36	6.34	1.81	36 (36)	1 (1)	80 (79)	17 (17)	LAB
IU	CCM	373	256	H- <i>k</i> Stack	45	6.5	1.79	46 (46)	2 (1)	88 (86)	19 (13)	???
IU	DWPF	312	201	H- <i>k</i> Stack	35	6.24	1.91	36 (36)	2 (2)	-	-	-
IU	HKT	340	275	H- <i>k</i> Stack	34	5.73	1.61	36 (35)	2 (2)	76 (73)	9 (9)	LAB
IU	HRV	635	421	H-<i>k</i> Stack	31	6.4	1.66	33 (33)	2 (2)	111 (106)	7 (7)	LAB
IU	SSPA	242	140	Crust 2.0	39	6.32	1.76	-	-	109 (108)	38 (12)	???
IU	TUC	500	241	H- <i>k</i> Stack	31	6.22	1.68	33 (33)	3 (3)	80 (79)	7 (7)	LAB
IU	WCI	289	177	Crust 2.0	41	6.47	1.77	52 (52)	5 (4)	97 (93)	7 (7)	MLD
IU	WVT	298	203	H- <i>k</i> Stack	36	6.46	1.99	38 (38)	1 (1)	113 (110)	7 (6)	MLD
LB	BMN	406	178	H- <i>k</i> Stack	29	6.28	1.77	30 (30)	2 (2)	51 (51)	32 (30)	LAB
LB	DAC	172	89	H- <i>k</i> Stack	32	6.28	1.75	33 (33)	4 (4)	-	-	-
LB	MVU	62	39	H- <i>k</i> Stack	38	6.23	1.79	42 (41)	2 (3)	-	-	-
LB	TPH	402	190	H- <i>k</i> Stack	35	6.28	1.77	36 (36)	2 (2)	83 (81)	8 (7)	LAB
LD	ACCN	311	204	H- <i>k</i> Stack	44	6.44	1.79	45 (45)	2 (2)	73 (73)	12 (10)	???
LD	BRNJ	246	153	H- <i>k</i> Stack	40	6.14	1.79	40 (41)	3 (3)	89 (87)	8 (8)	LAB
LD	LOZ	26	22	H- <i>k</i> Stack	41	6.42	1.86	42 (43)	3 (4)	-	-	-
LD	MVL	263	161	Crust 2.0	33	6.31	1.79	45 (44)	6 (6)	75 (72)	16 (15)	LAB
LD	PAL	456	298	H- <i>k</i> Stack	40	6.17	1.67	42 (42)	3 (3)	79 (77)	10 (9)	LAB
LD	PTN	104	58	H- <i>k</i> Stack	39	6.42	1.81	38 (37)	2 (3)	69 (68)	11 (11)	MLD
LD	SDMD	253	159	H-<i>k</i> Stack	30	6.3	1.71	32 (32)	3 (3)	105 (101)	9 (9)	???
LI	LTH	285	138	H- <i>k</i> Stack	33	6.55	1.87	33 (33)	1 (1)	76 (75)	14 (14)	LAB
NE	FFD	86	64	Crust 2.0	37	6.42	1.76	29 (29)	2 (2)	-	-	-
NE	HNH	19	57	Crust 2.0	34	6.45	1.78	32 (30)	7 (5)	-	-	-
NE	PQI	87	70	H- <i>k</i> Stack	33	6.42	1.79	34 (34)	3 (2)	-	-	-
NE	WES	73	62	H- <i>k</i> Stack	30	6.36	1.73	32 (32)	2 (2)	-	-	-
NE	WVL	78	45	H- <i>k</i> Stack	35	6.42	1.74	35 (36)	2 (3)	98 (94)	12 (11)	LAB
NE	YLE	73	51	H- <i>k</i> Stack	30	6.25	1.74	32 (32)	3 (2)	-	-	-

Table 2. Data table of pertinent receiver function information. For each station, the number of waveforms used in the deconvolution and the crustal model used in the migration are given. The depth picks of the most prominent positive *Ps* (M) and negative *Sp* (N) phase from the receiver functions are presented for both the 3-D and 1-D (in parentheses) migration models, along with their errors (ϵ). Based on the 3-D VS model used here, we interpret the negative *Sp* phase as the lithosphere-asthenosphere boundary (LAB), a mid-lithospheric discontinuity (MLD), or ambiguous (???). See the beginning of Section 4 for an explanation of the basis for these interpretations. Individual stations shown in Figures 4-8 are in bold (VTV, ULM, HRV, FVM, and SDMD).

NET	STA	# <i>Ps</i>	# <i>Sp</i>	Source	Moho	V_p	V_p/V_s	M_{3D} (M_{1D})	$\epsilon_{M3D}(\epsilon_{M1D})$	N_{3D} (N_{1D})	$\epsilon_{N3D}(\epsilon_{N1D})$	LAB?
NM	BLO	398	234	Crust 2.0	41	6.48	1.77	52 (51)	2 (2)	111 (106)	14 (13)	MLD
NM	FVM	210	125	Crust 2.0	42	6.46	1.78	46 (45)	3 (3)	101 (99)	11 (10)	???
NM	MPH	188	215	Crust 2.0	38	6.31	1.78	38 (38)	2 (2)	-	-	-
NM	PLAL	211	148	Crust 2.0	36	6.51	1.77	45 (44)	8 (7)	83 (81)	8 (7)	MLD
NM	SIUC	331	206	Crust 2.0	41	6.44	1.77	49 (49)	3 (2)	75 (73)	8 (8)	???
NM	SLM	416	230	H- <i>k</i> Stack	44	6.45	1.93	45 (44)	2 (3)	77 (76)	13 (13)	???
NM	UALR	200	165	H- <i>k</i> Stack	29	6.25	1.93	30 (30)	3 (3)	104 (100)	12 (15)	???
NM	UTMT	209	180	H- <i>k</i> Stack	21	6.44	2	23 (23)	1 (1)	76 (72)	7 (6)	???
NR	NE70	97	55	Crust 2.0	30	6.26	1.78	-	-	52 (50)	11 (9)	LAB
NR	NE71	169	73	H- <i>k</i> Stack	33	6.26	1.82	34 (34)	2 (2)	67 (66)	7 (7)	LAB
TS	PAS	869	500	H- <i>k</i> Stack	27	6.26	1.76	28 (28)	3 (3)	70 (66)	8 (8)	LAB
TS	PFO	356	211	H- <i>k</i> Stack	29	6.2	1.72	29 (29)	2 (2)	77 (73)	12 (14)	LAB
UO	PIN	254	129	Crust 2.0	37	6.42	1.77	45 (42)	11 (8)	66 (65)	6 (6)	LAB
US	AAM	215	114	H- <i>k</i> Stack	41	6.49	1.91	43 (41)	6 (7)	90 (86)	11 (9)	MLD
US	ACSO	251	159	H- <i>k</i> Stack	43	6.25	1.86	44 (44)	16 (15)	-	-	-
US	BINY	309	209	Crust 2.0	36	6.3	1.76	43 (43)	11 (10)	122 (116)	44 (43)	???
US	BLA	266	168	Crust 2.0	37	6.43	1.76	35 (36)	2 (2)	69 (68)	12 (9)	MLD
US	CBKS	330	173	H- <i>k</i> Stack	47	6.45	1.77	48 (48)	1 (1)	69 (68)	8 (8)	???
US	CBM	6		H- <i>k</i> Stack	39	6.42	1.78	41 (41)	3 (3)	-	-	-
US	CBN	187	120	Crust 2.0	32	6.37	1.78	-	-	90 (87)	20 (22)	???
US	CEH	25	15	H- <i>k</i> Stack	34	6.54	1.81	34 (34)	2 (2)	99 (95)	11 (10)	MLD
US	ELK	397	145	H- <i>k</i> Stack	31	6.28	1.74	32 (32)	2 (2)	56 (54)	6 (7)	LAB
US	EYMN	212	75	H- <i>k</i> Stack	41	6.47	1.79	44 (44)	2 (2)	94 (89)	10 (10)	MLD
US	GOGA	151	98	Crust 2.0	31	6.53	1.78	41 (32)	3 (2)	101 (90)	5 (5)	LAB
US	GWDE	35	29	Crust 2.0	29	5.95	1.8	38 (38)	3 (3)	-	-	-
US	ISA	38	22	H- <i>k</i> Stack	31	6.24	1.92	32 (32)	2 (1)	62 (61)	12 (11)	LAB
US	ISCO	386	162	H- <i>k</i> Stack	25	6.19	1.98	45 (43)	4 (3)	79 (76)	21 (20)	LAB
US	JCT	212	151	H- <i>k</i> Stack	44	6.35	1.86	45 (44)	2 (2)	94 (94)	9 (12)	LAB
US	JFWS	194	97	H- <i>k</i> Stack	35	6.53	1.98	37 (36)	3 (3)	59 (58)	12 (11)	MLD
US	LBNH	297	189	Crust 2.0	39	6.42	1.75	40 (40)	3 (3)	99 (93)	27 (35)	LAB
US	LKWY	265	186	H- <i>k</i> Stack	30	6.2	1.61	30 (30)	2 (2)	90 (87)	7 (6)	???
US	LRAL	189	127	Crust 2.0	32	6.46	1.78	49 (48)	2 (2)	91 (91)	21 (19)	MLD
US	LSCT	26	53	H- <i>k</i> Stack	25	6.37	1.9	24 (24)	3 (3)	-	-	-
US	LTX	202	125	H- <i>k</i> Stack	36	6.34	1.8	37 (37)	2 (2)	81 (79)	12 (22)	LAB
US	MCWV	280	178	Crust 2.0	44	6.22	1.76	47 (45)	3 (2)	91 (87)	11 (12)	MLD
US	MIAR	264	175	H- <i>k</i> Stack	47	6.29	1.77	49 (49)	2 (2)	-	-	-
US	MNV	256	67	H- <i>k</i> Stack	36	6.28	1.73	37 (37)	2 (1)	80 (78)	10 (12)	LAB
US	MYNC	54	42	Crust 2.0	37	6.45	1.77	55 (54)	2 (2)	-	-	-
US	NCB	319	204	H- <i>k</i> Stack	39	6.42	1.91	40 (40)	2 (2)	109 (106)	19 (19)	???
US	NHSC	148	116	Crust 2.0	30	6.2	1.79	36 (36)	5 (5)	57 (55)	6 (6)	???
US	SAO	183	100	Crust 2.0	28	6.33	1.79	34 (32)	28 (21)	79 (75)	13 (12)	LAB
US	WMOK	167	119	H- <i>k</i> Stack	31	6.39	1.53	30 (30)	3 (3)	84 (81)	7 (8)	???
UW	GNW	191	151	Crust 2.0	38	6.58	1.77	-	-	104 (104)	6 (6)	LAB
UW	LTY	195	104	Crust 2.0	40	6.52	1.76	-	-	104 (104)	7 (6)	LAB

Table 2 (continued).

