

Lithospheric and Asthenospheric Seismic Anisotropy at Collisional and
Passive Margins

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Preface

This thesis represents an exploration of the causes and indications of seismic anisotropy within the lithosphere and asthenosphere in a variety of tectonic environments. Seismic anisotropy is defined as the directionally-dependent variation of P- and S-wave velocities within a material. Anisotropies at lithospheric and asthenospheric depths are often used as proxies to represent fossilized deformation from when the minerals within the rocks were still at hot enough temperatures to deform and rotate, and the direction of asthenospheric mantle flow, respectively. With these proxies, I have investigated both the frozen-in LPO fabrics and the asthenospheric flow regimes at both compressional and passive margins.

Seismic anisotropy is an important phenomenon to understand. In addition to being a useful constraint in seismic tomography and other global modeling methods, seismic anisotropy itself is vital to quantify because seismic anisotropy indicates the presence of either an LPO (lattice preferred orientation) fabric, or an SPO (shape preferred orientation) fabric. This thesis will be focusing on LPO fabrics, which can refer to any arrangement of anisotropic minerals, be it the alignment of mineral crystals in the flowing asthenospheric mantle (Chapters 1, 2, 4, and 5), or fossilized anisotropy in the lithosphere (Chapters 3, 4, and 5).

Seismic Anisotropy in the Asthenospheric Mantle

Two anisotropic minerals, olivine and orthopyroxene, comprise a significant amount of the asthenospheric mantle mineralogy. The flowing asthenospheric mantle causes these mineral crystals, en masse, to rotate to a preferential alignment, creating LPO fabrics. For instance, in this type of flow, olivine crystals typically rotate, by way of dislocation glide, to a position where their a -axes are parallel to sub-parallel to the direction of mantle flow. When a shear-wave encounters such an anisotropic material, it splits and propagates along two axes: the relatively ‘fast’ and ‘slow’ axes. Assuming that a given shear-wave samples the same type of anisotropy continuously, the arrival of the ‘slow’ wave will be increasingly delayed compared to that of the ‘fast’ wave. A shear-wave splitting analysis constrains seismic anisotropy using two parameters: the fast direction (ϕ ; the orientation of the fast axis, measured from North), and the splitting time (dt ; the difference in arrival times between the ‘fast’ and ‘slow’ shear waves). In the case of the asthenospheric mantle, the fast direction is assumed to be representative of the direction of mantle flow, and the splitting time is representative of the path length through or the strength (the degree of mineral alignment) of anisotropy. Shear-wave splitting is used to assess the direction and strength of mantle flow in multiple tectonic environments in Chapters 1, 2, 4, and 5.

For many years, it was believed that subduction-induced flow in the asthenospheric mantle wedge was two-dimensional in nature; mantle material is drawn in towards the trench, and is then entrained downward with the slab. Only later was it

discovered that mantle flow patterns often deviated from the ideal 2-D corner flow model; Chapters 1 and 2 explore the complex mantle flow above and below a subducting slab potentially induced by along-strike slab dip variation and/or slab rollback using local S and SK(K)S shear-wave splitting measurements. In Chapter 1, I measure, interpret, and model the shear-wave splitting results in the context of the southern South American subduction zone. These data were recorded on temporary CHARGE (2000-2002) and Chile RAMP (2010) arrays, as well as three permanent stations: PEL, PLCA, and TRQA. In Chapter 2, I attempt to recreate the observed shear-wave splitting patterns from Chapter 1 in an analogue subduction tank fluids model where I can impose both variable slab geometries and slab rollback. Both the shear-wave splitting measurements and tank modeling showed that 1) mantle flow three-dimensionality increases with decreasing slab dip, and 2) when a shallowly-dipping slab segment is in close proximity to a slab with a relatively steeper dip, mantle flow is drawn away from the steeply-dipping slab to subduct on the shallowly-dipping slab. Additionally, the presence of a slab gap, as well as slab rollback in the tank models also allowed us to draw the conclusions that 1) penetration of mantle material from beneath the slab into the mantle wedge occurs, but the reverse does not, and 2) slab rollback increases the speeds of all markers within the mantle wedge, including flow through the slab gap. Synthetic shear-wave splitting measurements constrained by laboratory proxies for olivine *a*-axes in the subduction tank model showed that, in the mantle wedge above the slab gap close to the trench, slab rollback tends to generate shear-wave splitting fast directions that are within $\sim 10^\circ$ of trench normal. When slab rollback is absent, the fast directions tend to rotate to an oblique orientation (up to $\sim 30^\circ$ away from trench normal) that points towards the

shallowly-dipping slab, following the aforementioned sense of mantle flow from the steeply-dipping slab to the shallowly-dipping slab. While slab dip variations can cause localized trench-parallel flow in the mantle wedge right above the slab gap, the anisotropy elsewhere in the mantle wedge was not rotated enough to generate the trench-parallel shear-wave splitting observations made in arcs and back-arcs of subduction zones around the world, and the southern South American subduction zone in particular; some other source of anisotropy must be present at subduction zones to fully explain these trench-parallel shear-wave splitting measurements.

Seismic Anisotropy in the Lithosphere vs. Lateral Variation

In Chapters 1 and 2, I found that one of the best ways to constrain the seismic anisotropy of a deeper layer (i.e. sub-slab mantle) was to model the anisotropy of the uppermost layer (i.e. mantle wedge), and then account for that anisotropy in the waveforms that sample both layers (teleseismic SK(K)S). In Chapter 3, I attempted to define the lithospheric anisotropy of the southeastern US. With the dearth of local seismic events in the southeastern US, I used three regional events, the 2011 November 6 Shawnee, Oklahoma ($M_w = 5.6$) event, the 2011 August 23 Mineral Virginia ($M_w = 5.8$) event, and the 17 May 2012 Timpson, Texas ($M_w = 4.8$) event to constrain the seismic anisotropy in the upper mantle by analyzing P-wave velocity anisotropy. These data were recorded on the temporary SESAME (SouthEastern Suture of the Appalachian Margin Experiment) array (2010-2014). By modeling the delay in Pn arrival times at certain SESAME stations (through both arrival time differentiation and cross-correlation), I

determined that slower upper mantle Pn velocities (~ 7.9 km/s) reside beneath the core of the Appalachian Mountains, and that velocities increase (~ 8.4 - 8.5 km/s) to the south. These observations do not require the presence of seismic anisotropy; instead, they can be explained as the results of lateral variation. Two possible models for the low velocity zone beneath the Blue Ridge Province include a zone of buoyant upper mantle material that, together with a root of thickened crust, supports the continued high elevation of the Appalachian Mountains in this area, or an area of partially eclogitized lower crustal material that is a part of the crustal root, but lies below the velocity increase otherwise interpreted as the crust-mantle boundary. While lithospheric seismic anisotropy is not needed to explain our results, it is possible that a small amount of azimuthal anisotropy exists, and is just not resolvable within our uncertainty range. By fitting some of our best-fitting Pn velocity values to sinusoidal 2θ variation curves, I found that if $\sim 9\%$ anisotropy exists, then it has a fast direction roughly parallel to both the strike of the Appalachian Mountains. This analysis gives us an estimate for potential lithospheric seismic anisotropy in this area, which informs some of the anisotropy models for the southeastern US that are discussed in Chapter 4.

Seismic Anisotropy in the Lithosphere vs. Asthenosphere

In the interest of fully describing the seismic anisotropy in the southeastern United States, I conducted a shear-wave splitting analysis using teleseismic SK(K)S phases from the SESAME array and SKS phases from the TA (Transportable Array) in Chapter 4. Both the SESAME and the TA analyses show that the overall fast polarization

direction is NE/SW, which is sub-parallel to the direction of absolute plate motion (APM) and normal to the direction of accretion-induced compression. However, different local fast direction patterns are also evident, and I am able to distinguish between different fossilized anisotropic fabrics that reside in different terranes, such as in the Blue Ridge vs. the Inner Piedmont terranes. In eastern Georgia, where a distinctive variation in splitting with back-azimuth is observed, lithospheric anisotropy may be caused by a combination of fossilized collisional anisotropy of the Carolina Terrane and deformation due to Mesozoic rifting in the same area. Rift-related anisotropy may also be present in the lithosphere of southern Georgia. The data from the TA allows me to see a similar pattern of fast direction variation on a broader regional scale; the distinct fast direction change from E-W in Florida to NE-SW in South Carolina may be caused by the different fabrics of the Suwannee terrane and cratonic Laurentia, respectively. Another possibility is that asthenospheric flow is being actively driven by the presence of a cratonic keel or a remnant slab (the Farallon Plate) at greater mantle depths.

Chapter 5 explores lithospheric and asthenospheric anisotropy in the Hindu Kush-Pamir region, and employs many of the methods and ideas discussed in the previous chapters. Beneath the Hindu Kush and Pamir Mountains, seismological evidence suggests that there are either adjacent subduction zones with oppositely-dipping slabs or one continuous, but highly deformed slab. While the exact nature of the regional tectonics is still highly debated, I incorporated the possibility of two distinct anisotropy layers into our model: the lithosphere and the flowing asthenosphere. These anisotropy models are constrained by shear-wave splitting results from local s , SKS, and SKKS phases recorded

at stations from the Tajikistan National Seismic Network. My results indicate that the asthenospheric mantle is toroidally responding to the transport of the subducting slabs and overlying continental lithosphere with absolute plate motion. The lithospheric anisotropy models are consistent with fabrics aligned due to internal deformation (i.e. compression) that is primarily driven by the continued thrusting of India into the Eurasian continent.

Seismic anisotropy, a phenomenon that occurs at both small and large spatial scales, exists and is generated at both tectonically active and inactive locations. Seismic anisotropy is important to quantify; it can yield valuable information about the tectonic evolution and/or the current tectonic regime of a material within the lithosphere or the asthenosphere. In this thesis, I have examined the supra- and sub-slab asthenospheric flow at a subduction zone; variable slab geometries as a method of generating that supra-slab asthenospheric flow; upper mantle, lithospheric, and asthenospheric anisotropy at a former continental collision zone that is now a passive margin, and an ongoing continental collision example utilizing the theories, methods, and ideas from the previous four projects.

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Chapter 1

Seismic Anisotropy Above and Below the Subducting Nazca Lithosphere in Southern South America

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Abstract

The goal of this study is to better constrain anisotropy and mantle flow above and below the Nazca slab from 28°S to 42°S through modeling of shear-wave splitting in local S, SKS and SKKS (SK(K)S) phases. Comparisons of local S splitting times and path lengths in the slab, mantle wedge, and upper plate indicate that splitting times for arc and back-arc stations are consistent with anisotropy in the mantle wedge, but long slab paths to fore-arc stations imply that slab anisotropy is also significant. SK(K)S shear-wave splitting observations and models for sub-slab anisotropy show that significant anisotropy is present below the slab, and that the orientation of sub-slab anisotropy sometimes differs from anisotropy above the slab. Anisotropy both above the slab and below the slab in the South American subduction zone is consistent with mantle flow that is driven by a combination of entrainment with down-going slab motion and flow complexity related to variations in slab shape and slab rollback.

1. Introduction

Seismic anisotropy in mantle regions of subduction zones is an important tool for probing mantle flow and subduction dynamics. Studies over the years have revealed that anisotropy either below the slab or in the mantle wedge seldom matches the predictions of simple two-dimensional corner flow driven by mantle entrained with the down-going slab (e.g. Russo and Silver, 1994; Yang et al., 1995; Fouch and Fischer, 1996; Klosko et al., 1999; Hall et al., 2000; Smith et al., 2001; Pozgay et al., 2007; Hoernle et al., 2008; Long and Silver, 2008, 2009). In many subduction zones, anisotropy inferred to exist below the slab from shear-wave splitting in teleseismic phases such as SKS and SKKS aligns parallel or sub-parallel to the strike of the subducting slab (Russo and Silver, 1994; Fouch and Fischer, 1996; Peyton et al., 2001; Audoin et al., 2004; Anderson et al., 2004; Civello and Margheriti, 2004; Long and van der Hilst, 2005, 2006; Baccheschi et al., 2007; Long and Silver, 2008, 2009; Abt et al., 2010; Kiraly et al., 2012). In a few cases, such as Cascadia (Currie et al., 2004; Eakin et al., 2010), Mexico (Pinero-Feliciangeli and Kendall, 2008; Leon Soto et al., 2009), central Alaska (Christensen and Abers, 2010) and Sumatra (Hammond et al., 2010), the dominant orientation inferred for sub-slab anisotropy is in the down-dip direction of the slab. Fast polarization directions that are parallel and normal to slab-strike are hereafter referred to as “slab-parallel” and “slab-normal”. Slab-parallel anisotropy has been interpreted to reflect along-strike deflection of the sub-slab mantle in response to retrograde motion of the slab combined with decoupling of the slab and the underlying mantle, while slab-normal anisotropy has been taken as an indication of greater slab-mantle coupling (e.g. Russo and Silver, 1994; Long

and Silver, 2008, 2009). Subduction zones where the direction of sub-slab anisotropy varies significantly, sometimes over length-scales of a few hundred kilometers, provide important cases against which to test models for subduction dynamics and coupling between the slab and surrounding mantle. Japan is one such subduction zone (Long and van der Hilst, 2005; Long and Silver, 2009) and South America is another (Bock et al., 1998; Polet et al., 2000; Anderson et al., 2004; Russo et al., 2010a, 2010b; Kneller and van Keken, 2007b).

The goal of this study is to more accurately isolate anisotropy above, within, and beneath the Nazca slab in the southern South American subduction zone through modeling of shear wave splitting measurements from local S phases, which sample anisotropy in the slab, mantle wedge and upper plate, and from SK(K)S phases, which sample anisotropy from the core-mantle boundary to the station, but are typically interpreted in terms of upper mantle anisotropy. This approach, combined with data from new stations in this area that have not been included in previous shear wave splitting studies, reveals sub-slab fast symmetry axes whose orientation relative to slab strike varies from parallel to normal, as well as coherent mantle wedge anisotropy that suggests multiple drivers of mantle flow.