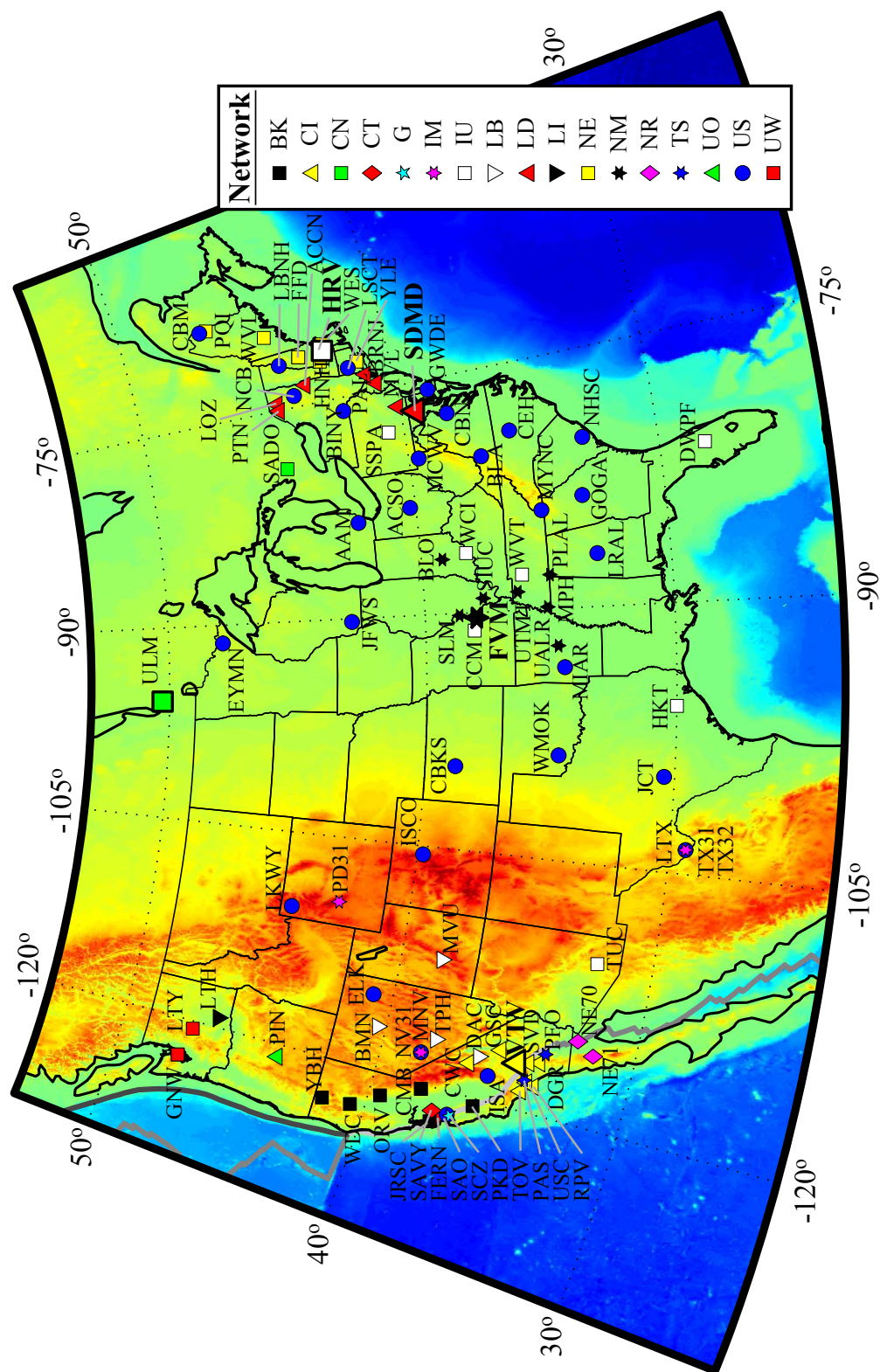


Figure S1. Station locations for interpretable P_s and S_p receiver functions (Table 2). Topography and bathymetry are from *Smith and Sandwell* [1997], and network affiliations are in Table 1.

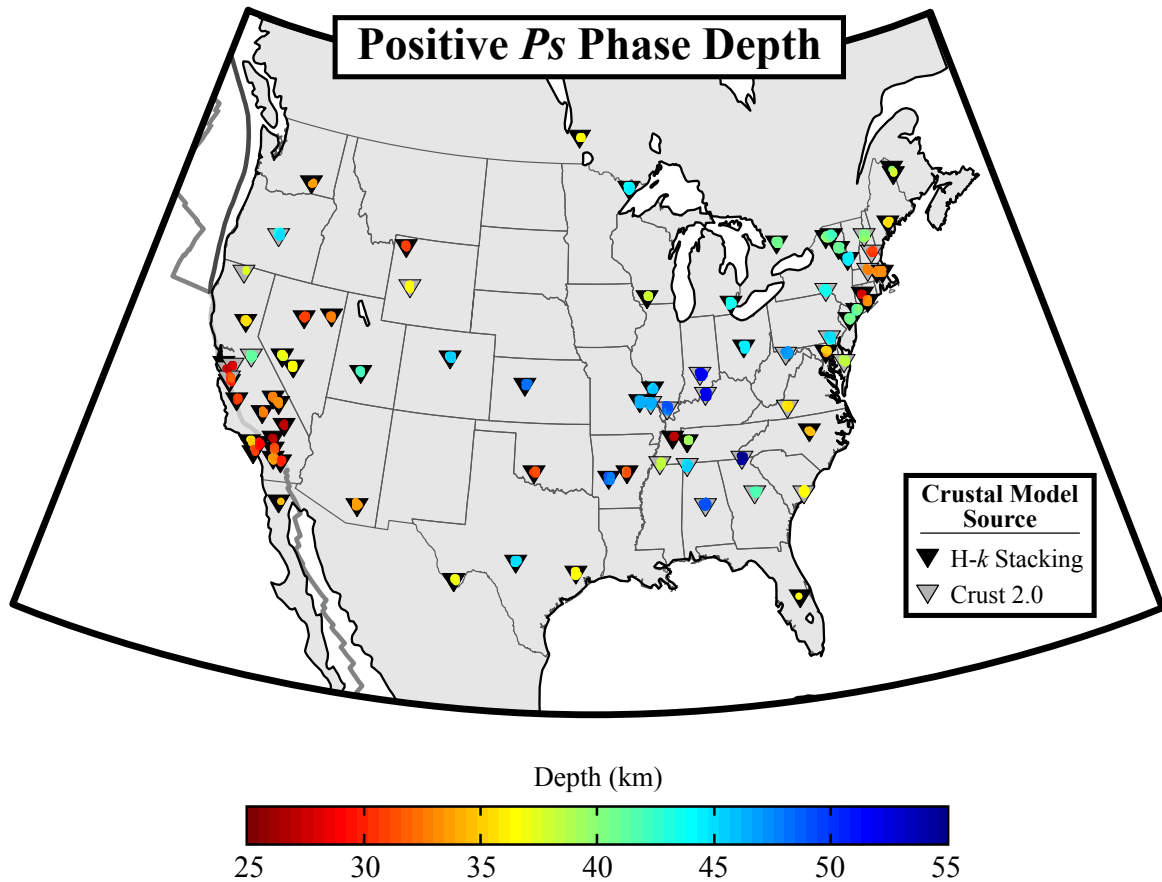


Figure S2. Depth of the positive P_s phase (i.e., the Moho). These points are generated in the same fashion as those in Figure 9a, except that here the picks for Moho depth were made on the P_s receiver functions. The station color indicates whether the crustal model was calculated by H- k stacking (black) [Zhu and Kanamori, 2000] or extracted from Crust2.0 (gray) [Bassin *et al.*, 2000, available online at <http://mahi.ucsd.edu/Gabi/crust2.html>]. These Moho depths are in good agreement with those from Chulick and Mooney [2002] where we have station coverage, with the exception of WMOK in southern Oklahoma where we observe a crustal thickness of 30 km compared with >45 km from Chulick and Mooney [2002].

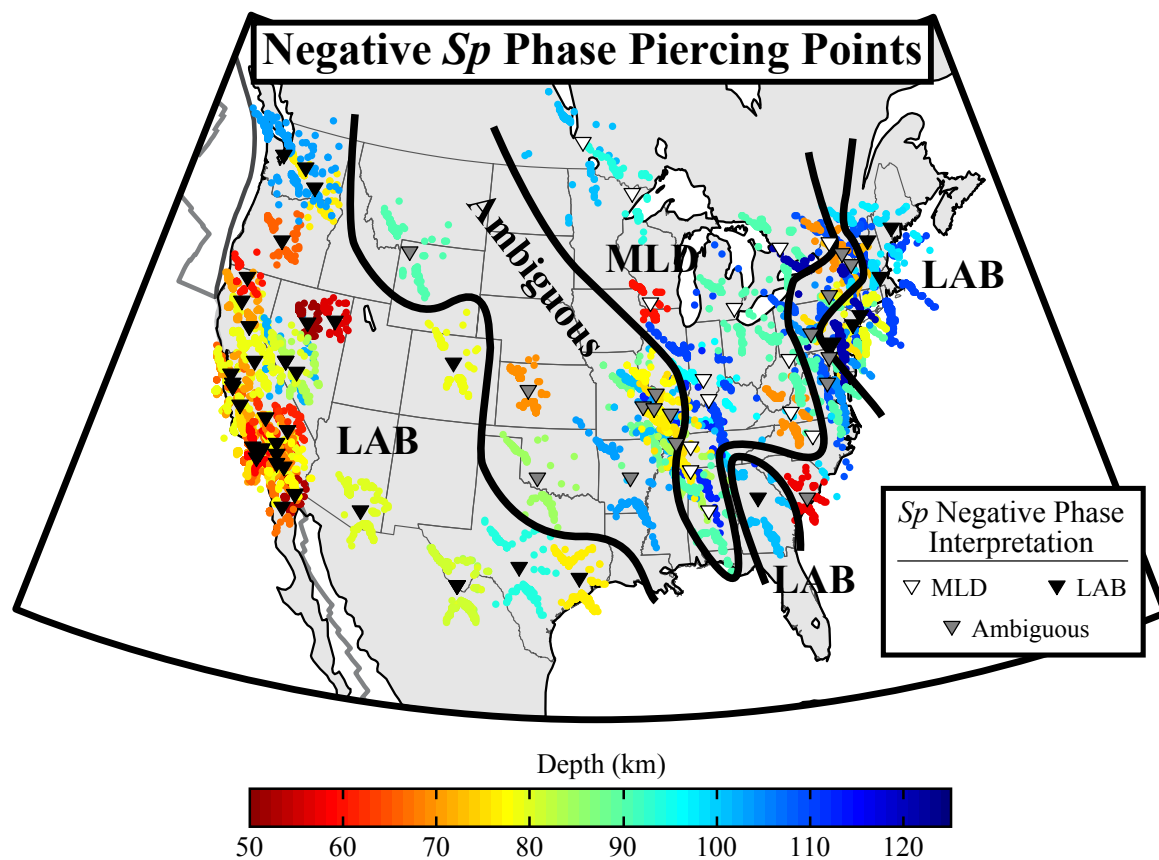


Figure S3. Same as Figure 9a, except the points shown are individual piercing points, rather than averaged values at grid points that have then been smoothed. Given the short wavelength of smoothing relative to the spread of piercing points around each station, there is often very little difference between the two plots.