2. Tectonic Setting

In the South American subduction zone, the Nazca Plate has been subducting along the western edge of the South American Plate since ~69 Ma (Pardo-Casas and Molnar, 1987). While the exact rate and direction of convergence have varied over time

(Pardo-Casas and Molnar, 1987; Yañez et al., 2002), the current rate between the Nazca and South American Plates is 7-8 cm/year in our study region (28°S-42°S; Syracuse and Abers, 2006). Depending on reference frame, the trench is migrating with retrograde motion of up to 4.5 cm/year or is essentially stationary (Schellart et al., 2008). Features of particular tectonic interest include the area of flat slab subduction between 28°S and 33°S and the subduction of the Chile Ridge at 46°S.

From its southern edge at 46°S to 34°S, the Nazca slab dips to the east at an angle of 30°-35° (Syracuse and Abers, 2006). While the exact shape and extent of the flat slab is still debated (Cahill and Isacks, 1992; Syracuse and Abers, 2006; Anderson et al., 2007; Linkimer, 2011), most models agree that between 28°S and 30°S, the slab dip rapidly decreases to values of 0°-15° by 31°S (Figure 1). Based on eastward migration of arc magmatism, shallowing of slab dip began at 18-20 Ma; magmatism in the flat slab region ceased at 5-1.9 Ma (Kay and Abruzzi, 1996; Ramos et al., 2002). Although the causes of slab flattening are still debated, one hypothesis is that flattening is initiated by subduction of anomalously thick oceanic crust. In the case of the flat slab zone in our study region, subduction of the Juan Fernandez Ridge, a chain of seamounts generated by plume material from the Alexander Selkirk Island Hotspot (Flueh, 1998), is spatially and temporally correlated with the onset of slab flattening (Gutscher et al., 2000; Yañez et al., 2002).

The Chile Ridge forms the plate boundary between the Nazca Plate and the Antarctic Plate (Figure 1), and subduction of the Chile Ridge began at 14-15 Ma (Lagabriele, 2004). A gap (Murdie and Russo, 1999) may be found between the Nazca slab and the Antarctic slab because of the differing convergence rates of ~7 cm/year and

~2 cm/year, respectively; this gap is consistent with velocity models from body-wave tomography (Russo et al., 2010b).

3. Prior Shear-wave Splitting Studies in the Region

The Nazca-South American subduction zone is the site of the seminal paper that argued for the existence of sub-slab anisotropy with slab-parallel olivine a-axes and the generation of this anisotropy by slab-parallel flow induced by retrograde slab motion (Russo and Silver, 1994). However, this study also noted the presence of localized zones of slab-normal fast polarizations, and suggested that they could represent deflection of sub-slab flow due to pressure gradients created by variations in slab dip. Subsequent studies of shear-wave splitting at 12°-28° S (north of our study region) obtained a mix of slab-normal and slab-parallel fast polarizations in teleseismic shear-waves (Figure 2), and again invoked retrograde slab motion combined with slab morphology as the origin of this signal (Bock et al., 1998; Polet et al., 2000). Polet et al. (2000) also analyzed splitting in S phases from local earthquakes and found fast polarizations that were aligned predominantly northward and splitting times of less than 0.3 s. Further to the east at these latitudes, fast polarizations across the back-arc are largely ENE-ESE, with a more SE deflection apparent at the southern edge of the edge of the thicker lithosphere of the Sao Francisco craton (e.g. 21°S, 47°W in Figure 1) (James and Assumpção, 1996; Heintz et al., 2003; Assumpção et al., 2006 and 2011).

Between 30°S and 36°S, Anderson et al. (2004) measured splitting in SK(K)S phases recorded by the Chile Argentina Geophysical Experiment (CHARGE) array, and

found fast polarizations that were predominantly aligned to the north, with the exception of fast polarizations rotated clockwise to the east at stations above the eastern portion of the flat slab (Figure 2). Anderson et al. (2004) interpreted this pattern in terms of sub-slab flow entrained to the east with the subducting slab in the flat slab zone, and flow deflected to the south by retrograde slab motion flow south of 33°S. Anderson et al. (2005) analyzed splitting in local S phases at CHARGE stations and found fast polarization trends similar to the SKS results, but significantly smaller splitting times (0.1-0.4 s), which they attributed to a remnant of mantle wedge (now too cold to form new anisotropy) preserving slab-parallel flow in the wedge during the time of slab flattening, on the basis of mantle properties determined via interpretation of local body wave tomography (Wagner et al., 2006). Kneller and van Keken (2007b, 2008), however, used numerical models to show that the simple change in slab dip across the southern boundary of the flat slab zone could create significant extension in the mantle wedge parallel to slab strike. Kneller and van Keken (2007b) applied slab contour data from Anderson et al. (2007) to constrain mathematical models of the unique slab geometry of the Nazca Plate, and used this model to show that the wedge extension induced by this geometry produces a strong arc-parallel fast polarization direction trend that matches the local S data of Anderson et al. (2005), supporting the hypothesis that slab-strike-parallel fast directions could arise from current deformation within the active mantle wedge.

In the fore-arc to the south of the CHARGE array, Hicks et al. (2012) found teleseismic splitting with NNE-ENE fast polarizations (Figure 2), and local S fast polarizations with roughly ESE-oriented fast polarizations and splitting times of 0.1-0.4 s.

In the back-arc to the south of the CHARGE array, Helffrich et al. (2002) found ENE-oriented SKS fast polarization directions at station PLCA, and, further to the east, Assumpção et al. (2011) obtained SE-oriented SKS fast polarizations at station TRQA (Figure 2). In the vicinity of the Chile Ridge slab gap to the south of our study region, variable fast polarizations in teleseismic shear phases (Murdie and Russo, 1999; Russo et al., 2010a, 2010b) have been interpreted as evidence of mantle flow through the slab gap (Russo et al., 2010a, 2010b).

4. Shear-wave Splitting Measurements

4.1 *Data and methods*

We measured shear-wave splitting in shear-waves from local events (to isolate anisotropy within and above the Nazca slab) and in teleseismic SK(K)S phases (to sample anisotropy from the core-mantle boundary to the station). We made local S measurements at two temporary broadband networks, the 22 station CHARGE array (2000-2002) (Anderson et al., 2004; Wagner et al., 2005) and the 65 station Chile RAMP Experiment (2010) [Incorporated Research Institutions for Seismology (IRIS) and the University of Chile]. We also acquired local S splitting at three permanent broadband stations TRQA (2000-2010; IRIS/USGS Global Seismograph Network), PEL (1995-2008; GEOSCOPE) and PLCA (1998-2009; USAF/USGS Global Telemetered Seismograph Network). We collected SK(K)S splitting measurements at PEL, PLCA, TRQA and selected CHARGE stations (JUAN, LENA, BARD and NIEB). Average SK(K)S splitting values at all CHARGE stations are adopted from Anderson et al.

(2004). Stations at which splitting measurements were obtained are shown in Figure 1. We obtained all of our data from the IRIS Data Management System.

For the local S measurements at CHARGE array stations, and for PEL during 2000-2002 when the CHARGE array was in the field, we used event locations from Anderson et al. (2007). For local S phases at Chile RAMP stations, we used event locations from Lange et al. (2012). For all SK(K)S phases, for local S at permanent stations PLCA and TRQA, and for local S at PEL except for the period overlapping the CHARGE array, we use the USGS National Earthquake Information Center (NEIC) catalog.

For a local S event to be selected for splitting analysis, we required a ray path incidence angle at the station of less than 40° - 45° from vertical (to minimize particle motion perturbations associated with the free surface or crustal layering [e.g. Booth and Crampin (1985)]). For the CHARGE and permanent stations, events were limited to hypocentral depths of more than 40 km and incidence angles of less than 40° . For Chile RAMP stations, limits were 20 km and 45° , due to the position of these stations in the fore-arc where seismicity is shallower. SK(K)S phases were limited to epicentral distances of 90° - 120° . All waveforms were visually inspected and only those with clear local S or SK(K)S phases were retained in the splitting results.

Splitting measurements were made on the horizontal components of the shear-wave using the eigenvalue minimization method of Silver and Chan (1991). In practice, a grid search was performed over all possible fast polarization directions (ϕ) in increments of 1° and splitting times (dt) of 0 to 5 s in increments of 0.01 s to find the value that yielded the most singular horizontal component covariance matrix. Examples of splitting

in a local S phase (Figure 3a) and an SKKS phase (Figure 3b) illustrate that, as expected, removing the effects of the best-fitting splitting parameters from the waveforms shifts the apparent slow component of the split shear wave so that it aligns in time with the fast component. This operation alters the initially elliptical particle motion so that it becomes more linear.

Shear-wave splitting measurements were typically made using waveforms filtered with a 0.05 Hz – 2 Hz bandpass. An analysis of frequency dependence in the observed splitting was also carried out for select stations, and will be described later. The time windows over which the splitting analysis was applied vary from 2 to 5 s for local S phases and 10 to 30 s for SK(K)S phases. Multiple measurements were made on each waveform with different window lengths, and a splitting result was retained only if parameters did not significantly vary between windows.

Uncertainties in the splitting parameters were calculated using the approach described in Abt and Fischer (2008). At each point in the grid of trial ϕ and dt values, the minimum eigenvalue of the horizontal component covariance matrix is divided by the minimum eigenvalue for the best-fitting splitting parameters. The 95% confidence region for the best-fitting parameters is then defined by eigenvalue ratios that are less than $(1 + F(f, d-f, 95\%)(f/(d-f)))^{1/2}$ where f is the number of free parameters (two for ϕ and dt), d is the number of pieces of independent information, and $F(f, d-f, 95\%)$ is the 95% F-distribution value assuming a given d and f . To determine d for a given waveform, the length of the waveform time window is divided by the minimum resolvable time in the waveform. The latter value is defined by the first zero-crossing of the autocorrelation function for the 60 s prior to the local S arrival on the SV component. The uncertainties

for the best-fitting ϕ and dt values are taken to be the absolute value of the maximum difference between the best fitting values and any ϕ or dt within the 95% confidence region. Events that yielded fast polarization direction uncertainties of greater than 35° or splitting time uncertainties of more than 65% of the splitting time were discarded.

4.2 *Local S shear-wave splitting observations*

Well-constrained shear-wave splitting measurements were obtained for 291 shear phases from local events. Across our study region, fast polarizations vary significantly from slab-normal to slab-parallel; splitting times that vary from 0.07 to 1.02 s, with an average of 0.26 s (Figures 4 and 5, Supplemental Table 1). Although the first impression of the subduction zone is that fast polarizations in some regions are very scattered, a closer examination reveals that the fast polarizations are dominated by coherent trends, even if some of these trends vary at relatively small spatial scales. Such small-scale variation is not uncommon for local S phases in mantle wedge regions (e.g. Levin et al., 2004; Abt et al., 2009; Hicks et al., 2012). Due to the incidence angles of the local S phases, even uniform anisotropy with a horizontal fast symmetry axis can produce significant variations in fast polarization with back-azimuth. Alternatively, uniform anisotropy with plunging fast symmetry axes and/or spatial variations in anisotropy may also be responsible for some of the apparent small-scale complexity. However, in some cases, closely adjacent paths yield distinct fast polarizations whose errors do not overlap, and in a few of these cases, the shear phases have similar particle motions after the best-fitting splitting parameters are removed. These latter results are difficult to explain with anisotropic structure, and more likely reflect small-scale scattering or other structural

perturbations to particle motion (Booth and Crampin, 1985), or noise that effectively masquerades as splitting and is not reflected in the splitting parameter uncertainties.

At stations in the subduction zone fore-arc (the region to the west of the bold line in Figure 5), fast polarizations at some latitudes (32° - 34° and 36° S- 37.5° S) are dominated by a N-NE orientation, whereas at others (34° S- 36° S and roughly 38° S), fast polarizations are highly variable and include values that are roughly normal to slab strike (Figure 7). However, even in these zones of complexity, paths with distinct fast polarizations can often be separated into spatially coherent groups, a result that suggests they represent real anisotropic structure.

At stations within the arc and back-arc, fast polarizations also span orientations from slab-parallel to slab-normal (Figures 4, 5, and 7). Fast polarizations at the southernmost station, PLCA, lie between NE and ENE, roughly in between slab-normal and slab-parallel. At arc and back-arc stations at latitudes of 35° - 37° S (LENA, BARD and NIEB), fast directions are aligned roughly N-NNE, closer to a slab-parallel trend. This orientation also dominates fast polarizations at fore-arc stations just west of the arc at these latitudes (AMER and the three CHILE RAMP stations that lie closest to the arc; Figure 5b). Moving further north, most fast polarizations at station USPA are N-NE, bending roughly parallel to the 100 km slab depth contour (Syracuse and Abers, 2006) as it curves to the east into the region of flat slab subduction. Anderson et al. (2005) obtained similar fast polarization trends for CHARGE stations LENA, BARD, NIEB and USPA.

At stations within the latitudes of the flat slab, fast polarizations show considerable variation, but with coherence at relatively short length scales. At stations in

the back-arc of the flat slab zone (JUAN, HEDI, PACH and RINC), fast polarizations range from nearly N to ESE (Figures 4 and 5a). A consistent variation with back-azimuth is most evident at JUAN where fast polarizations are N-ENE for the easternmost events, and ENE-ESE for $\sim$ S back-azimuths roughly parallel to slab-strike, and in general NNE-E for all other back-azimuths (Figures 4 and 6). As will be described more fully later in the modeling section, this back-azimuthal pattern is largely explained by a simple uniform model of anisotropy with a horizontal a-axis, although a model with a small a-axis plunge (15°) provides a slightly better fit.