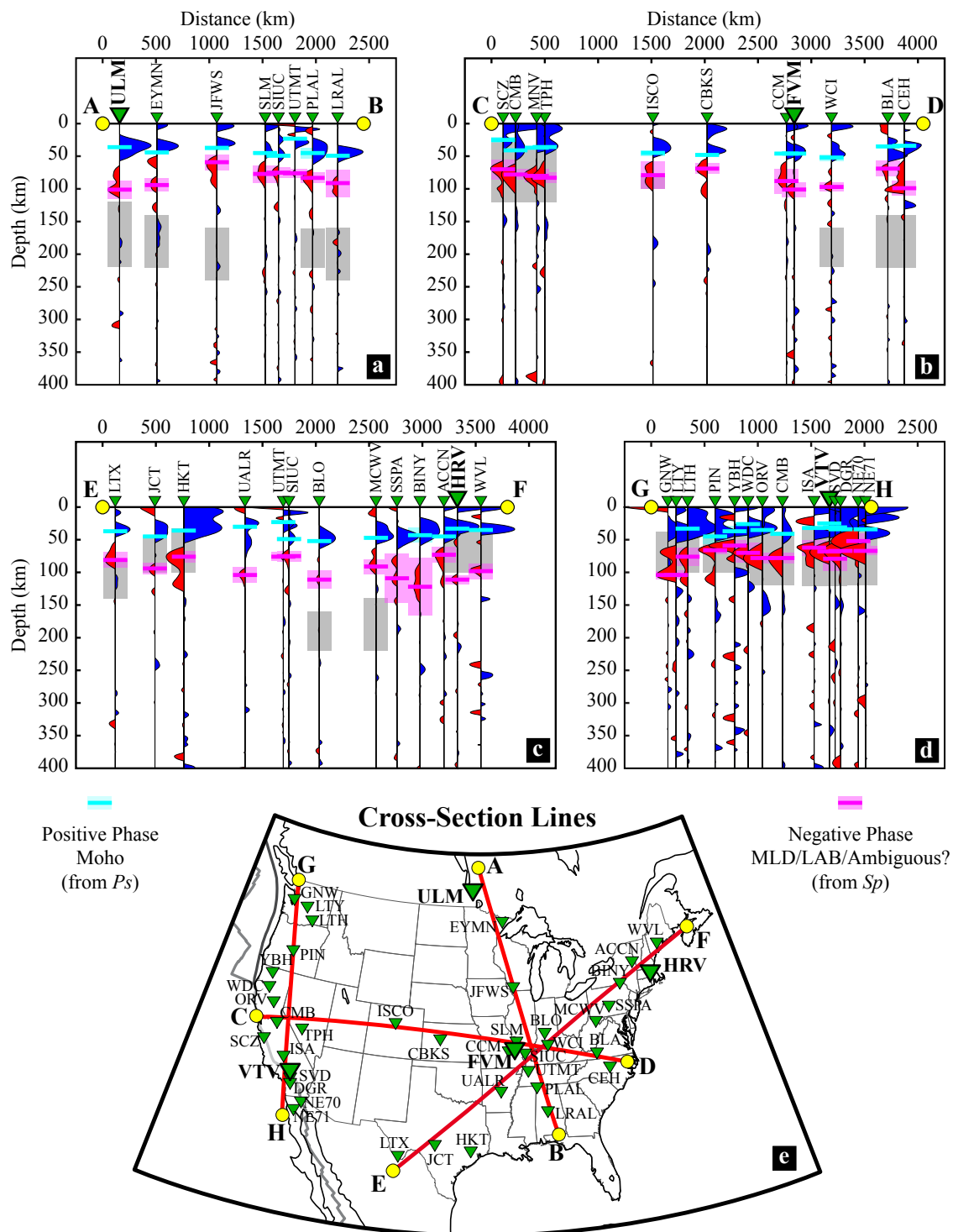


Figure S4. Same as Figure 10, except the receiver functions extend down to 400 km.

Appendix 1

Arc-Parallel Flow in the Mantle Wedge Beneath Costa Rica and Nicaragua

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1. Summary

Resolving flow geometry in the mantle wedge is key to understanding the thermal and chemical structure of subduction zones, subducting plate dehydration, and melting that leads to arc volcanism, which can threaten large populations and alter climate through gas and particle emission. Here we show that isotope geochemistry and seismic velocity anisotropy provide strong evidence for trench-parallel flow in the mantle wedge beneath Costa Rica and Nicaragua. This finding contradicts classical models which predict trench-normal flow due to the overlying wedge mantle being dragged downwards by the subducting plate. The isotopic signature of central Costa Rican volcanic rocks is not consistent with its derivation from the wedge¹⁻³ or eroded fore-arc complexes⁴ but instead from seamounts of the Galapagos hotspot track on the subducting Cocos Plate. This isotopic signature decreases continuously from Central Costa Rica to northwestern Nicaragua. Since the age of the isotopic signature beneath Costa Rica can be constrained and its transport distance is known, minimum northwestward flow rates can be estimated (63-190 mm/yr) and are comparable to the magnitude of subducting Cocos Plate motion (~85 mm/yr). Trench-parallel flow needs to be taken into account in models evaluating thermal and chemical structure and melt generation in subduction zones.

2. Data and Discussion

Preferential alignment of the minerals olivine and orthopyroxene occurs as a result of deformation and produces anisotropy in seismic-wave velocities within the upper mantle. Assuming classical trench-normal “corner flow” and standard models of crystallographic fabric development, the direction of fast shear-wave polarization will be normal to the arc

in the warmer wedge beneath the arc and back-arc, although the fast shear-wave polarization may be parallel to the arc in the cold corner of the wedge beneath the fore-arc^{5,6}. While this pattern of seismic anisotropy is observed in some subduction zones^{7,8}, many display more variable fast directions further from the trench beneath the arc and back-arc, often with a roughly arc-parallel trend⁹⁻¹². The origin of this arc-parallel-fast anisotropy, in particular if it is related to along-arc flow within the mantle wedge or some other process, is vigorously debated^{5,6,13-16}.

Although arc volcanic rocks from Costa Rica have Ocean-Island-Basalt (OIB)-type compositions similar to those found in Galapagos Hotspot rocks¹⁷, the origin of this geochemical signature is controversial. In some models, OIB-signatures are contained in the mantle wedge beneath Central Costa Rica, reflecting 1) residual Galapagos-type mantle remaining after the formation of the Caribbean Large Igneous Province (CLIP) in the Cretaceous¹ or 2) flow of OIB-type asthenospheric mantle around the northwest margin of South America to the mantle beneath Central Costa Rica² or 3) through a slab window beneath Central Costa Rica³. Alternatively, subduction erosion of older Galapagos and CLIP terranes in the Costa Rican fore-arc may have introduced this signature beneath Costa Rica⁴. The geochemical evidence presented here, however, suggests that the OIB-signature is primarily derived from the subducting Galapagos hotspot track.

Central American volcanism results from subduction of the Cocos Plate beneath the Caribbean Plate (Fig. 1). Normal oceanic crust formed at the East Pacific Rise (EPR) subducts beneath Guatemala to northwestern Costa Rica, whereas crust formed at the Cocos-Nazca Spreading Center (CNS) with the Galapagos hotspot track on it subducts

beneath central Costa Rica to Panama. The Galapagos hotspot tracks (and islands) are chemically zoned^{19,20} with the Seamount Province having distinctly higher $^{208}\text{Pb}/^{204}\text{Pb}$ (Figs. 1,2) and lower $^{143}\text{Nd}/^{144}\text{Nd}$ (Supplementary Information (SI) Section 1, Fig. S1.1) for a given $^{206}\text{Pb}/^{204}\text{Pb}$ isotope ratio compared to the Cocos and Coiba Ridges.

Radiogenic isotope ratios, which are not fractionated by physical processes such as partial melting and magma differentiation, can be used as tracers to determine the sources contributing to arc magmatism. Along the volcanic front (VF) from Central Costa Rica to NW Nicaragua, Nd isotope ratios increase and Pb isotope ratios decrease continuously (Figs. 2a,b). $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ lead isotope data from the VF rocks also form an excellent linear correlation ($R^2=0.997$; Fig. 2c). Although normal arc magmatism ceased in the Late Miocene-Pliocene in southern Costa Rica and western Panama, Pliocene-Quaternary adakitic/alkalic volcanic rocks are widely distributed but volumetrically insignificant^{3,24,25}. Lead isotope data from these rocks also form an excellent linear correlation ($R^2=0.976$) below, but sub-parallel to, the trend formed by VF rocks. The highest Pb isotope ratios in the arc rocks occur in the areas situated above the subducting hotspot track, also characterized by radiogenic Pb.

At least three endmembers are required to explain the Pb and Nd isotope data of the Nicaraguan to Panamanian volcanic rocks (see also Fig S1.1). The NW Nicaraguan endmember with unradiogenic Pb and radiogenic Nd reflects addition of a slab fluid with Pb primarily from the subducting crust, as represented by samples from a subducting seamount, to the depleted mantle wedge (Figs. 1,2). The southern Costa Rica/Panama endmember with radiogenic Pb but intermediate $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ could be derived from the subducting Cocos/Coiba Ridges³ and/or eroded fore-arc igneous

complexes⁴, consisting of CLIP basement and accreted early Cenozoic Galapagos seamounts^{22,23,26}. The only source with the appropriate Pb (radiogenic Pb and high $^{208}\text{Pb}/^{204}\text{Pb}$) and Nd isotopic composition to derive the Central Costa Rica endmember is the Seamount Province subducting beneath Central Costa Rica. Other potential sources for Pb and Nd, such as tectonically eroded Costa Rica fore-arc⁴ and input of Galapagos hotspot mantle through a slab window³ (which would be expected to have a composition similar to the voluminous Cocos/Coiba Ridges – representing the primary composition of the Galapagos plume), have isotopic compositions distinct from the Seamount Province and thus cannot explain the observed mixing trends.

If the Galapagos seamount slab component has melt-like properties (i.e., supercritical fluid/hydrous melt), it could readily transport rare earth (e.g., La, Nd) and high field strength elements (e.g., Nb), consistent with decrease in La/Yb and increase in Ba/La, U/Th and Ba/Th observed in the VF lavas from Central Costa Rica to NW Nicaragua^{2,21,27} (see SI Section 1 for the role of subducted sediments). The transition from a dominantly fluid-like slab component beneath Nicaragua to melt-like slab components in Costa Rica and Panama may be related to the influx of hot mantle through a slab window in the south³.

The along-arc variations in VF isotopic composition indicate that the Galapagos seamount component diminishes in quantity to NW Nicaragua. These trends cannot be produced by upper plate variation. Although CLIP lithosphere underlies Costa Rica, its composition (included in the fore-arc igneous terranes field in Fig. 2c) is distinct from that of the Seamount Province^{22,23}. The VF isotopic trends also cannot be explained by systematic changes in the composition of the subducting plate. During ODP Leg 170,

gabbro sills with OIB-type geochemical characteristics were drilled about 80 km north of the Seamount Province²⁸, suggesting that the Galapagos hotspot may have affected a larger area than represented by the morphological expression of the hotspot track. However, it is unlikely that the plume spread out 500 km north of the Cocos Ridge. When the hotspot track off Costa Rica formed, the CNS was located to the south of the hotspot²⁰, and plume material would have flowed south towards the ridge (not north). In addition, the gabbro sills from Leg 170 do not have Pb isotopic compositions that plot within the Seamount Province, but rather compositions that plot within the Cocos/Coiba Ridge field (Fig. S1.1). Finally, a seamount on the Cocos Plate subducting beneath Central Nicaragua shows no evidence for the presence of Galapagos-type material in this area²⁰ but rather has the appropriate composition to serve as the depleted Nicaraguan endmember.

Since the seamount component only appears to be present on the incoming plate outboard of Central Costa Rica, this signature must have been introduced into the wedge by a melt-like slab component from the subducting Seamount Province and then transported NW in the mantle wedge in melt pockets (possibly crystallized to form pyroxenite after leaving the slab and reacting with the overlying mantle wedge peridotite). Fluids from the subducting plate beneath Nicaragua flux the wedge causing melting and mixing with the Galapagos seamount component. As the mantle is transported northwestwards, the seamount component will be progressively flushed out of the wedge, causing a decrease in Pb and increase in Nd isotopic composition.

Seismic anisotropy in the mantle wedge beneath Nicaragua and Costa Rica provides corroborating evidence for arc-parallel flow. Shear-wave splitting was measured in local

S phases recorded by a dense temporary deployment of broadband seismometers (Fig. S2.1), and a three-dimensional model of anisotropy was obtained (Fig. 3) by tomographically inverting the splitting measurements²⁹. Beneath the arc where the Galapagos geochemical signature is observed and further into the back-arc, anisotropy is dominated by roughly arc-parallel alignment of the olivine fast symmetry axis (*a*-axis). These *a*-axes extend well into the warmer wedge where they should align roughly parallel to flow⁶. A zone of arc-normal *a*-axes appears beneath the NW end of the Nicaraguan arc where geochemical evidence for the Galapagos seamount component is nearly absent, perhaps suggesting a reduction in arc-parallel flow.

The isotope data allow an along-arc flow rate for the mantle in the wedge to be directly estimated. Given that samples from the Miocene Nicaraguan arc (7-24 My) adjacent to Quaternary SW Nicaraguan rocks (Fig. 1) have Pb isotope ratios that overlap the range in the Quaternary NW Nicaraguan VF (Fig. 2), the component with elevated Pb isotope ratios was presumably introduced into the mantle beneath Nicaragua after the Miocene. Igneous rocks from the Miocene arc (6-26 My) in Costa Rica have slightly more radiogenic isotopic compositions than those from Nicaragua, which could reflect interaction with CLIP lithosphere or derivation from subduction of older Galapagos hotspot tracks²⁶. The Seamount Province composition however is not observed in the Miocene volcanic rocks, providing a maximum age for the appearance of this component of ~6 My. Assuming that the oldest part of the subducted Seamount Province lies along its boundary with the Panama Fracture Zone (Fig. 1), it would have passed beneath the Central Costa Rican VF 2-3 My ago²⁵ (reversing subduction at a rate of 85 mm/yr),

providing a minimum age of 2 My for the appearance of the seamount component beneath the arc.

These dates (2-6 My) imply that the minimum arc-parallel wedge flow ranges from 63-190 mm/yr, assuming transport of the Galapagos seamount component 380 km. This distance is measured from the northwesternmost Nicaraguan volcano (Casitas) that shows a clear geochemical influence of the subducting Seamount Province (more radiogenic Pb and less radiogenic Nd than Cosigüina Volcano and the crust outboard of Nicaragua) to the maximum Pb isotopic composition (greatest amount of Galapagos seamount component) in central Costa Rica. These transport rates are on the order of the Cocos Plate subduction rate (~85 mm/yr off Costa Rica), suggesting that lateral (arc/trench-parallel) flow in the mantle wedge can compete with flow entrained by subduction.

3. Conclusions

Lead isotope data demonstrate that rapid arc-parallel flow occurs in the mantle wedge beneath Costa Rica and Nicaragua, allowing arc-parallel-fast anisotropy to be explained without atypical mechanisms of generating anisotropic fabrics^{5,14}. Trench rollback, variations in slab dip, and arc-parallel shearing within the upper plate may all contribute to along-arc wedge flow^{9,13,16}. In Costa Rica and Nicaragua, the fore-arc is translating to the NW, although at less than 15 mm/yr³⁰. Westward migration of the volcanic front since the early Miocene was greatest in Nicaragua and Honduras/El Salvador¹⁸ (Fig. 1) where slab dips are steepest, suggesting differential slab roll-back that could draw wedge material towards NW Nicaragua possibly from the slab window to the south. Collision (indenture) of the thickened Cocos Ridge crust with the margin could also have helped

drive arc material northwards. Regardless of the mechanism for generating flow, this study demonstrates a rate of arc-parallel flow beneath Costa Rica and Nicaragua that rivals downgoing Cocos plate motion.

4. Methods

4.1. Sample Preparation and Isotope Analyses

Samples used in this study were cleaned of surficial dirt and then coarsely crushed into chips smaller than 4 cm. The chips were sieved and washed with deionized water, followed by ~12 hrs drying in a warm (50°C) oven. Approximately 30-60 g of clean chips were hand-picked under a binocular microscope and ground to sand-sized particles in a mechanized agate mortar and pestle, followed by fine grinding in an agate mill by a shatter box. A subset of 0.5-1.0 mm sized chips was saved for Pb isotope analyses. Sample dissolution and element chromatography were carried out at the Leibniz Institute of Marine Sciences IFM-GEOMAR in Kiel (Germany) in Class 1000 clean rooms that are equipped with Class 100 laminar flow hoods. All reagents used were either double distilled in a PicoTrace Teflon distillery (HCl and HNO₃) or certified ultra pure HF and HBr acids from SEASTAR[®]. An ELGA[®] purifying system provided 18.2MΩ water. Nd and Pb isotope analyses were carried out on whole rock powders (Nd) and rock chips (Pb) that were leached in hot 2N HCl (70°C, 1h) for Pb and Nd in order to minimize the effects of alteration and sample handling. About 100mg of sample were weighed into a Teflon beaker and then dissolved for 2 days in a 5:1 mixture of ultra pure HF and HNO₃ at 150°C.

The ion chromatography followed established standard procedures³¹. These include a two-pass Pb separation and clean-up using 100 μ l Teflon micro-columns filled with BIORAD[®] AG 1x8 (100-200 mesh) resin that is equilibrated with 1 M HBr for highest Pb retention and from which Pb is released with 1ml of 6 M HCl. The rare earth elements (REEs) are obtained in 6 ml 6 M HCl at the final washout during Rb-Sr separation on quartz glass columns filled with BIORAD[®] AG50W-X8 (100-200 mesh) resin. The REE's are then loaded in 0.25 M HCl onto 4 ml quartz glass columns filled with EICHROM[®] Ln-Spec resin (100-150 μ m) to obtain the Nd fraction. Pb and Nd isotopic ratios were determined by thermal ionization mass spectrometry (TIMS) at IFM-GEOMAR on a MAT262 RPQ²⁺ and TRITON TIMS respectively. Both instruments operate in static multi-collection mode. Nd isotopic ratios are normalized within each run to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ and all errors are reported as 2 sigma of the mean. Over the course of the study, La Jolla produced $^{143}\text{Nd}/^{144}\text{Nd} = 0.511846 \pm 0.000005$ ($N=49$) and our in-house Nd-monitor SPEX $^{143}\text{Nd}/^{144}\text{Nd} = 0.511711 \pm 0.000006$ ($N=39$). Nd replicate analyses (separate digests) of 4 samples were within the external errors of the standards. The long-term reproducibility of NBS 981 ($N=184$) is $^{206}\text{Pb}/^{204}\text{Pb} = 16.899 \pm 0.007$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.437 \pm 0.009$, $^{208}\text{Pb}/^{204}\text{Pb} = 36.525 \pm 0.028$. Pb isotope ratios are normalized to published NBS 981 values³². Pb replicate analyses (separate digests) of 13 samples were on the average better than 0.022%/amu. Total chemistry blanks were <100 pg for Nd and Pb and thus are considered negligible.

4.2. Shear-wave Splitting Analysis and Tomography

We analyzed shear-wave splitting in local S phases recorded by the TUCAN Broadband Seismic Experiment and by three permanent stations, maintained by the GSN (IRIS), GEOSCOPE (France), and GEOFON (Germany). Over the 20-month deployment, we obtained 791 high-quality splitting measurements (Fig. S2.1) using an eigenvalue minimization method³³. We invert these local S splitting measurements for a 3-D model of anisotropy (Figs. 3 and S2.3).

Here we briefly highlight the key aspects of this technique; the method is described in detail in (29). The elastic coefficients of the anisotropic model are assumed to be an average of olivine and orthopyroxene³⁴⁻³⁶. At crustal depths, olivine *a*-axes in the model should be taken as a proxy for anisotropy of crustal origin (e.g., stress-induced cracks or deformation fabrics in crustal mineralogies). The model is divided into blocks (25 km per side) and parameters in each block are the azimuth and plunge of the olivine *a*-axis and a scalar value that controls the strength of anisotropy relative to its single-crystal value. The model in Figures 3 and S2.3 assumes hexagonal symmetry for the anisotropy, but inversions assuming orthorhombic symmetry yield similar results.