4.3 *Local S splitting parameters as a function of path length*

Local S paths sample the subducting Nazca plate, the mantle wedge and the upper plate; anisotropy in any of these zones may contribute to observed splitting. To evaluate contributions to splitting from the slab, the wedge and the upper plate, trends in local S splitting times as a function of path lengths in a given zone were examined. Rays were traced in the AK135 1D reference velocity model (Kennett et al., 1995), and we used the hypocenter-defined slab contours of Syracuse and Abers (2006) as the top of the slab; for simplicity, the upper plate lithosphere was assigned an average thickness of 65 km. This average upper plate thickness exceeds the maximum crustal thicknesses found by Gilbert et al. (2006) but does not extend into what appears to be the asthenospheric mantle wedge as defined by the body-wave tomography of Wagner et al. (2005). Because of the potential for 3D ray-bending, ideally the ray paths would be calculated with respect to 3D velocity structure. However, a 3D velocity model of sufficiently high resolution (e.g. Wagner et al., 2005) does not exist throughout our study region. Instead we evaluated 3D

ray-bending in a simplified model that contains a slab with a constant dip of 30° . Seismic velocities were 8% faster than those given in the AK135 model both within the slab and within the overriding plate, and 4% slower than those values given in the AK135 model for the mantle wedge. We found that, across a range of backazimuths (180° - 360°) and incidence angles (20° - 40°), the lengths of ray path segments in the slab, mantle wedge and upper plate differ by less than 10 km between 1D and 3D cases.

For stations in much of the study region (south of 31°S), local S splitting times averaged in 20 km bins of path length show a well-resolved increase of splitting time with mantle wedge path length (Figures 8c and 8d). However, a consistent positive correlation between splitting time and mantle wedge path length is not observed for the most northern stations in the flat slab region (Figure 8b), nor is one seen when all fore-arc stations are considered together (Figure 8a). However, the positive correlations observed at more southern latitudes (south of 31.5°S ; Figure 8c-d) becomes apparent only for the longest mantle wedge paths ($> 80\text{ km}$), and the absence of this correlation in the flat-slab region may reflect a paucity of long paths. Fast polarization directions show no consistent trends with path length in the mantle wedge, and we are not able to resolve independent layers of anisotropy with specific olivine a-axis orientations.

When local S path lengths in the slab are less than 75 km, as is the case for all but 6 of the 291 well-constrained splitting measurements, no well-resolved increase in splitting time with slab path length is observed (Figure 9). For three CHARGE fore-arc stations (AMER, CONS, and SJAV), paths from 6 events travel unusually long paths within the slab ($> 75\text{ km}$), and the splitting times for these paths are relatively large (Figures 9a and 9d). Only one of these 6 paths travels through the mantle wedge, and its

mantle wedge path length is only 6 km. The other five ray paths have zero path length in Figure 8d. Finally, splitting times show no correlation with path length in the upper plate (Figure 10).

These splitting time trends with path length suggest that significant anisotropy is present in the mantle wedge south of approximately 31°S and that its orientation is uniform enough beneath each station that shear-wave splitting accrues with path length. The lack of a clear positive correlation between splitting time and wedge path length for the most northern stations (HURT, ELBO, NEGR, PACH, HEDI and RINC) (Figure 8b) may simply reflect a smaller number of long wedge path lengths (particularly those that exceed ~ 70 km), but it could also mean that mantle wedge anisotropy is weaker and/or more complex within this zone of the flat slab region. The splitting time trends also indicate that neither slab anisotropy nor upper plate anisotropy dominates the observed splitting, except for the small number of phases at AMER, CONS, and SJAV that have unusually long paths in the slab but mostly do not sample the mantle wedge (Figures 8d and 9d).

The slab contours of Syracuse and Abers (2006) differ from those in Gans et al. (2011) who used receiver functions to independently determine the depth to the slab in the flat slab segment. Slab contours in this region are reflected in Figures 8a-c and 9a-c. However, over most of the Gans et al. (2011) study region, the contours of Syracuse and Abers (2006) underestimate the depths given by Gans et al. (2011) by about 10 km. Contour values differ the most in the northwestern corner of the flat slab where Gans et al. (2011) locate the slab 40 km deeper than in Syracuse and Abers (2006), but only four events with splitting measurements originate from this area (along the 70°W longitude

line from 30°S-31°S, recorded at stations NEGR, HURT, and ELBO). Thus, employing the Gans et al. (2011) contours would not significantly alter the trends in Figures 8 and 9.

To further constrain the relative contributions to the observed splitting from anisotropy in the mantle wedge, slab, and upper plate, phases with zero path lengths in the mantle wedge or in the slab were isolated. For the 39 phases with no path length in the slab, the best-fit line, as determined by a least squares solution (weighted by observed splitting time errors), has a slope of $2.2\text{e-}03$ s/km (which defines the linear relationship between the wedge path length and the splitting time), and a zero path length intercept of 0.1 s. For the 68 phases with no path length in the wedge, the weighted least squares solution for a linear relationship of splitting time with slab path length yields a value of $2.0\text{e-}03$ s/km, with a zero path length intercept of 0.2 s. These relationships are consistent in that the strength of anisotropy in the slab is similar to the strength of anisotropy in the mantle wedge. The zero path length intercepts indicate the amount of splitting present if the ray paths do not sample either the slab or the wedge, leaving only the overriding plate as a source of anisotropy. Therefore, upper plate anisotropy typically contributes 0.1-0.2 s of splitting to the overall splitting time, assuming that upper plate anisotropy is relatively uniform. The upper plate contribution to shear-wave splitting inferred here is somewhat larger than the ~ 0.07 s inferred from the upper plate in Nicaragua and Costa Rica by Abt et al. (2009).

4.4 *Frequency dependence in local S splitting*

To assess possible frequency dependence in local S splitting, splitting was re-measured using a variety of bandpass filters for 21 S phases that produced high quality

splitting measurements with the standard filter (0.05 – 2 Hz). For each phase, the low frequency corner of the bandpass was fixed at 0.05 Hz, and the high frequency corner was increased incrementally between 0.5 Hz and 2 Hz. For approximately 70% of the phases, fast polarizations (~72% of the measurements) and splitting times (~67% of the measurements) did not vary outside their 95% uncertainties across all filters. The remaining phases had well-constrained measurements with error bars that did not overlap, indicating a small degree of frequency dependence. Because of the relatively small dependence on frequency, splitting measurements are reported only for the standard filter (0.05 – 2 Hz) elsewhere in this paper.

More extreme variations in bandpass filters also suggest that frequency dependence is relatively mild. For example, Anderson et al. (2005) measured local S splitting at the CHARGE stations using higher frequency bandpass filters (0.4 Hz to 8, 10, and 15 Hz) and obtained a range of splitting times (0.05-0.55 seconds) that largely overlaps with the values for CHARGE stations obtained in this study with the 0.05 – 2 Hz filter (0.09-0.89 seconds), particularly when the 95% uncertainties are considered. This frequency dependence in splitting time is less than was observed in Japan by Wirth and Long (2010) with filters of 0.02 – 0.125 Hz versus 0.125 – 0.50 Hz. With their higher frequency filters, Anderson et al. (2005) obtained overall fast polarization trends for CHARGE stations AMER, LENA, BARD, NIEB, USPA, JUAN, HEDI and NEGR that are broadly similar to those in this study, again suggesting that splitting is relatively stable as a function of frequency. Larger discrepancies at other CHARGE stations likely reflect greater spatial variation in fast polarizations coupled with small numbers of events that do not overlap between the two studies.

In summary, local S splitting measurements show the simplest fast polarization patterns for stations in the arc and back-arc south of the flat slab region where they vary from an azimuth in between slab-normal and slab-parallel at PLCA to roughly slab-parallel at BARD, NIEB, LENA, and USPA. Splitting time correlations with slab, wedge, and upper plate path lengths are consistent with a mantle wedge anisotropy as the dominant source of the splitting. Stations in the fore-arc and within the flat slab region show greater complexity in fast polarization orientations. Relatively large splitting times for a few long paths in the slab suggest that slab anisotropy is comparable in magnitude to that in the wedge. Overall, the local S splitting results are relatively independent of local S frequency content.

4.5 *SK(K)S shear-wave splitting observations*

37 well-constrained shear-wave splits in SK(K)S phases were obtained at permanent stations PLCA, TRQA and PEL and a sub-set of the CHARGE stations (BARD, NIEB, LENA, and JUAN; Figures 11 and 12, Supplemental Table 1). SK(K)S splitting measurements at a given station are dominated by fast polarizations that are internally consistent (most 95% uncertainties overlap), in contrast to the local S splitting at some stations.

Fast polarizations at PLCA vary from NE-E with one exception, similar to the results of Helffrich et al. (2002). In contrast, fast polarizations at TRQA (a station that lies farther from the arc to the west) range from SE-E, in agreement with the station-averaged SK(K)S splitting for TRQA reported by Assumpção et al. (2011). While we focus on the more local variations of SK(K)S observations close to the trench, we note

that SK(K)S far in the back-arc are consistently E-W oriented (Figure 2). Fast polarizations at NIEB, BARD and LENA are oriented roughly N, and those at JUAN are largely NE, in agreement with the station-averaged SK(K)S splitting of Anderson et al. (2004). At PEL, we obtained fast polarizations that vary from NNE to ENE. This trend differs from the more northerly average fast polarization of Anderson et al. (2004), although Anderson's study determined a high variability for data at PEL, and the 95% uncertainties for some of the PEL SK(K)S fast polarizations in this study do overlap the average fast polarization of Anderson et al. (2004).

SK(K)S splitting times range from 0.35 s to 2.42 s, with an average of 1.32 s. At a given station, SK(K)S splitting times are typically much larger than local S splitting times, a result that broadly argues for the presence of anisotropy beneath the local event hypocenters. Later in this paper, the strength of anisotropy both above and below the local events is quantitatively determined.

4.6 Frequency dependence in SK(K)S splitting

In order to directly compare the effects of frequency dependence between local S and SK(K)S phases, the same range of filters used for the analysis of frequency dependence in local S phases was applied to the SK(K)S shear-wave splitting measurements at PLCA. These filters produced almost no variation in SK(K)S splitting parameters, and none that was resolved at the 95% confidence level. This result makes sense because the dominant periods of the SK(K)S arrivals are on the order of 5-10 s, and most of the SK(K)S energy lies below the lowest frequency of the upper corner (0.5 Hz) used in the analysis. Our standard bandpass filter (0.05 to 2 Hz) is reasonably similar to

the typical filter used by Anderson et al. (2004) (0.02 to 1 Hz), although in some cases, Anderson et al. (2004) employed 0.02 to 0.4 Hz or 0.125 Hz. In any case, the range of splitting times obtained by Anderson et al. (2004) (0.3 to 3.3 s with an average of 0.95 s) is roughly similar to ours.

Overall, SK(K)S splitting observations suggest anisotropy whose seismic fast axis is parallel to slab-strike beneath some stations (NIEB, BARD, LENA, JUAN, and perhaps PEL) but neither parallel nor perpendicular to slab-strike beneath others (PLCA and TRQA). However, to make a more definitive interpretation of the SK(K)S splitting, contributions from anisotropy above and below the slab need to be constrained.

5. Modeling Anisotropy Above and Below the Slab

A particular goal of this study is to better isolate mantle anisotropy above the Nazca slab using local S phases, and, where possible, to remove the effects this shallower anisotropy from SK(K)S splitting in order to isolate anisotropy below the slab. To accomplish this goal, we first modeled splitting in local S phases (method described below) for individual stations and groups of stations to find the best-fitting uniform orientations and strengths of anisotropy for the mantle above the events. The effects of these uniform local S models were then removed from SK(K)S phases, and the residual SK(K)S splitting was modeled to determine the best-fitting uniform orientations and strengths of sub-slab anisotropy.

One limitation of this approach is that we employ uniform models to parameterize the anisotropy. Some stations, for example PEL and JUAN, manifest spatial variations in

local S fast polarization directions that are averaged out in their best-fitting uniform anisotropy model. However, the density of local S ray paths is not sufficiently high to permit spatially variable models to be well-constrained, or to attempt shear-wave splitting tomography (e.g. Abt and Fischer, 2008; Abt et al., 2009, 2010). Uniform anisotropy models likely provide a better approximation of real structure beneath stations where observed local S splitting is more uniform (PLCA, LENA, BARD and NIEB).

Another issue is the difference in frequency content between the local S and SKS phases. Even though the same bandpass filter was applied to local S and SK(K)S waveforms (0.05 to 2 Hz), the S(K)KS phases still have intrinsically larger dominant periods (typically 5-10 s) compared to the 1 to 2 s dominant periods in the local S phases. Because we are fitting the observations with uniform models of anisotropy, there is no inherent frequency-dependence in the forward calculation of predicted splitting parameters, as there might be with vertically- or laterally-varying anisotropic models (e.g. Rumpker and Silver 1999; Rumpker and Ryberg, 2000; Saltzer et al., 2000; Chevrot et al., 2004; Chevrot, 2006; Fischer et al., 2005, Abt and Fischer, 2008). Therefore, we employ a simple ray-based method of calculating predicted splitting based on the Christoffel equation. If real structure is not in fact uniform, the discrepant dominant periods of the local S and SK(K)S phases may cause them to sample the variations in anisotropy differently. However, we have no way to estimate these effects, other than to point out that they are less likely beneath stations with fairly uniform splitting parameters, and to reiterate that neither the local S splitting measurements nor the SK(K)S splitting observations manifest a strong frequency dependence.

5.1 *Local S models*

Local S splitting measurements were fit by models of peridotite anisotropy characterized by two or three parameters: the azimuth of the olivine a-axis (θ), the strength of the anisotropy relative to single crystal values (α), and, for station JUAN, a-axis plunge. Following Abt and Fischer (2008), and Abt et al. (2009, 2010), the relationships of the elastic constants were based on a blend of 70% olivine and 30% orthopyroxene. Forsterite elastic constants (Anderson and Isaak, 1995; Abramson et al., 1997) were used to represent olivine, and bronzite elastic constants (Frisillo and Barsch, 1972) were used to represent orthopyroxene. Based on Mainprice and Silver (1993), we assumed that the a-axis of olivine is aligned parallel to the c-axis of orthopyroxene, and that the b-axis of olivine is parallel to the a-axis of orthopyroxene. The plunge of the olivine a-axis is resolved independently from θ and α only when splitting observations span a wide range of incidence angles and back-azimuths (Abt and Fischer, 2008; Abt et al., 2010). Because the local S measurements at most stations do not provide sufficient sampling of incidence angle or back-azimuth, we fixed the olivine a-axis at horizontal (with the exception of station JUAN in Figure 6, which will be discussed further below). The strength of the anisotropy (α) is a scalar value that varies from 0% for an isotropic model to 100% for a model in which all of the olivine a-axes are perfectly aligned.