The inversion is conducted using an iterative, damped least squares approach. The starting model is constructed by prescribing the average fast direction for all phases that sample a block as the olivine *a*-axis azimuth in that block. The fabric strength in the starting model is based on the average splitting times from each ray within a block relative to that predicted from the single-crystal elastic coefficients. Synthetic splitting is calculated by propagating, for each ray, an initially linearly polarized wavelet through successive anisotropic blocks along its path and accruing splitting from each block until

the phase reaches the station³⁷. Shear-wave splitting parameters are determined from the resulting waveform using the same method applied to the data³³ and then used to calculate data-model residuals and partial derivatives.

Inversions were performed using a wide variety of starting models, block sizes and damping parameters. Even when optimal inversion parameters used, some model blocks in the back-arc remained poorly resolved because of reduced ray coverage. Therefore, effective block size is made larger in back-arc regions in order to allow for interpretation of a larger portion of the model space, albeit at coarser spatial resolution. The size of these larger (non-cubic) model volumes is variable, but typically 50 km thick and between 50 km and 100 km in each horizontal direction (Figure S2.4).

To test the accuracy with which the *a*-axis orientation and fabric strength can be retrieved, inversions of synthetic data predicted by a series of anisotropic target models were carried out using the source-station distribution for the TUCAN shear-wave splitting dataset²⁹. For target models that vary laterally in better-sampled model regions (beneath the fore-arc, arc, and regions of the back-arc near the dense arc-normal lines of stations) with uniform anisotropy elsewhere, *a*-axis orientation is retrieved to within 25° on average; *a*-axis azimuth is, in general, recovered at high accuracy down to at least 100 km depth, and *a*-axis dip recovery is typically good to roughly 75 km. Lateral variations at spatial scales of 50-75 km can be imaged beneath the arc, but retrievable spatial variations are obviously coarser in back-arc regions where individual blocks are grouped into larger volumes (Figure S2.4).

5. Supplementary Information

5.1 The Role of Sediments

The Be and Th isotope systematics of some of the Central Nicaraguan magmas show evidence for the involvement of subducted sediments³⁸, yet evidence for subducted sediments are not clearly seen in the Pb isotopes. This is likely due to different behavior (decoupling) between highly fluid-mobile Pb and relatively fluid-immobile Be and Th and suggests that fluid-mobile Pb is flushed out of the sediments at shallower depths than fluid-immobile Be and Th. *Rüpke et al.*³⁹ point out that fluids derived from serpentinite on the downgoing slab beneath Nicaragua are likely to be highly reactive and could mobilize Be (and possibly Th) as they pass through the overlying subducted sediments. The excellent linear correlation on the radiogenic Pb isotope diagram (Fig. 2c) indicates that just two components are necessary to explain the Pb isotopes for Costa Rica and Nicaragua volcanic front volcanoes. Mixing between three components would not generate a binary mixing array unless two components have identical compositions, which is not the case with sediments and depleted MORB-type crust. The other alternative is that one of the components forming the binary array is itself a mixture of multiple components, for example, that we have an integrated slab fluid coming off the slab under Nicaragua containing Pb from the magmatic crust and sediments. This integrated slab fluid could then mix with the mantle wedge containing the Galapagos seamount component; however, due to the very unradiogenic Pb isotopic composition of the depleted endmember of the array (seen at Cosigüina and the incoming plate crust) an integrated slab fluid could not contain much sediment Pb.

Subducting sediments are also likely to be the main source of fluid-mobile elements such as Ba and U^{23,38}. Variations in ratios of fluid-mobile to fluid-immobile elements (e.g. Ba/La, U/Th and Ba/Th) in northwest Nicaragua are likely to reflect variable fluid fluxes from the subducting sediments beneath northwest Nicaragua.

5.2. Shear-wave Splitting Observations

The TUCAN Broadband Seismic Experiment, funded by the NSF-MARGINS program, was in the field from July 2004 through March 2006, and included 48 broadband stations (borrowed from the IRIS/PASSCAL program) in a fore-arc, back-arc, and two cross-arc lines with station spacing of 10-50 km (Fig. S2.1). The very dense cross-arc lines provide excellent resolution in the mantle wedge beneath Costa Rica and Nicaragua.

We measured well-resolved shear-wave splitting parameters in 791 local S waves recorded by the TUCAN array and three permanent global stations (IRIS/GSN station JTS, GEOFON station BOA, and GEOSCOPE station HDC). These data show strong variations in both fast direction and splitting time, but when measurements are broken down in terms of similar paths, coherence emerges, making this dataset a good candidate for tomographic inversion. For example, shallow events that sample the wedge corner produce splitting in which fast directions are arc-normal and then rotate to arc-parallel towards the volcanic arc (Fig. S2.2a). In addition, the majority of measurements to stations in the back-arc line (Fig. S2.2b) show fast directions which are roughly parallel to the arc, with the exception of a prominent group of arc-normal fast directions in the northwest portion of the array and sparser arc-normal measurements in the southeast.

The three-dimensional anisotropy model shown in this study (Figures 3 and S2.3) is one realization of a damped, linearized, least squares inversion of the shear-wave splitting data. This model minimizes data-model misfit, and contains structure characteristic of most of the inversions obtained during testing (i.e., with different damping, starting model, crystallographic symmetry). Effective block size in the model is intentionally variable to account for less dense sampling of back-arc and deeper model regions. This is accomplished by requiring model perturbations to be equal in certain groups of model blocks (outlined in black in Figure S2.4) in later inversion iterations.

Anisotropy with largely arc-parallel *a*-axes dominates in much of the mantle wedge (depths of roughly 50-150 km), although a zone of arc-normal *a*-axes occurs in the wedge beneath the NW end of the array (right in Fig. S2.3). Arc-parallel anisotropy is also present in the deepest model layer, but lies too close to the slab interface to be clearly attributed to the wedge. Assuming typical development of crystallographic orientation, *a*-axes should align roughly parallel to flow in the wedge beneath the arc and back-arc. In contrast, olivine *a*-axis alignment normal to the flow direction has been observed experimentally at hydrated, low temperature, and high stress conditions^{5,40}, but should be a factor only in the cold wedge corner⁶ trenchward of the slab surface at roughly 70 km depth. In our model, the arc-parallel *a*-axes extend into the mantle wedge well beyond the shallow wedge corner where the B-type olivine fabric may exist⁶. Corroborated by the Pb isotope data, we associate this overall pattern of anisotropy with along-arc flow towards the NW, rather than with anomalous crystallographic preferred orientation fabrics influenced by melt¹⁴. The rotation of the along-arc *a*-axes to a more arc-normal orientation at the NW end of the arc occurs where geochemical evidence for the

Galapagos seamount component is nearly absent. These a -axes may represent a transition to more arc-normal flow in the wedge, although this hypothesis cannot be fully tested with existing data due to a drop-off in sampling at the NW end of the TUCAN array.

6. Acknowledgements and Author Contributions

6.1. Acknowledgements

Reviewers (J Walker, S Schwartz and one anonymous), J Phipps-Morgan, T Plank, R Werner, M Portnyagin, members of SFB574 and MARGINS are thanked for comments/discussions that helped significantly improve the manuscript. The IRIS PASSCAL program provided seismometers and technical assistance to the TUCAN experiment. This research was supported by the German Science Foundation Collaborative Research Center (SFB574, Contribution Nr. 119) and the National Science Foundation MARGINS Program.

6.2 Author Contributions

Hoernle, Abt and Fischer collected samples/data, processed and interpreted the data, developed the ideas and wrote the paper, with significant input from Abers and discussions with other co-authors. Nichols, Hauff and Heydolph generated the geochemical data and van den Bogaard age data. Abers, Protti and Strauch were key in collecting the seismic data; van den Bogaard, Alvarado and Nichols assisted in the collection of the volcanic samples. Central American partners provided invaluable geological overviews and logistical support.

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8. List of Figures and Tables

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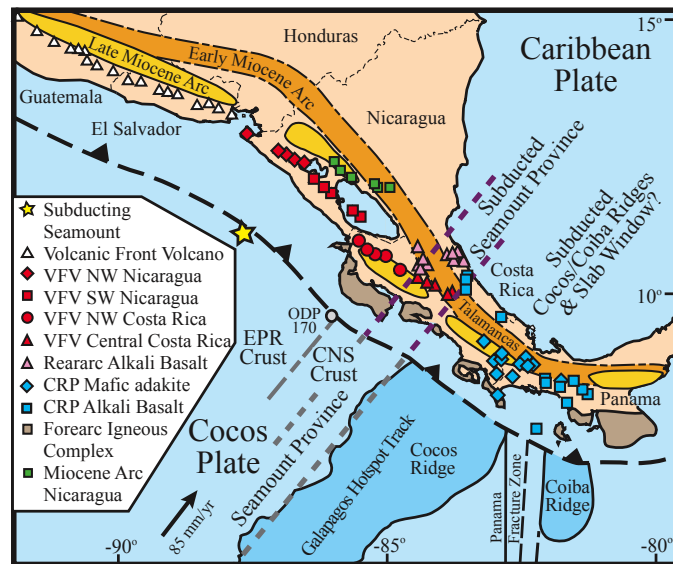


Figure 1. Map of Central America illustrating the westward migration of the volcanic arc in NW Costa Rica to SW Guatemala from the Early Miocene (11-24 Ma) to Late Miocene (6-12 Ma) to the Quaternary¹⁸. The Seamount Province of the Galapagos hotspot track is subducting beneath the Central Costa Rican volcanic front and Late Pliocene-Holocene (0-2 Ma) rear-arc alkali basalts (NE occurrences dated at 2 My). The Cocos and Coiba Ridges of the Galapagos hotspot tracks are subducting beneath Pliocene-Holocene mafic adakites and alkali basalts from SE Costa Rica and western Panama (CRP), where there has been no volcanic front volcanism since the Late Miocene. The heavy dashed purple lines project the Seamount Province beneath Costa Rica. The boundary between crust formed at the East Pacific Rise (EPR) and the Cocos Nazca (Galapagos) Spreading Center (CNS) and the Panama Fracture Zone are also shown. Volcanic front volcanoes (VFV) are indicated. A window in the subducting Cocos and Nazca Plates may be located beneath southern Costa Rica and western Panama³.