The particle motion perturbation method of Fischer et al. (2000) was used to predict splitting for the ray-path of a given shear phase. Rays were traced from the hypocenter to the station assuming the AK135 1D reference velocity model (Kennett et al., 1995) and the direction cosines for the ray were determined in 10 km depth increments. The phase velocities and particle motions for the fast and slow S-waves in

each 10 km layer were obtained from the eigenvalues and eigenvectors of the Christoffel equation for the layer. A 2 s wavelet with an initially linear particle motion at the bottom of the model was rotated to the fast and slow S-wave polarizations for the deepest layer and time-shifted according to their relative velocities; this operation was repeated for each layer until a depth of 50 km was reached. For simplicity, we assumed that the crust is isotropic, and set the base of the crust at an average depth of 50 km based on the receiver function modeling of Gilbert et al. (2006). This assumption will not significantly bias our model because we find that the upper plate is only slightly anisotropic, and contributes less than 0.1-0.2 seconds of splitting. Splitting was calculated from the synthetic particle motions using the same method that was applied to the observed waveforms (the eigenvalue minimization method of Silver and Chan (1991)).

We predicted splitting for each local S phase at a given station for models representing a grid of θ and α values; θ varied over 180° in increments of 10 degrees, and α varied from 10 to 100% of the strength of single crystal anisotropy in increments of 10%. (For station JUAN and in the SK(K)S modeling described later in the paper, where non-zero a-axis plunges were tested, θ values were tested over 360° .) The best-fitting model was defined as the combination of θ and α that minimized the misfit between predicted and observed fast polarizations and splitting times. ‘Misfit’ was defined as the mean normalized error-weighted misfit given by equations 3 and 4 of Abt et al. (2010), with the assumption that uncertainties in synthetic splitting measurements are equal to those for the corresponding observations. The 95% confidence limits of the best-fitting model were specified using an F-test for a two-parameter model (θ and α) in which the number of observations equals the number of modeled splitting parameters.

At JUAN, where ray path distribution is sufficient to delineate a clear dependence of fast polarization with backazimuth (Figure 6), a-axis plunges of 0° - 45° were tested. For each candidate plunge, the best-fitting values of θ and α were determined. Although a plunge of 15° produced the minimum weighted average fast polarization misfit (for $\theta = 70^{\circ}$ and $\alpha = 10\%$) the misfit for a horizontal a-axis (for $\theta = 60^{\circ}$ and $\alpha = 10\%$) was only slightly higher, while larger a-axis plunges produced significantly worse fits. Although the distribution of paths at other stations is insufficient to rule out larger a-axis plunges, the 0° - 15° a-axis plunges supported by the modeling at JUAN indicate that assuming the a-axis to be horizontal at other stations is not unreasonable.

The local S phases at each station (Table 1) were modeled, but at some stations splitting measurements were either too few in number or too variable to be fit with a meaningful uniform model. In these cases, splitting parameters for neighboring stations were combined. These grouped stations were BARD, LENA and NIEB, HURT and ELBO, and HEDI and PACH. In the end, models with interpretable 95% confidence limits were obtained for all but four stations (CONS, SJAV, and HURT/ELBO, which are the stations in Figure 13 that are closest to the trench and are not accompanied by best-fit model values) (Figure 13). The successful models for fore-arc stations have a-axis azimuths (θ) that are deflected either in a clock-wise (AMER) or counter-clockwise (PEL and NEGR) direction from a slab-parallel orientation, and they indicate anisotropy whose strength is 10-20% of single crystal values. At back-arc stations, a-axis azimuths are in between slab-parallel and slab-normal at PLCA and roughly slab-parallel at BARD/NIEB/LENA, USPA, JUAN, HEDI/PACH and RINC. Acceptable ranges of

anisotropic strength for local S models varied from 10-15% to 10-40% of single crystal anisotropy.

5.2 *Sub-slab models*

To remove the effects of the local S model at a given station from the SK(K)S splitting observations for that station, the splitting in each 10 km layer of the local S model was calculated for an individual SK(K)S ray using the Christoffel equation. The observed SK(K)S waveform for that ray was then “un-split” (Figure 14) by rotating to the predicted fast and slow S-wave polarizations for the shallowest layer and time-shifting to remove that layer’s predicted splitting time (e.g. Abt et al., 2010). This operation was repeated for successively deeper layers until the slab surface (Syracuse and Abers, 2006) was reached. The resulting waveform was then analyzed for residual shear-wave splitting. The ensemble of residual SK(K)S splitting measurements for each station was then modeled to find the best-fitting uniform structure, using the same procedure described in the preceding section on local S model, with three exceptions. The first is that the initial SK(K)S synthetic wavelet had a period of 10 s. The second is that the residual SK(K)S splitting models were assumed to extend from 410 km to the point where the phase exits the slab. The third is that models with an olivine a-axis plunge equal to the local slab dip were tested, as were models with horizontal olivine a-axes.

The residual S(K)KS splitting reflects anisotropy in both the slab and the sub-slab mantle. However, for 21 of 27 SKS paths (we do not use SK(K)S phases from TRQA, as those rays are too far in the back-arc to sample the slab), slab segments are less than 95 km in length, assuming a slab thickness of 80 km. This slab thickness is based on the

depth of the 1300°C isotherm assuming the crustal age range of the Nazca lithosphere subducting beneath our study region (20-40 Ma; Capitanio et al., 2011) and a plate-cooling model with an asthenospheric temperature of 1350°C and an asymptotic plate thickness of 95 km (Schmerr, 2012). Using the slope of the linear relationship between local S splitting time and slab path length (2.0×10^{-3} s/km), slab paths of 95 km or less yield splitting times less than or equal to 0.19 s. Even for the longest slab path, 134 km, this approach yields a slab splitting time of only 0.26 s. Because residual SK(K)S splitting times range from 0.31 s to 2.88 s with a mean 1.32 s, and no station has residual splitting times small enough to be explained by slab anisotropy alone, we infer that most of the anisotropy represented by the residual SK(K)S splitting is located below the slab. We note, however, that residual SK(K)S splitting in some subduction zones may contain a much larger contribution from slab anisotropy, as was determined by Hammond et al. (2010) for the Java-Sumatra subduction zone.

Residual SK(K)S splitting was modeled at permanent stations PLCA and PEL, and several CHARGE stations with a sufficient number of observations (JUAN and the combined splitting measurements of BARD, NIEB and LENA). For all of these stations, the models with the plunging a-axes fit the residual SK(K)S splitting better than the horizontal a-axis models, although this difference is not significant at the 95% confidence level. The best-fitting a-axis azimuth (θ) and anisotropic strength (α) values for plunging and horizontal a-axis cases either agree exactly, or overlap at 95% confidence. The best-fitting models with plunging a-axes are shown in Figure 13, although both cases are given in Table 1. No local S splitting measurements exist at TRQA, and for this

station, total SK(K)S splitting was modeled with a uniform anisotropic layer extending from 410 km to 50 km.

In the cases with plunging a-axes for PLCA, PEL, and JUAN, the azimuth and plunge of the best-fitting sub-slab a-axis and its 95% uncertainties point roughly in the slab's down-dip direction (SE, NE, and ENE, respectively). This is not the case at the station group that consists of BARD, NIEB, and LENA. For these stations, both the best-fitting non-plunging and plunging a-axis azimuths ($\sim 30^\circ$) are oriented somewhat counter-clockwise of slab-parallel. The downward plunge of the plunging a-axis case therefore points slightly away from the down-dip direction, although it is largely aligned with the strike of the slab. At PLCA, the best-fitting sub-slab a-axis azimuth is rotated considerably clockwise of the mantle wedge a-axis from the local S modeling. This result makes sense considering that a number of the SK(K)S fast polarizations start out clockwise of the local S a-axis, and that removing the effects the local S model tends to produce residual SK(K)S fast directions for these phases that are rotated even further clockwise (Figures 13 and 14). However, scatter in the raw and residual SK(K)S splitting measurements results in large uncertainties on the sub-slab a-axis azimuth. At TRQA, where recorded waves may not sample the Nazca slab at all, the best-fitting horizontal a-axis azimuth has a slab-oblique orientation. Acceptable ranges of anisotropic strength for both the plunging and non-plunging cases in Figure 13 varied from 10-20% to 10-40%.

6. Implications for subduction zone mantle flow and deformation

6.1 *Fore-arc*

Local S shear-wave splitting for paths in the fore-arc mantle manifests relatively simple slab-parallel fast polarizations in some regions, and more complex fast polarizations that range from slab-parallel to slab-normal in others. These results suggest variations in the orientation of anisotropy at relatively small spatial scales. This heterogeneity could reflect variations in strain geometry, variations in olivine water content, or variations in stress levels that create zones with different olivine fabrics, for example, B-type fabrics at high water and stress conditions versus A-, C-, or E-type fabrics (Katayama et al., 2005; Karato et al., 2008).

6.2 *Arc and back-arc mantle wedge*

Within the high temperature mantle wedge beneath the arc and back-arc, olivine lattice-preferred orientations are more likely to develop with C- or E-type fabrics, assuming the likely presence of significant water in olivine, or with an A-type fabric if the mantle wedge is anomalously dry for a subduction zone (Katayama et al., 2005; Kneller et al., 2007a; Karato et al., 2008). In all of these cases, olivine a-axes will tend to align sub-parallel to the mantle flow in relatively simple flow-fields that persist over longer length scales (hundreds of kilometers) (Kaminski, 2002; Kaminski and Ribe, 2002). Because mantle wedge flow fields may violate these conditions, combined numerical modeling of flow, LPO development and shear-wave splitting is desirable. However, in the absence of such modeling, we will make a few observations assuming that the a-axis azimuths in the local S models roughly reflect the azimuth of mantle flow.

The best-fitting olivine a-axis azimuths at PLCA are sub-parallel to the direction of absolute Nazca plate motion in the Indo-Australian hotspot reference frame, and

rotated counter-clockwise between 40-50° from Nazca plate motion in the Pacific hotspot and “no net rotation” reference frames (Schellart et al, 2008). Assuming that this anisotropy resides in the mantle wedge, as would be consistent with splitting time versus path length correlations (Figure 8), it could represent wedge material largely entrained by the subducting slab, but with some deflection due to another flow driver. Mantle flow driven through the Chile Ridge slab gap to the south (e.g. Russo et al., 2010a, 2010b) is one possible source of rotation away from flow driven purely by subducting plate motion. Large changes in slab dip can also deflect flow towards slab strike (Hall et al., 2000; Kneller and van Keken, 2007b), but this effect tends to be relatively local to the dip change (Kneller and van Keken, 2007b, 2008). Large slab dip variations are not evident in the slab contours of Syracuse and Abers (2006) in the vicinity of PLCA, and this station likely lies too far south to be affected by the dip variation associated with the flat slab at ~33°S.

Modeled a-axes further to the north in the mantle wedge appear to lie sub-parallel to slab strike, with ~N azimuths at BARD, LENA and NIEB, and NNE-NE azimuths at USPA, JUAN, RINC, PACH and HEDI. Although wedge flow lines at these latitudes very likely reflect an underlying entrainment of shallow material from the back-arc by the subducting slab, the decrease in slab dip towards the flat slab could create a zone of low pressure in the wedge that deflects flow lines to the north (e.g. BARD, LENA and NIEB), before they become entrained sub-parallel to subducting plate motion (e.g. USPA, JUAN, RINC, PACH and HEDI) (Hall et al., 2000; Kneller and van Keken, 2007b, 2008).

Further in the back-arc, beyond the eastward limit of the Nazca plate in the upper mantle, the best-fitting a-axis azimuth obtained for TRQA for the whole upper mantle

would be consistent with simple shearing of the South American plate relative to the asthenosphere, assuming the Indo-Atlantic hotspot absolute plate motion reference frame, but not the Pacific hot spot or no-net-rotation reference frames (Schellart et al., 2008). While the TRQA a-axis azimuth is rotated 30° - 60° clockwise with respect to the majority of SKS fast polarizations in the South American back-arc at latitudes of 15°S - 30°S (Figure 2), the entire far east of the back-arc can be characterized as having an a-axis orientation that is more normal than parallel to the trench and to slab strike.

6.3 *Sub-slab mantle*

A key finding of this study is the varying geometry of sub-slab anisotropy (Figure 13, Table 1) relative to the strike of the slab and to the direction of Nazca plate motion. Sub-slab anisotropy in this subduction zone segment cannot be explained by simple down-dip entrainment of mantle with the subducting slab, but neither is it consistent with purely slab-parallel (e.g. Russo and Silver, 1994; Long and Silver, 2008, 2009) or toroidal flow (Kincaid and Griffiths, 2003; Funicello et al., 2003; Civello and Margheriti, 2004; MacDougall et al., 2011; Faccenda and Capitanio, 2012) driven by slab rollback or trench migration. Rather, the sub-slab anisotropy is consistent with flow that is driven by a combination of slab entrainment and interaction between retrograde slab motion and slab dip.

At $\sim 40^{\circ}\text{S}$ (PLCA) best-fitting sub-slab a-axes are aligned sub-parallel to absolute Nazca plate motion in the Pacific hotspot reference frame, but given the large uncertainties on PLCA a-axis orientation, a-axes may also be considered parallel to Nazca plate motion in the “no net rotation” and Indo-Australian reference frames

(Schellart et al., 2008). Assuming A-type olivine fabrics below the slab (or C- or E-fabrics if some water is present), the modeled a-axes are consistent with shear in a layer of mantle entrained by subducting plate motion. This interpretation is consistent with Hicks et al. (2012) who concluded that SKS splitting observed at stations in the fore-arc to the NW of PLCA (Figure 2) is best explained by Nazca plate entrainment of the sub-slab asthenosphere.