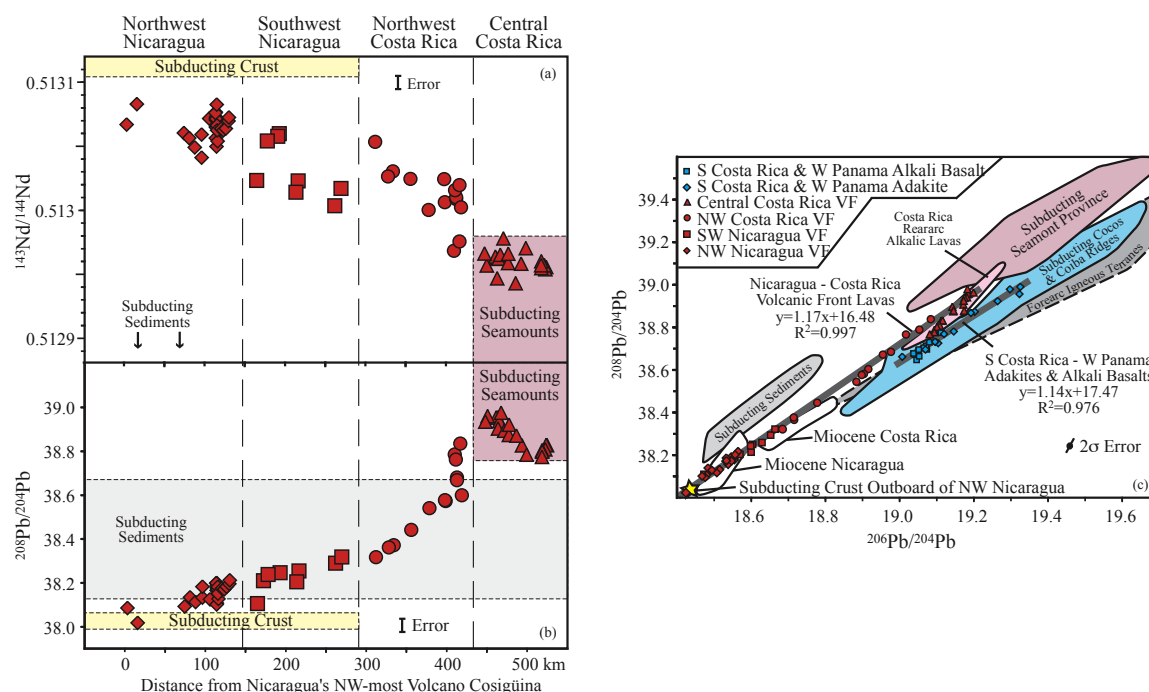


Figure 2. Systematic variation in lead and neodymium isotopic composition along the volcanic front (VF) from Costa Rica to northwest Nicaragua indicate northwest flow of mantle wedge material. Isotope ratios, (a) $^{143}\text{Nd}/^{144}\text{Nd}$ and (b) $^{208}\text{Pb}/^{204}\text{Pb}$, show systematic trends along the arc that indicate mixing of mantle wedge enriched with melts from the subducting Galapagos Hotspot Seamount Province with mantle wedge enriched with hydrous fluids from the ocean crust subducting beneath Nicaragua. In (c) Quaternary VF lavas from Nicaragua and Costa Rica form an excellent linear correlation (longer solid gray line, $R^2=0.997$) on the thorogenic Pb isotope diagram. Late Pliocene-Quaternary rear-arc alkalic lavas in NE Costa Rica overlap in composition with Costa Rica VF lavas but extend to more radiogenic compositions, consistent with a greater contribution from a melt-like slab component. Low volume, scattered Pliocene-Quaternary mafic adakites and alkali basalts from S Costa Rica and W Panama, where there has been no VF volcanism since the Late Miocene, form a distinct linear array (shorter solid gray line, $R^2=0.976$) below but subparallel to the VF lavas, having lower $^{208}\text{Pb}/^{204}\text{Pb}$ for a given $^{206}\text{Pb}/^{204}\text{Pb}$ isotope ratio. While the adakites may represent direct melts of subducted Galapagos rocks, the alkali basalts most likely reflect interaction between C-rich fluids/melts from Galapagos material and the overlying mantle wedge. The difference in Pb isotopic composition between the Pliocene-Quaternary central Costa Rica and the southern Costa Rica/western Panama lavas correlates closely with the boundary between the subducting Seamount Province and Cocos/Coiba Ridges of the chemically-zoned Galapagos hotspot track. The subducting Seamount Province and the Cocos/Coiba Ridges, which have distinct Pb and Nd (Fig. S1.1) isotopic compositions^{19,20}, could serve as the endmembers necessary for generating the distinct isotopic compositions of the central Costa Rica and southern Costa Rica/western Panama lavas, respectively. Fields for 1) fore-arc igneous terranes, which includes CLIP basement from Costa Rica and Panama²¹⁻²³ and 2) sediments and ocean crust outboard Nicaragua^{1,20} are also shown.

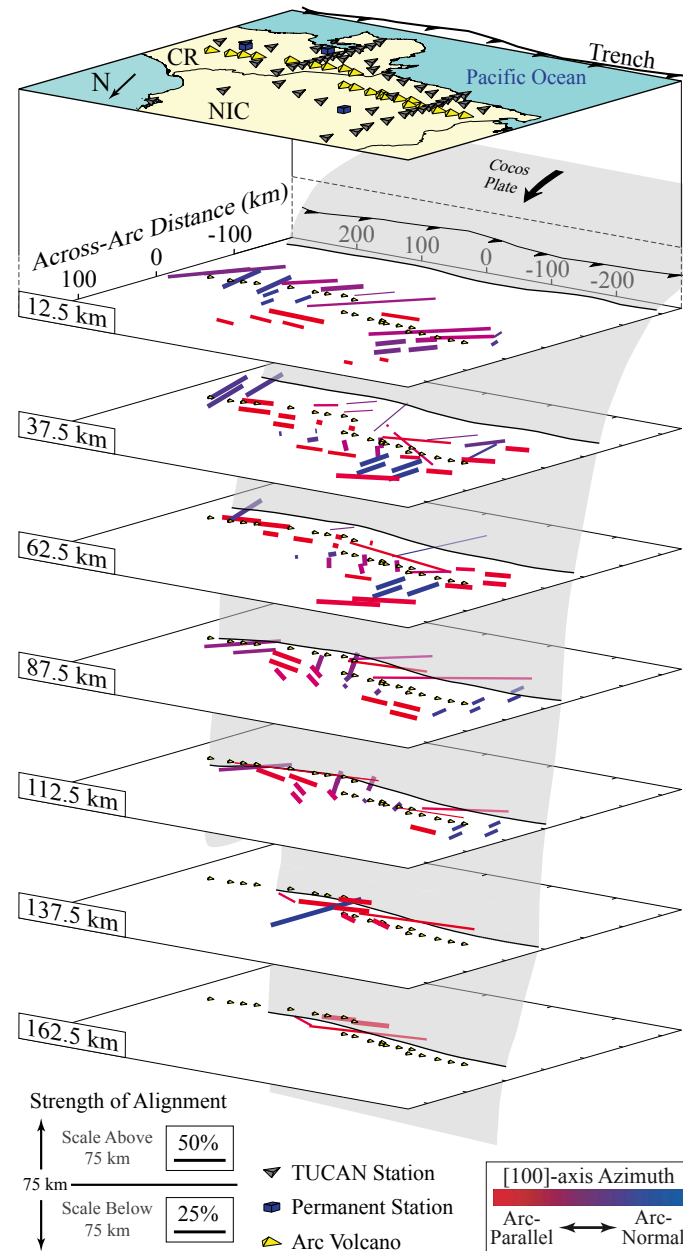


Figure 3. Model of anisotropy obtained by inverting shear-wave splitting measurements from events in the Nicaragua-Costa Rica subduction zone using a tomographic approach²⁸ (Methods and SI Section 2). The vectors represent well-resolved olivine *a*-axes in an olivine-orthopyroxene model, with orientation and color indicating horizontal azimuth. Vector length corresponds to the strength of anisotropy relative to single crystal values; note the change in strength scale at 75 km depth. We also invert for *a*-axis dip, but for clarity here we project *a*-axes onto map-view planes. Effective block size in the back-arc is larger. Only well-resolved regions of the model are shown, and vector thickness corresponds to model parameter resolution. The TUCAN seismic array and volcanic arc are shown at the surface, and for reference, the volcanic arc is plotted on each slice through the model. The slab-wedge interface, based on earthquake locations, is shown by yellow shading. The top two layers primarily lie within the upper plate; the mantle wedge is located in front of the slab in the layers spanning 50-175 km. Roughly arc-parallel *a*-axes dominate well-resolved wedge regions beneath the arc and back-arc at depths of 50-150 km, with the exception of a zone of arc-normal *a*-axes at the NW end of the arc in Nicaragua and one isolated model block deep beneath the back-arc.

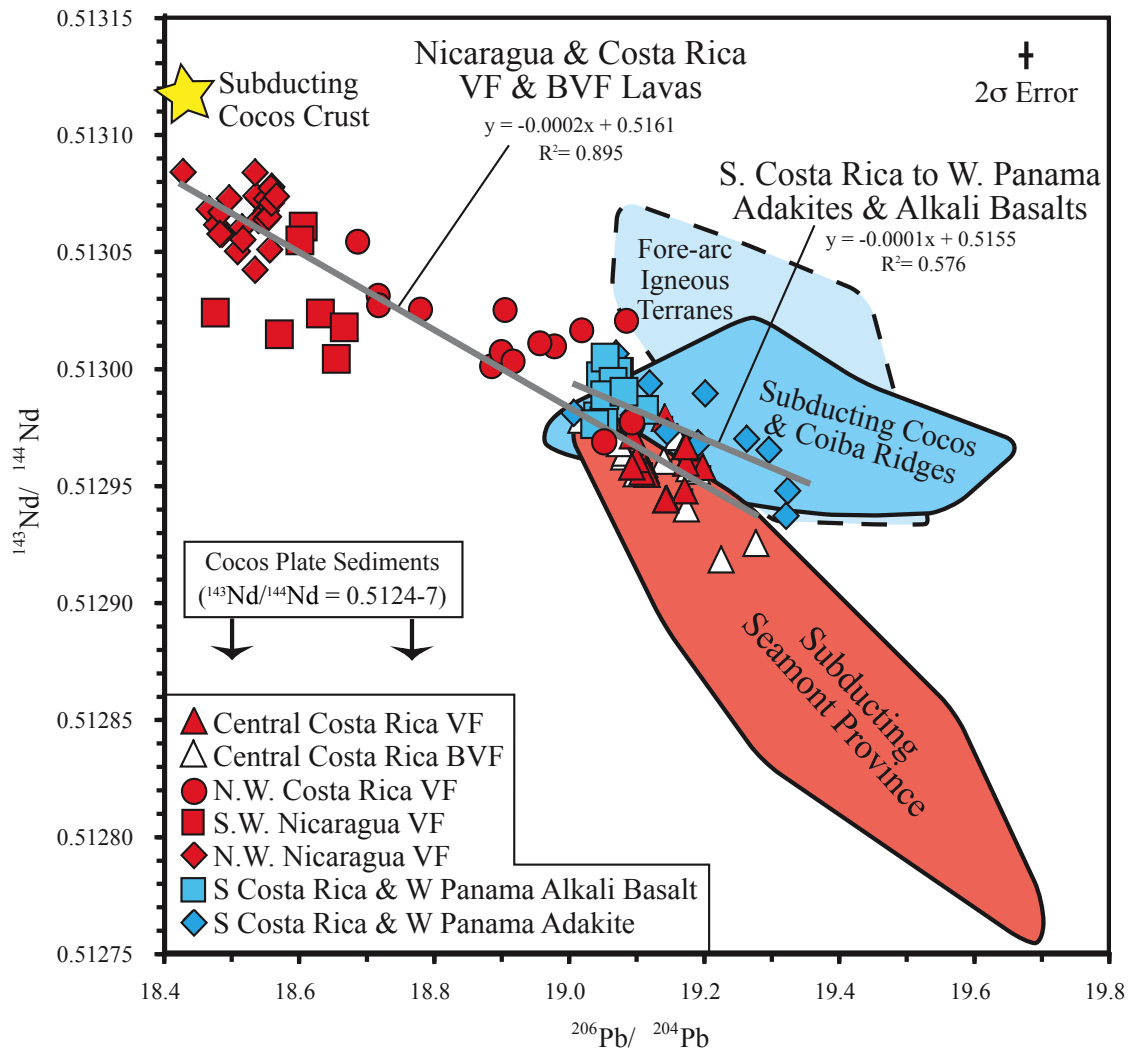


Figure S1.1. Pb vs Nd isotope data for Costa Rica and Nicaragua volcanic front (VF), Costa Rica behind the volcanic front (BVF) alkalic rocks, and southern Costa Rica and western Panama adakites and alkali basalts. At least three endmembers are required to explain the Pb and Nd isotope data of Central American volcanic rocks. The NW Nicaraguan endmember with unradiogenic Pb and radiogenic Nd can be explained by mixing slab fluids from subducting EPR oceanic crust (+/- sediments) with depleted mantle wedge. The southern Costa Rica/western Panama endmember with radiogenic Pb but intermediate $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ could be derived from the subducting Cocos/Coiba Ridges and/or eroded fore-arc igneous complexes. The only source with the appropriate Pb (radiogenic Pb and high $^{206}\text{Pb}/^{204}\text{Pb}$) and unradiogenic Nd isotopic composition to derive the central Costa Rica endmember is the Seamount Province subducting beneath Central Costa Rica. Other potential sources for Pb and Nd, such as tectonically eroded Costa Rica fore-arc and input of Galapagos hotspot mantle through a slab window (which would be expected to have a composition similar to the voluminous Cocos/Coiba Ridges – representing the primary composition of the Galapagos plume), have isotopic compositions distinct from the Seamount Province and so cannot explain the observed mixing trends. See Figure 2 caption for additional information and data sources.

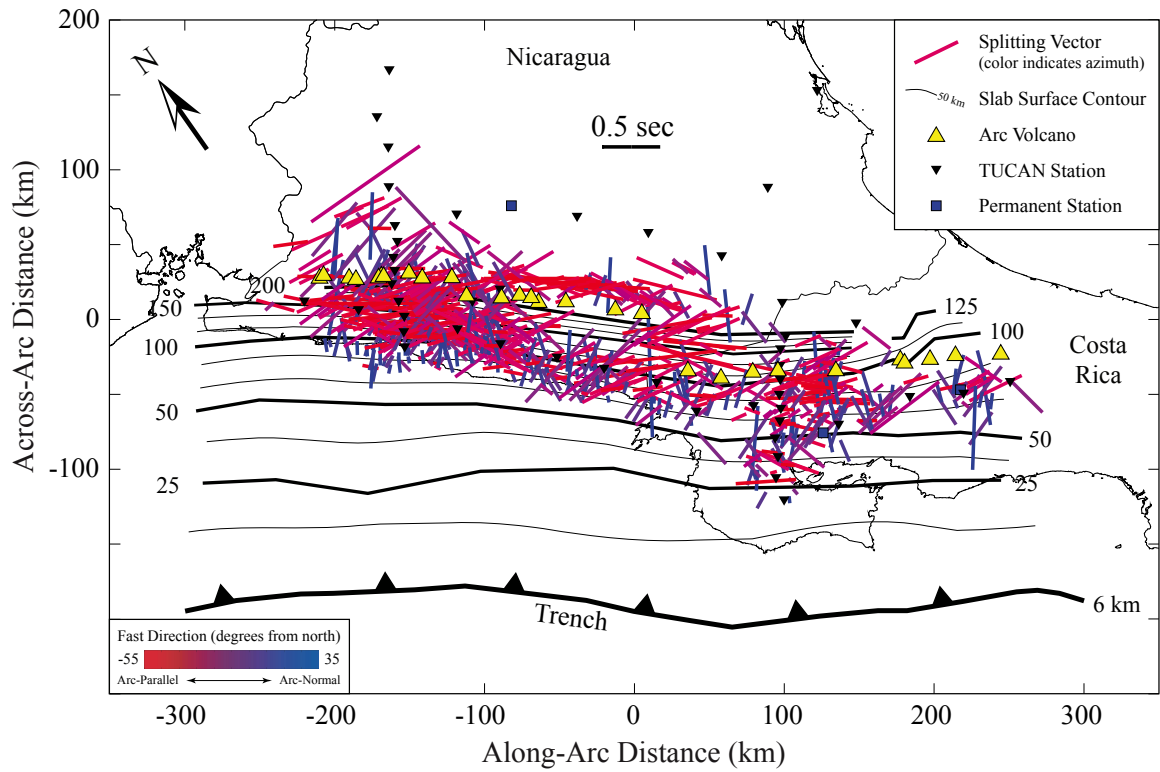


Figure S2.1. Shear-wave splitting measurements, plotted at ray-path midpoints, that were used to invert for the model of anisotropy shown in Figures 3 and S2.3. The vectors overlap, with splitting measurements from deeper events plotting below those from shallow events. The color scheme is used to help distinguish between different fast direction azimuths. When viewed all together, a dominant orientation is difficult to discern, except in the central Nicaraguan back-arc where arc-parallel fast directions are more prevalent.

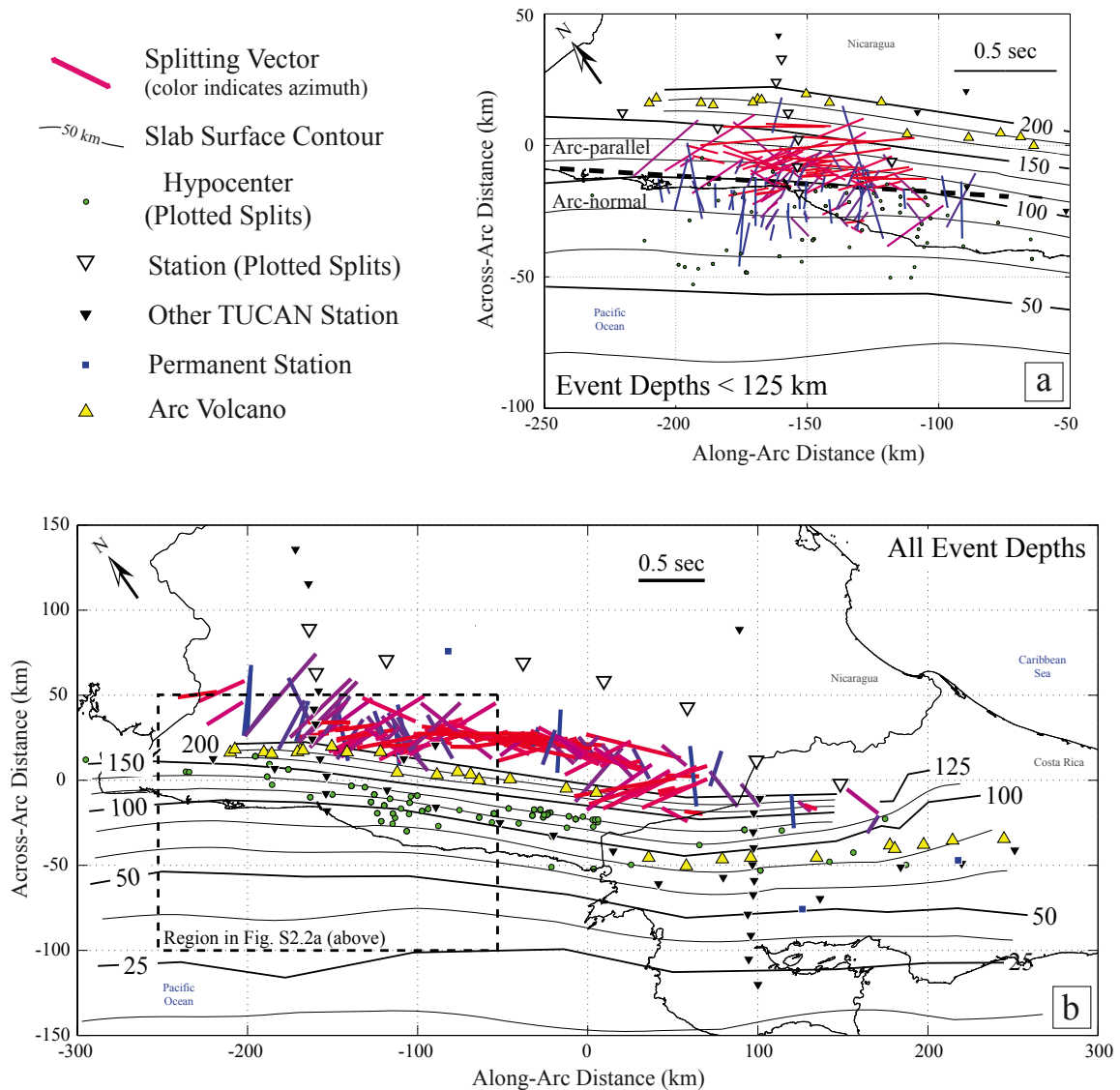


Figure S2.2. Subsets of the shear-wave splitting measurements show greater coherence than the entire dataset (Fig. S2.1). The larger inverted empty black triangles represent the stations at which the splits shown were measured. **(a)** In NW Nicaragua, for splits from shallow (<125 km) events, a relatively sharp transition (thick dashed line) occurs between arc-normal fast direction closer to the trench and arc-parallel fast directions towards the arc. This pattern is predicted for along-arc flow and B-type fabric in the cold wedge corner^{5,6}, but is the inverse of what would be expected for 2-D corner flow (i.e., arc-parallel close to the trench and changing to arc-normal in the arc and back-arc). **(b)** Across the central Nicaraguan back-arc region, fast directions are strongly arc-parallel, while more oblique fast directions are observed to the NW and SE.