In contrast, at 35°-36°S (BARD, LENA, NIEB), sub-slab a-axes are oriented roughly parallel to slab strike but roughly normal to Nazca plate motion, while at 31°S-33°S (PEL, JUAN), sub-slab a-axes follow the curve of slab strike, but also align sub-parallel to Nazca plate motion and the slab's down-dip trajectory. Although the a-axes at 31°S-33°S are consistent with simple entrainment of the sub-slab mantle by Nazca plate subduction, the more N-trending axes at 35°-36°S suggest an interaction of slab shape and motion. For example, westward migration of the slab (present in both the Indo-Australian and Pacific hotspot reference frames) coupled with the decrease in slab dip at ~33°S would tend to drive flow parallel to slab-strike towards the flat slab region, matching sub-slab anisotropy at BARD, LENA and NIEB. This mechanism differs from the sub-slab flow model of Anderson et al. (2004) in the sense that Anderson et al. suggest southward flow south of the flat slab zone. Alternatively, flattening of the slab from 20 Ma to at least 1.9 Ma (Kay and Abruzzi, 1996; Ramos et al., 2002) would be another mechanism for drawing sub-slab flow from the south into the flat slab zone where it could become entrained with downgoing plate motion.

7. Conclusions

Modeling of shear-wave splitting in local S, SKS, and SKKS phases in the South American subduction zone from 30°S to 40°S has revealed that significant anisotropy resides both above and below the subducting Nazca slab.

At arc and back-arc stations, correlations of local S splitting times with phase path lengths in the slab, mantle wedge, and upper plate are consistent with significant anisotropy in the mantle wedge. In the southern part of the study region, splitting fast polarization directions and models of anisotropy that best fit the local S splitting at each station are neither parallel to the direction of subducting Nazca plate motion nor to the strike of the slab, suggesting complexity in wedge flow that may be related to the Chile Ridge slab gap. Further to the north, modeled olivine a-axes above the slab align roughly parallel to slab strike, bending to the east at stations within the flat slab zone. This pattern is consistent with flow driven by pressure gradients related to the northward decrease in slab dip coupled with slab entrainment. Far in the back-arc, modeled anisotropy may be explained by the motion of the South American plate relative to the underlying mantle. At fore-arc stations, local S splitting results vary more rapidly at short spatial scales. A few paths to fore-arc stations that have especially long segments in the slab are consistent with slab anisotropy that is roughly comparable in magnitude to that inferred in the mantle wedge.

When the effects of shallow anisotropy inferred from local S phases are removed from SK(K)S waveforms, significant residual splitting remains. Based on the strength of slab anisotropy inferred from local S phases and the length of SK(K)S phases in the slab, most of the residual SK(K)S splitting derives from sub-slab mantle anisotropy. In the

south of the study region, sub-slab anisotropy is consistent with a layer of mantle entrained beneath the subducting Nazca plate. Further north, sub-slab anisotropy has an azimuth that varies from roughly N and slab-parallel at $\sim 36^\circ$ S, but then aligns parallel to slab entrainment beneath the flat slab zone. Overall, sub-slab anisotropy is consistent with models that include significant mantle entrainment with subducting plate motion, but another flow driver must exist at mid-latitudes. Possible mechanisms include retrograde motion of the slab coupled with present-day along-strike variation in slab dip and/or flattening of the slab in the north over last 20 My.

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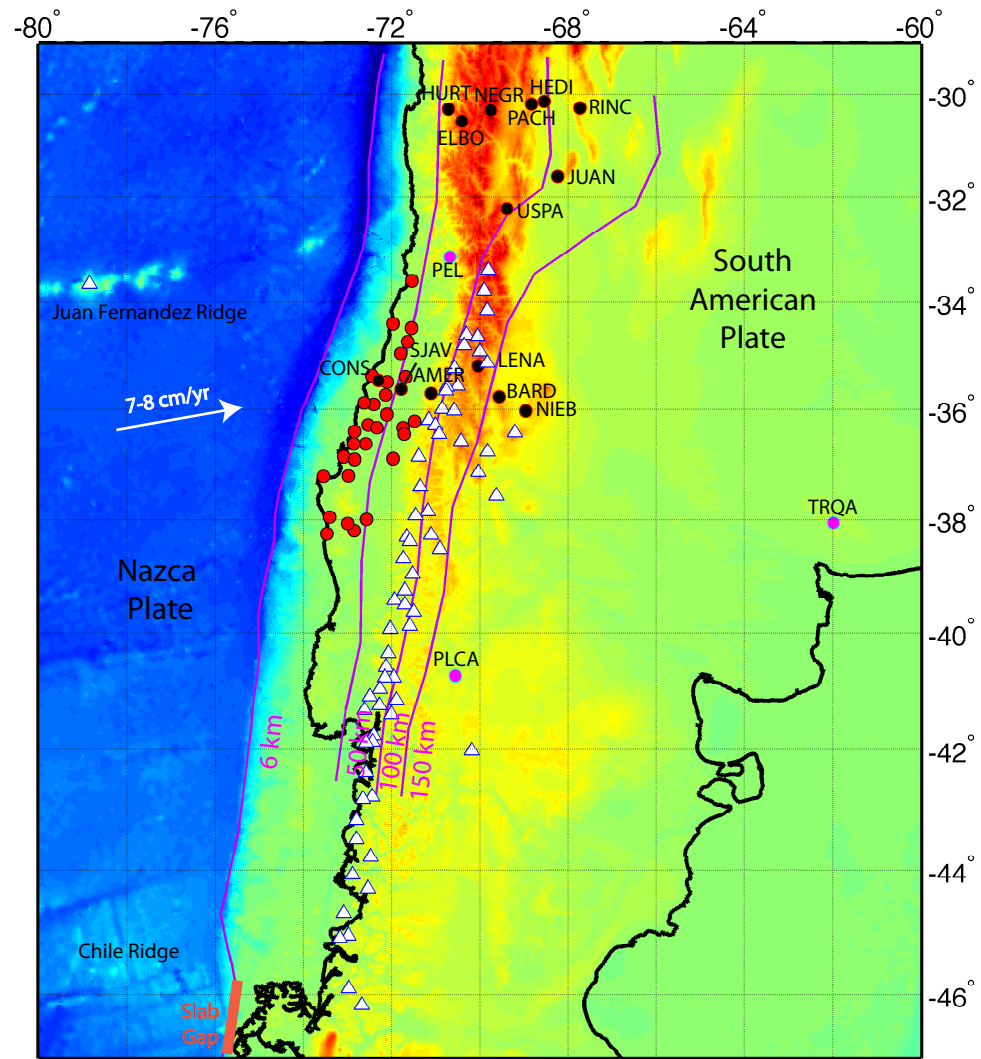


Figure 1. Map of topography with broadband stations where we obtained shear-wave splitting measurements. Black circles represent 14 of the 22 stations of the 2000-2002 CHARGE deployment; red circles denote 30 of the 65 stations of the 2010 Chile RAMP deployment (2010); pink circles are permanent stations PEL, TRQA, and PLCA. Slab contours are taken from Syracuse and Abers (2006). White triangles show volcanoes.

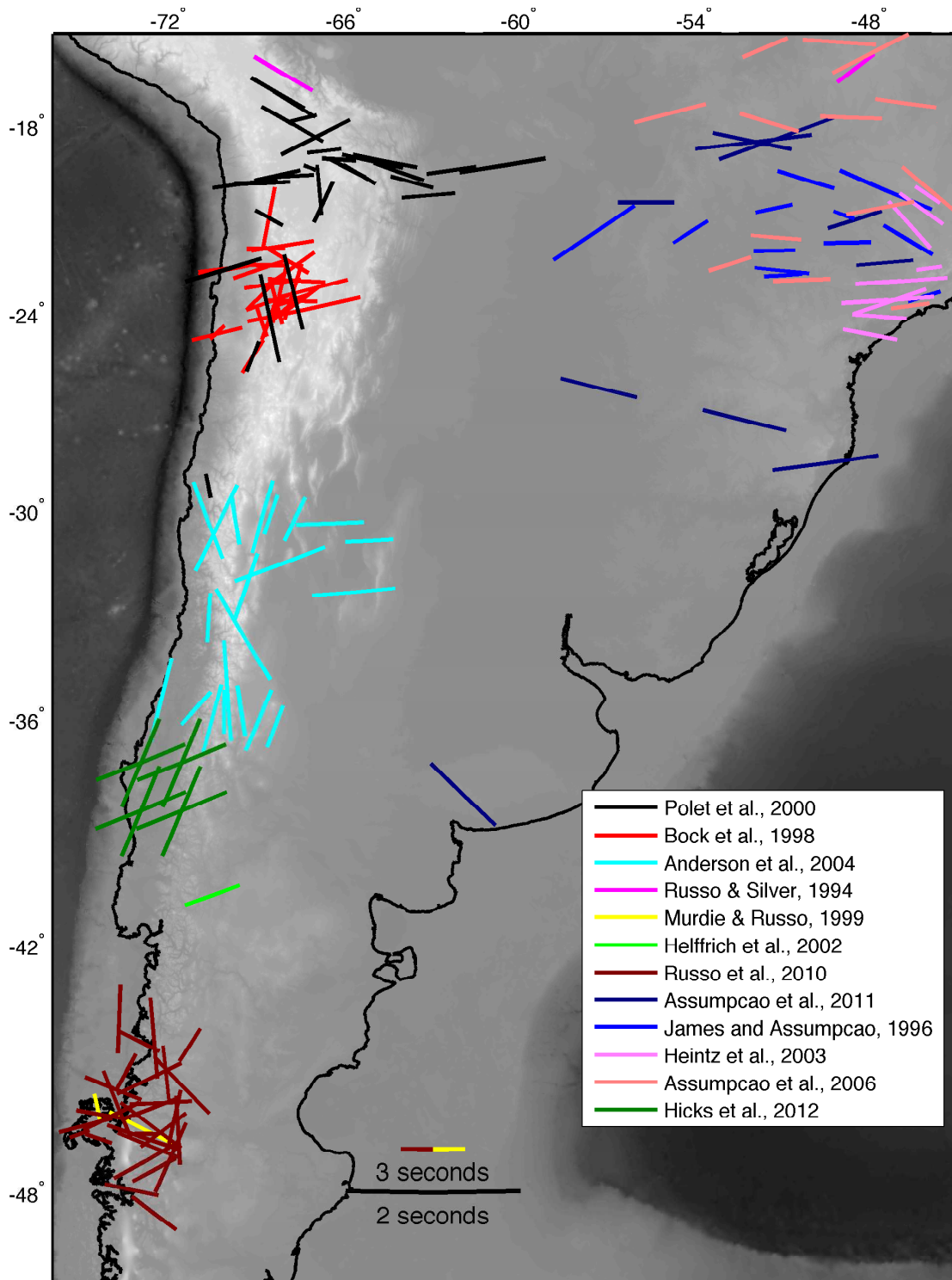


Figure 2. Previous *KS shear-wave splitting measurements in southern South America, where the orientation of the lines with respect to N is the fast polarization direction, and the scale bars showing the representative splitting time in seconds. (The maroon and yellow scale bar refers only to the Russo et al. (2010a and 2010b) and Murdie and Russo (1999) studies, as these measurements were scaled down for clarity.) Splitting measurements are represented by station averages, with the exception of Hicks et al. (2012) for which typical individual fast polarizations and average splitting time are shown schematically. Constraints from ScS are included in the Polet et al. (2000) measurements.

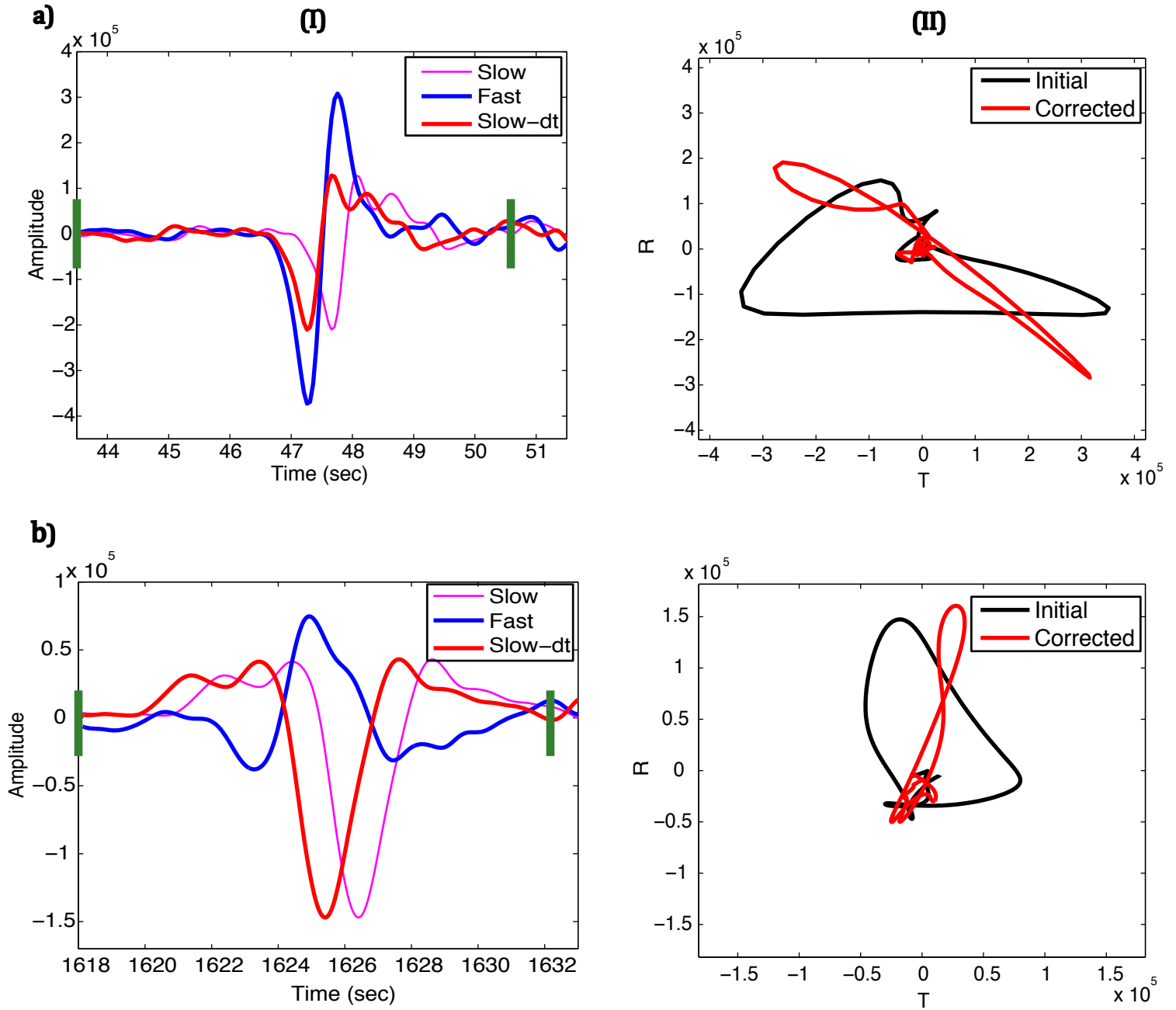


Figure 3. Examples of splitting measurements in a local S (a) and an SKKS (b) phase. Panels in Column I show fast (blue) and slow waveforms for the initial (pink) and corrected (red) particle motions. Green lines indicate the time window we used to calculate the splitting measurement. Panels in Column II show horizontal component particle motions for the initial waveform (black) and the waveform rotated and time-shifted to remove the effects of the best-fitting splitting parameters (red). The local S event (2009.061.07) was recorded at station PLCA and has an epicenter location of (40.3°S, 71.5°W, 179 km) and an origin time of 07:43:11; $\varphi = 44^\circ \pm 9^\circ$ and $dt = 0.38 \text{ s} \pm 0.06 \text{ s}$. The SKKS event (2004.207.14) was recorded at station PLCA and has a hypocenter location of (2.43°S, 103.98°E, 582 km) and occurred at 14:35:19; $\varphi = 66^\circ \pm 12^\circ$ and $dt = 1.24 \text{ s} \pm 0.26 \text{ s}$.

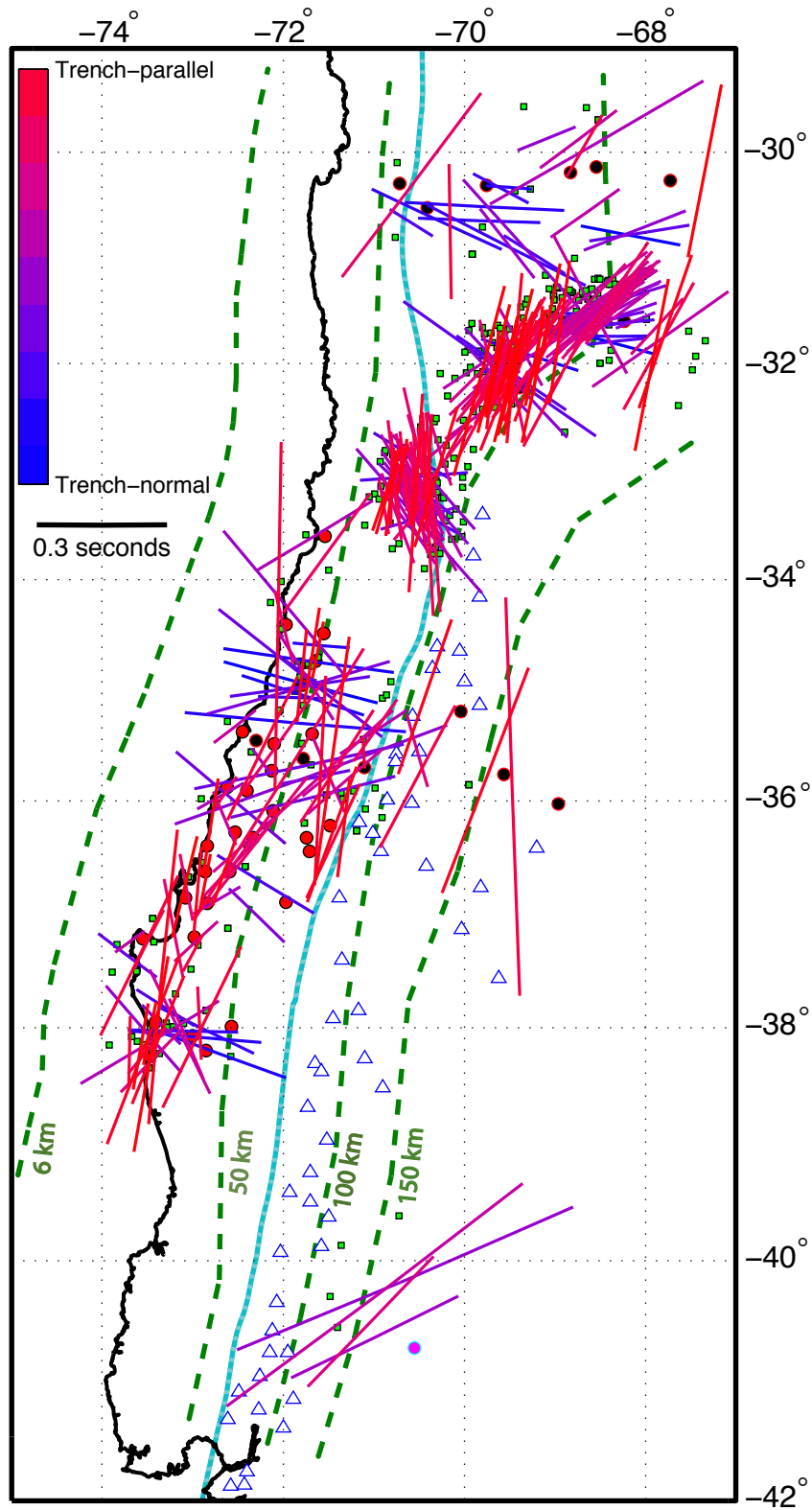
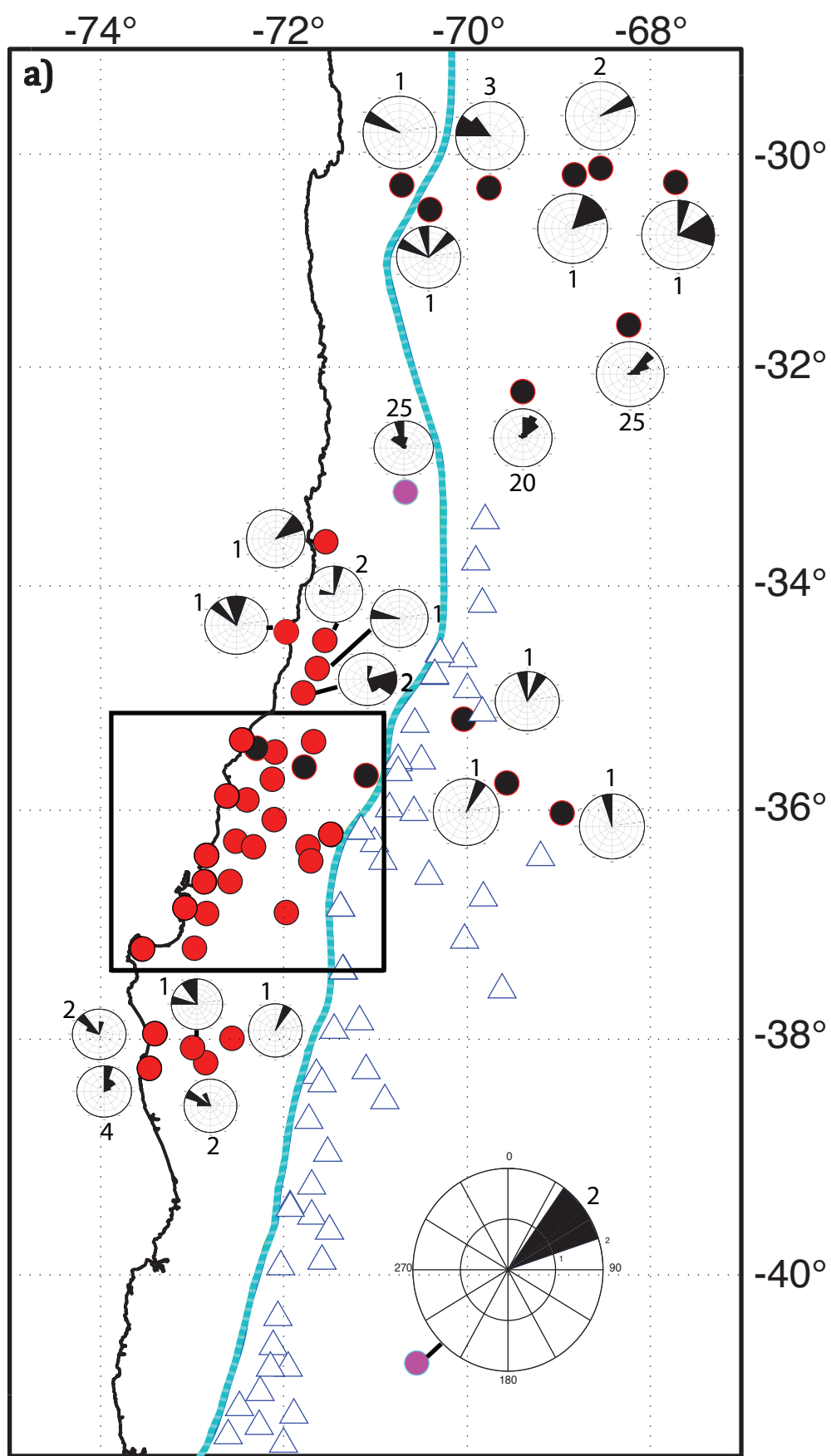


Figure 4. Local S shear-wave splitting results. Splitting vectors are aligned parallel to ϕ and their length scales to dt ; their colors range from trench-parallel fast direction (red) to trench-normal (blue), where the strike of the arc is defined as roughly $N13^{\circ}E$. Splitting vectors are plotted halfway between the event hypocenter (green squares) and the station (circles which are pink for permanent stations and black for temporary stations). White triangles show arc volcanoes. Slab contours from Syracuse and Abers (2006) are shown, as well as a light blue line, which we use to divide between the fore-arc/arc and the back-arc.



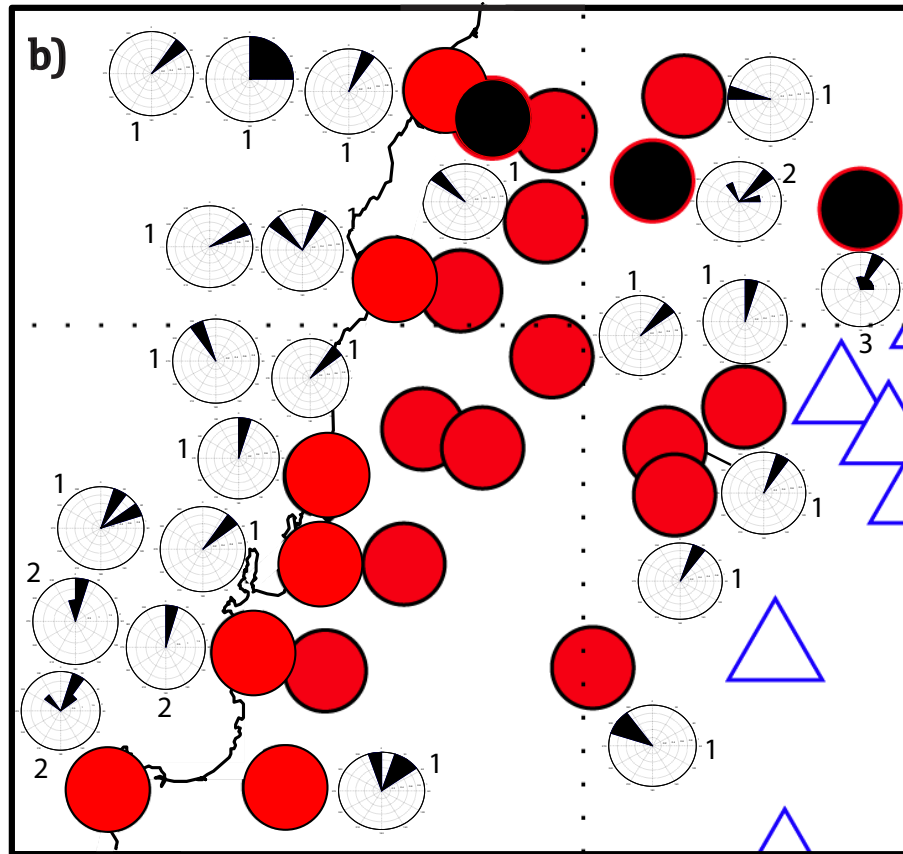


Figure 5. Rose diagrams of fast polarization directions at each station, with an example of a rose diagram enlarged for clarity at station PLCA. a) Entire study region. The bold blue line divides the region defined as the fore-arc (in the west) from the region defined as the arc and back-arc. b) Region shown by box in (a). Numbers outside each rose diagram represent the maximum radius (maximum number of occurrences in a given bin). Red circles are Chile RAMP stations, black circles are CHARGE stations, and pink circles are permanent stations PLCA and PEL.