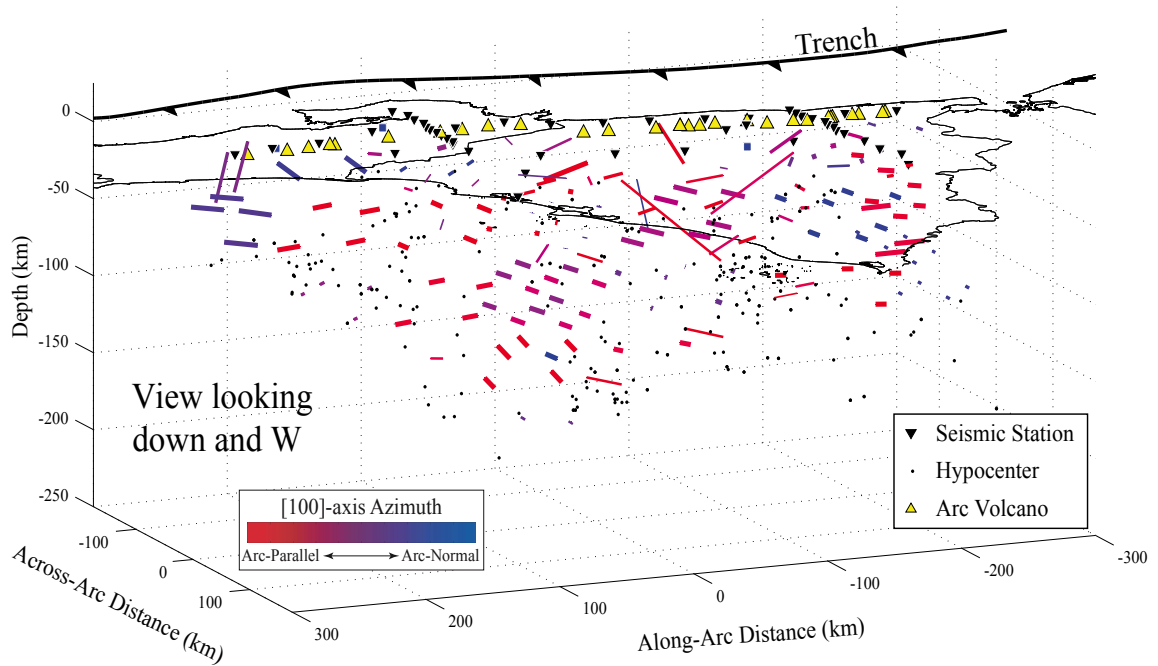


Figure S2.3. Three-dimensional model of anisotropy obtained by inverting local S splitting measurements. Each vector shows the orientation of the olivine a -axis in an olivine-orthopyroxene mantle, and its length indicates the strength of anisotropy relative to its single crystal value. Blocks are 25^3 km cubes, although blocks in back-arc regions are made effectively larger by setting their perturbations equal in the later stages of the iterative inversion. Vector thickness corresponds to formal resolution values obtained for the parameters in each block; results are shown only when the resolution of each model parameter in a block exceeds 0.5. Anisotropy vectors at depths of 50-150 km largely lie in the mantle wedge and are roughly oriented parallel to the arc, except in NW Nicaragua.

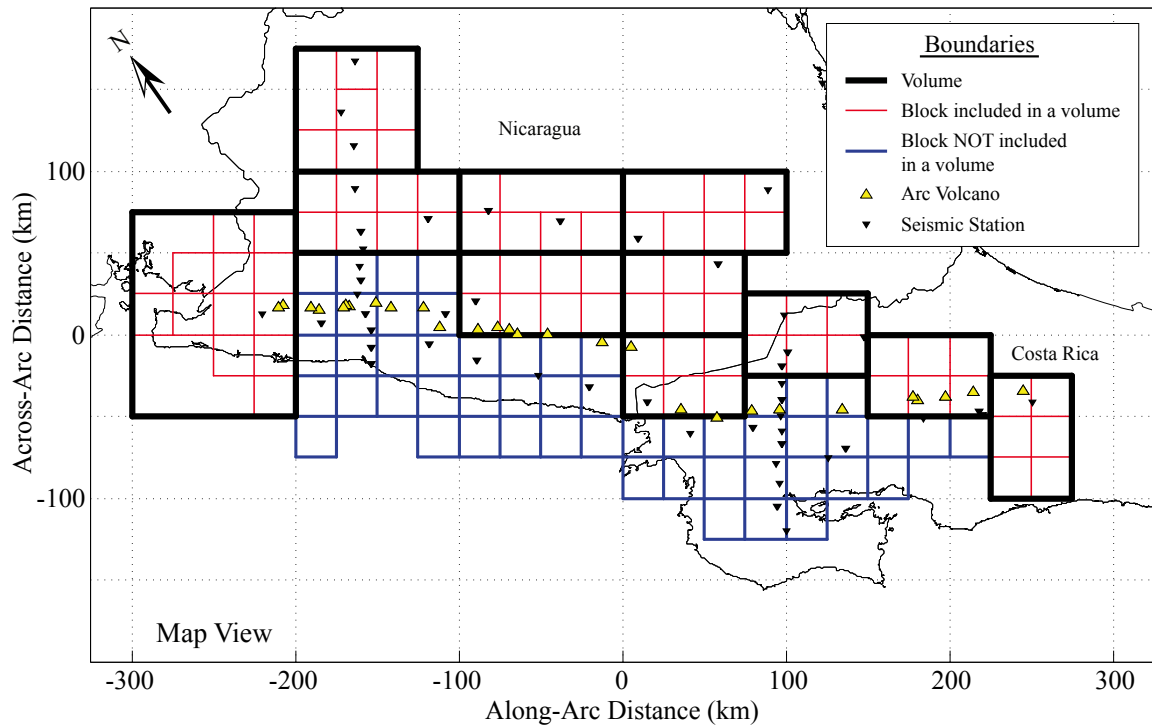


Figure S2.4a. Map-view of individual model blocks (red and blue) and larger back-arc model volumes (black) used in the inversion. Beneath much of the fore-arc and arc, model parameters in individual 25^3 km model blocks are well-resolved (blue). Individual model blocks that lack sufficient resolution for interpretation (red) are found in the back-arc regions of the model space, and we force many of these blocks to act as larger model "volumes" (outlined in black). This is accomplished by determining average model parameters for all blocks in a volume at an intermediate iteration, and then requiring that model parameter changes are equal for all blocks in volume in later iterations. The coarsening of model discretization in this manner produces better resolved model parameters for the volume, and allows us to confidently interpret the model in a larger portion of the mantle wedge, although the minimum spatial scale at which anisotropy can be resolved is increased. The lateral extent of these larger volumes does not vary in depth, however the actual number of blocks in each that are sampled by ray paths in the dataset does. In the map view shown here, only blocks sampled by ray paths in the dataset are shown, and in the models resulting from the inversion (Figures 3 and S2.3), only a -axes in well-resolved model blocks are displayed.

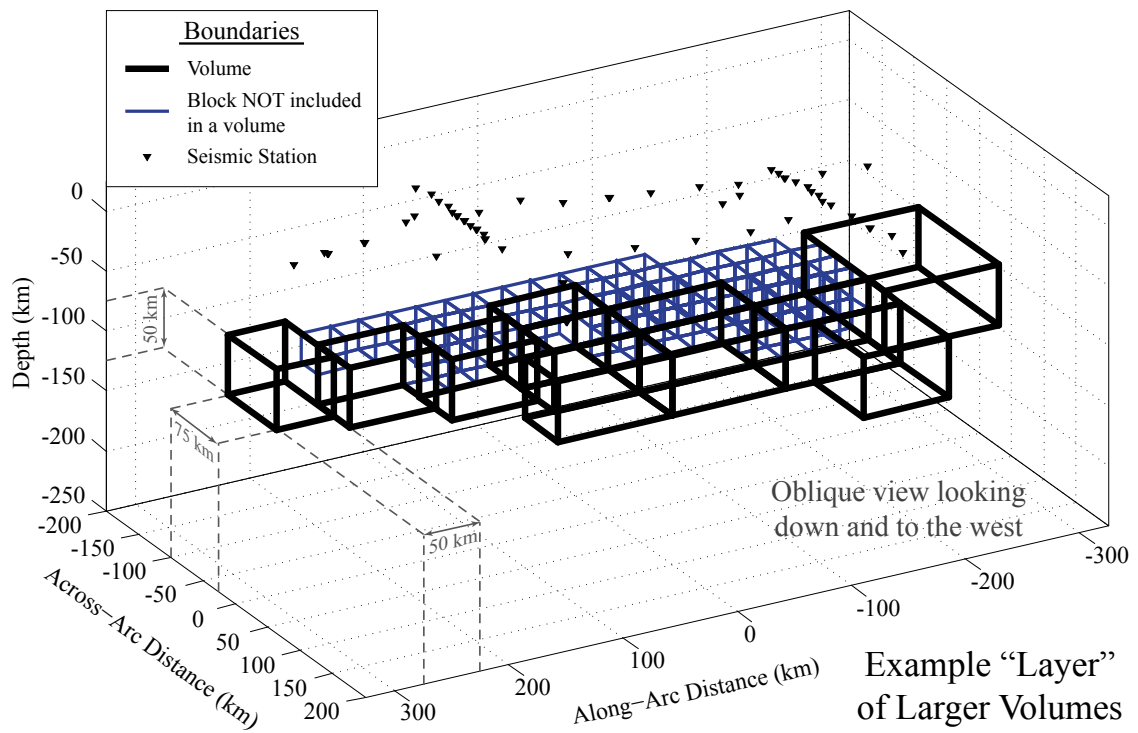


Figure S2.4b. Three-dimensional view of individual model blocks and larger back-arc model volumes used in the inversion. Except for the top layer of 25^3 km model blocks, the larger volumes are each 50 km thick (i.e., containing two layers of blocks) and have different lateral dimensions. The far left block, for example, is 50 km by 75 km. In this view the red blocks are not shown in order to improve visibility of the independent (blue) blocks.