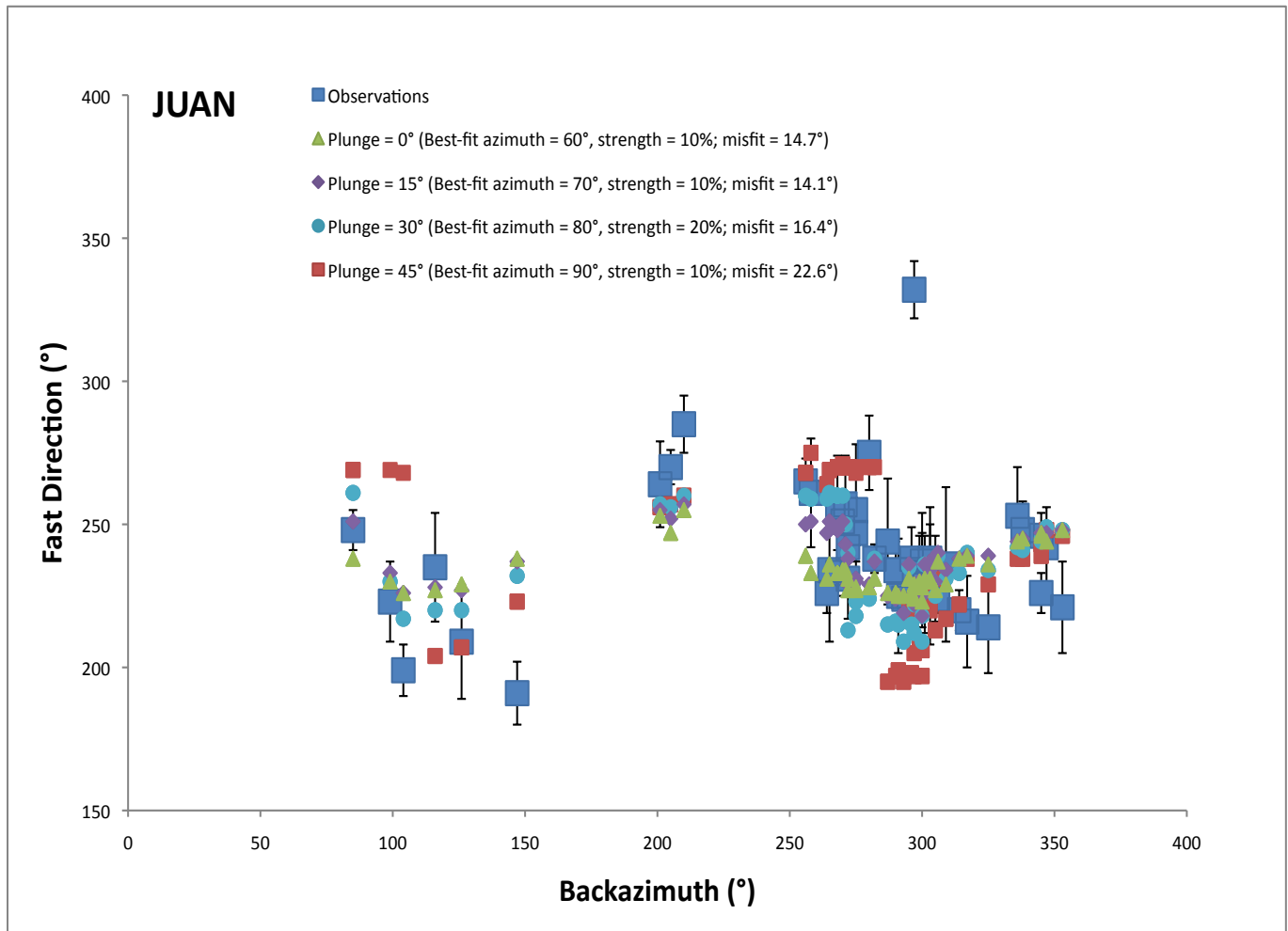


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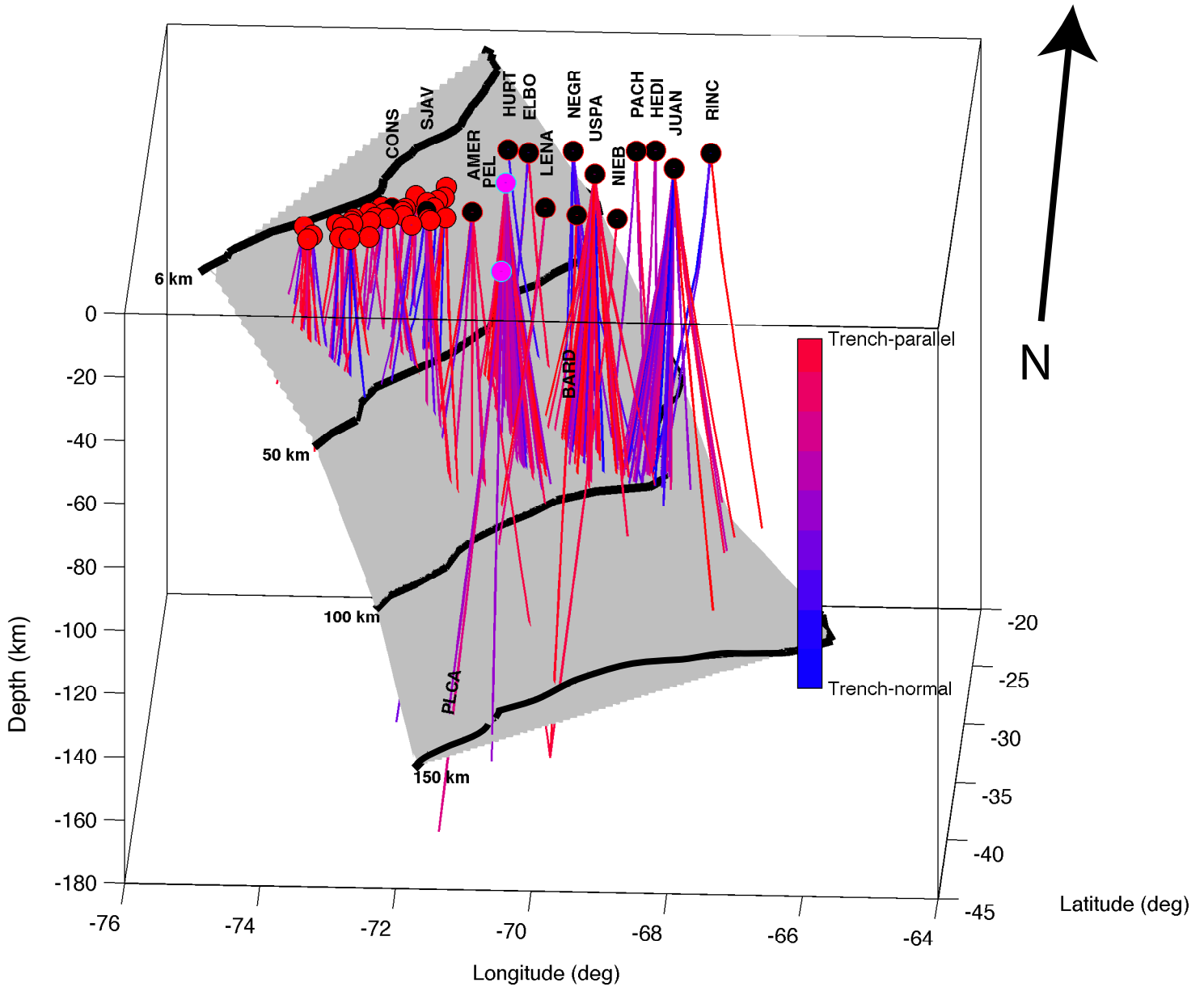


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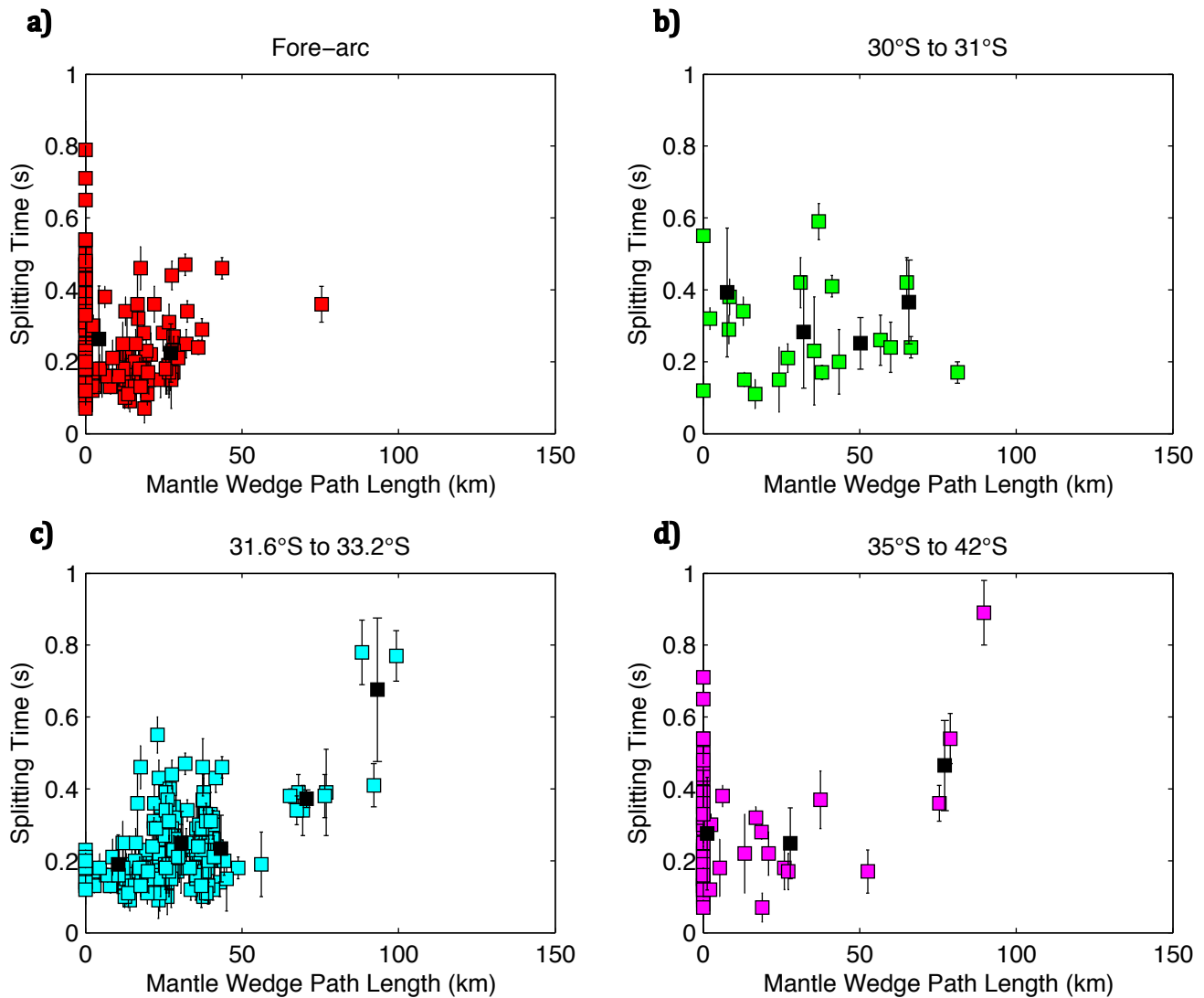


Figure 8. Plots of local S splitting times versus phase path length in the mantle wedge. Black squares represent error-weighted splitting time averages, binned in 20 km intervals of path length. Averages for bins with only one event are not displayed. Results are shown separately for stations in the fore-arc (a), and for fore-arc, arc and back-arc stations at latitudes of 30°S to 31°S (b), 31.6°S to 33.2°S (c), and 35°S to 42°S (d).

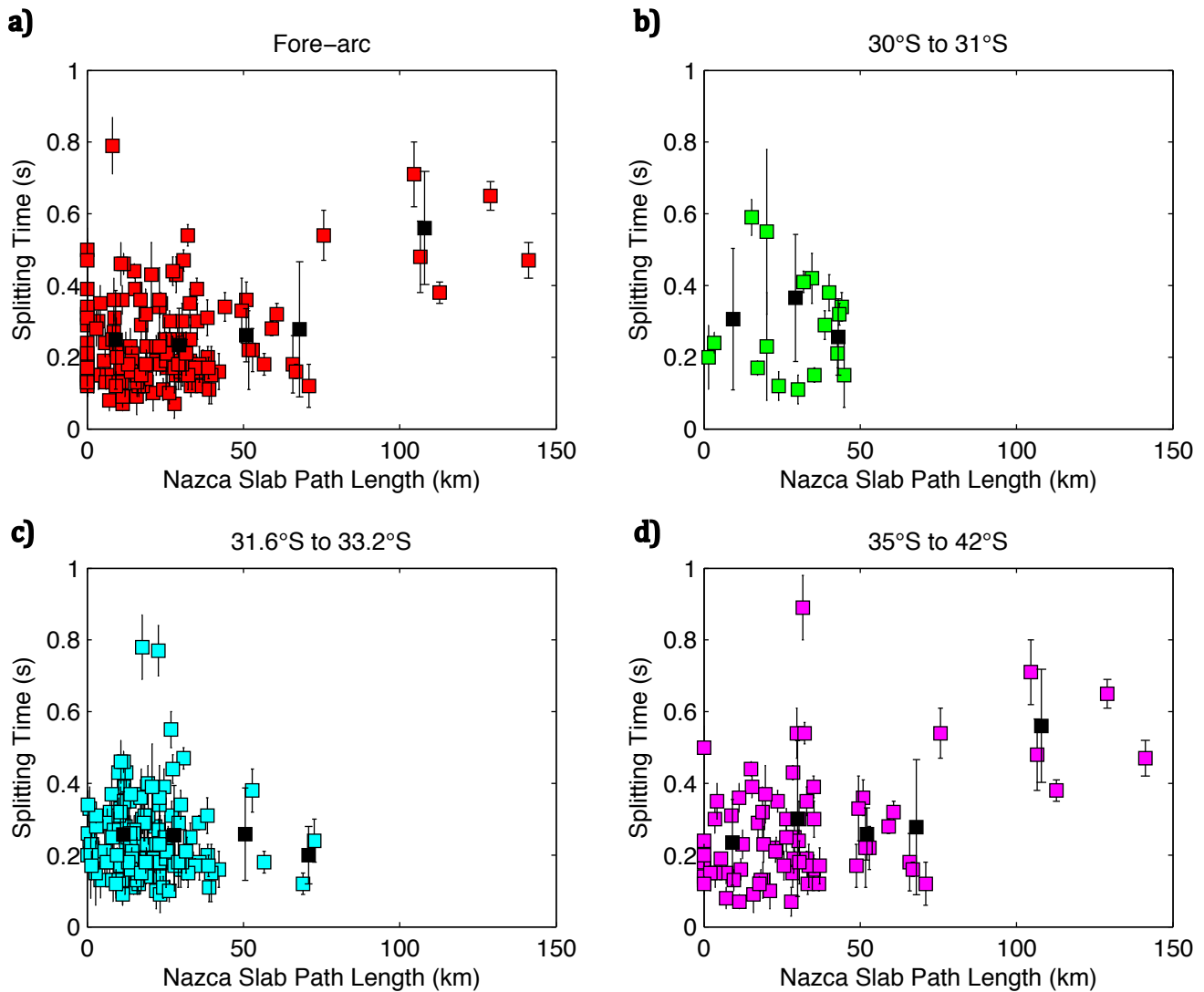


Figure 9. Plots of local S splitting times versus phase path length in the slab. Black squares represent error-weighted splitting time averages, binned in 20 km intervals of path length. Averages for bins with only one event are not displayed. Results are shown separately for stations in the fore-arc (a), and for fore-arc, arc and back-arc stations at latitudes of 30°S to 31°S (b), 31.6°S to 33.2°S (c), and 35°S to 42°S (d).

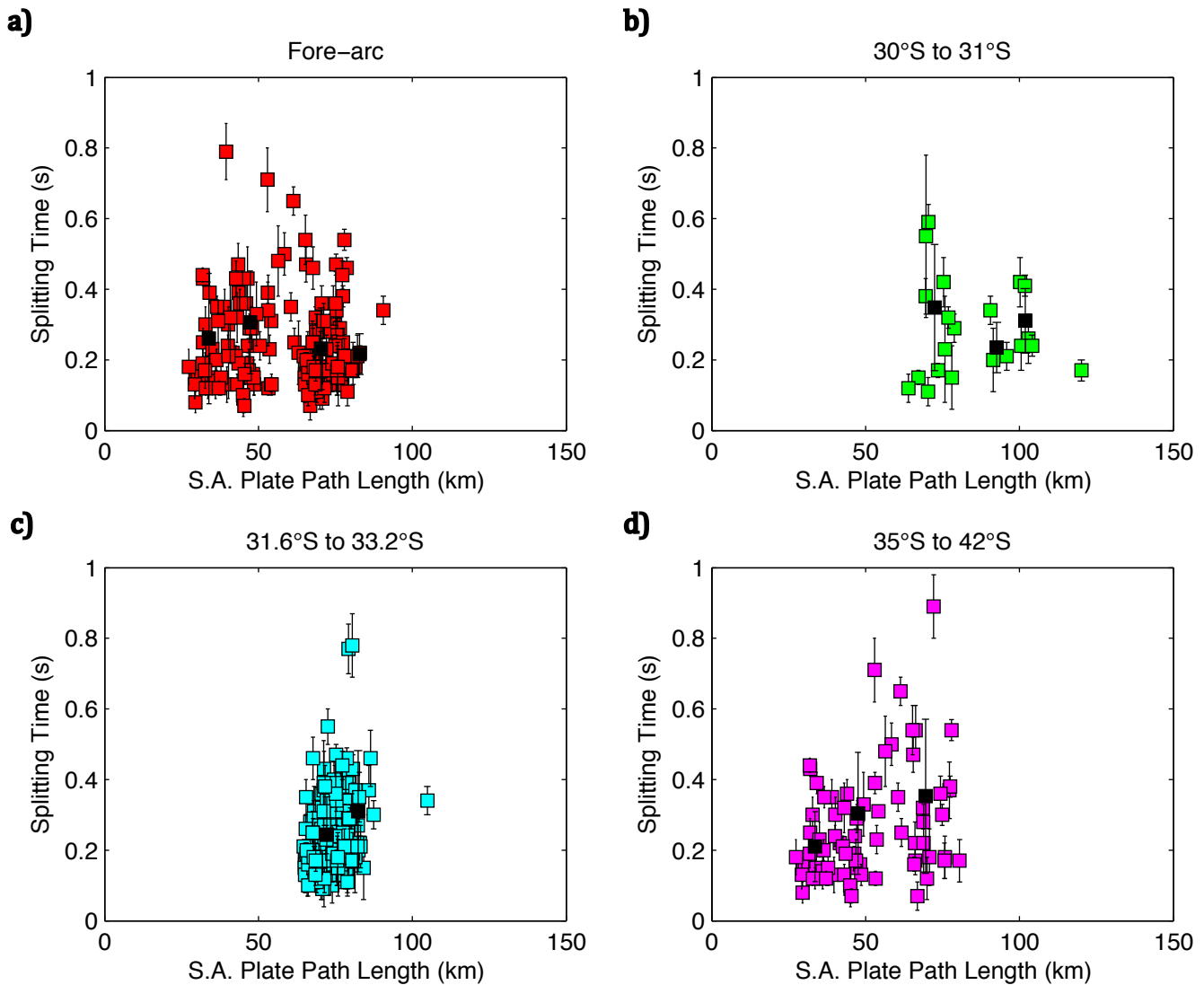


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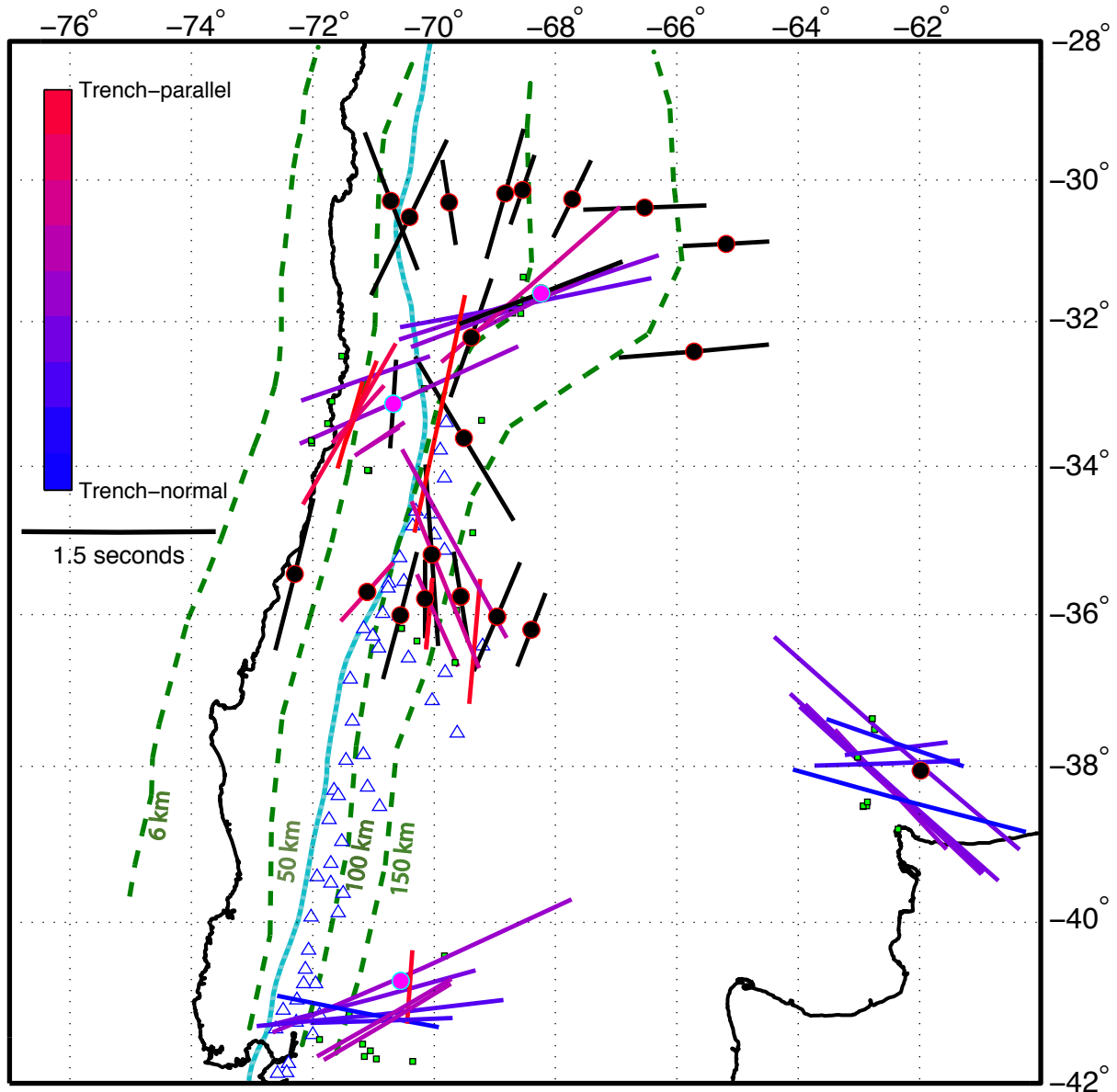


Figure 11. SK(K)S shear-wave splitting results. The split colors scale from trench-parallel (red) to trench-normal (blue). Splits are plotted halfway between where their ray path crosses into the upper mantle at 410 km (green squares) and the station. Black vectors show station-averaged SK(K)S measurements from Anderson et al. (2004). Black circles are CHARGE stations, and pink circles are permanent stations TRQA, PLCA, and PEL. Slab contours from Syracuse and Abers (2006) are displayed, as well as the light blue dividing line between fore-arc/arc and back-arc.

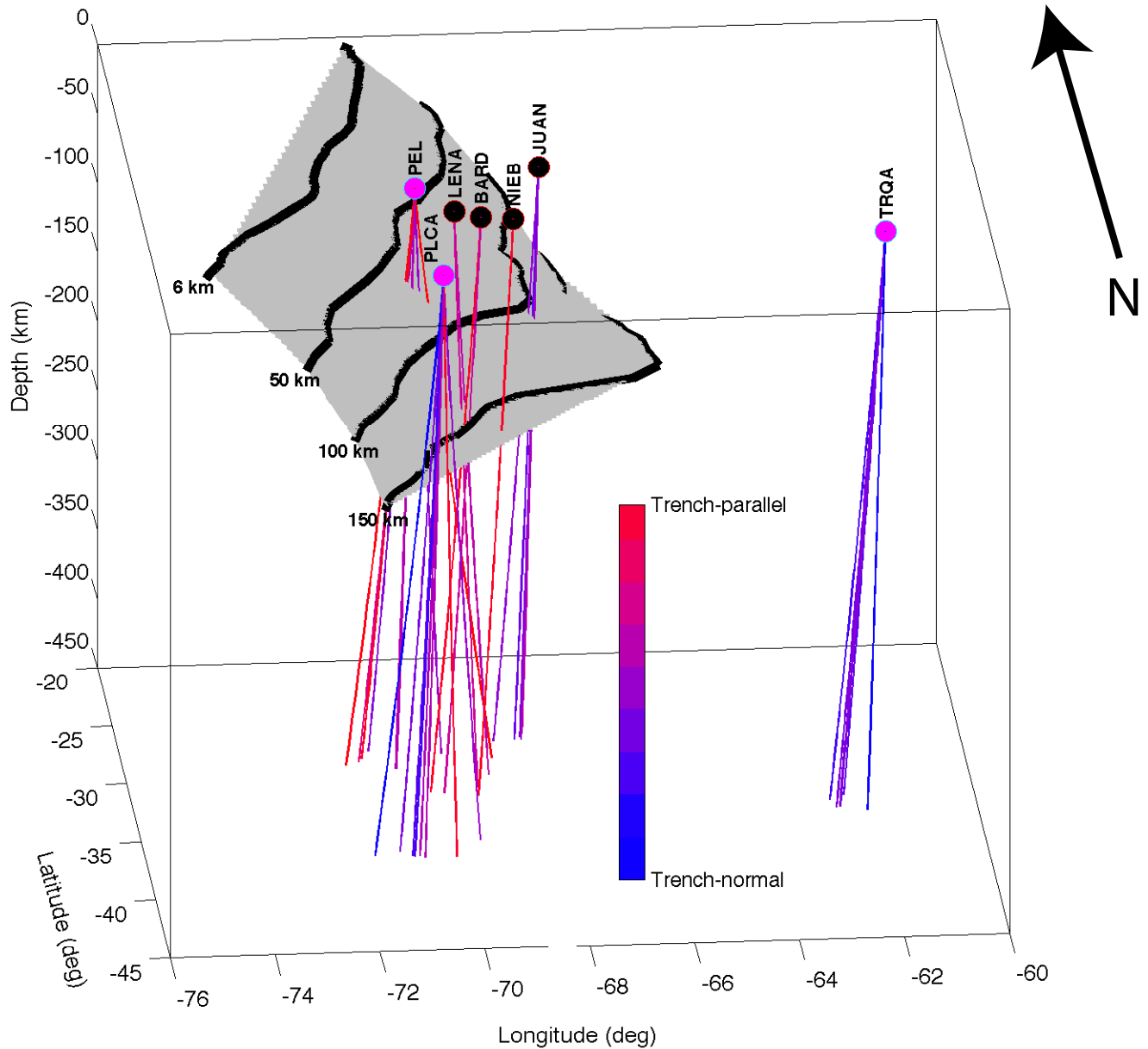


Figure 12. Ray paths for phases with well-constrained SK(K)S splitting measurements, relative to the upper surface of the slab (in grey) as defined from seismicity from Syracuse and Abers (2006). Rays are color-coded by fast polarization orientation. Black circles are CHARGE stations, and pink circles are permanent stations TRQA, PLCA, and PEL.

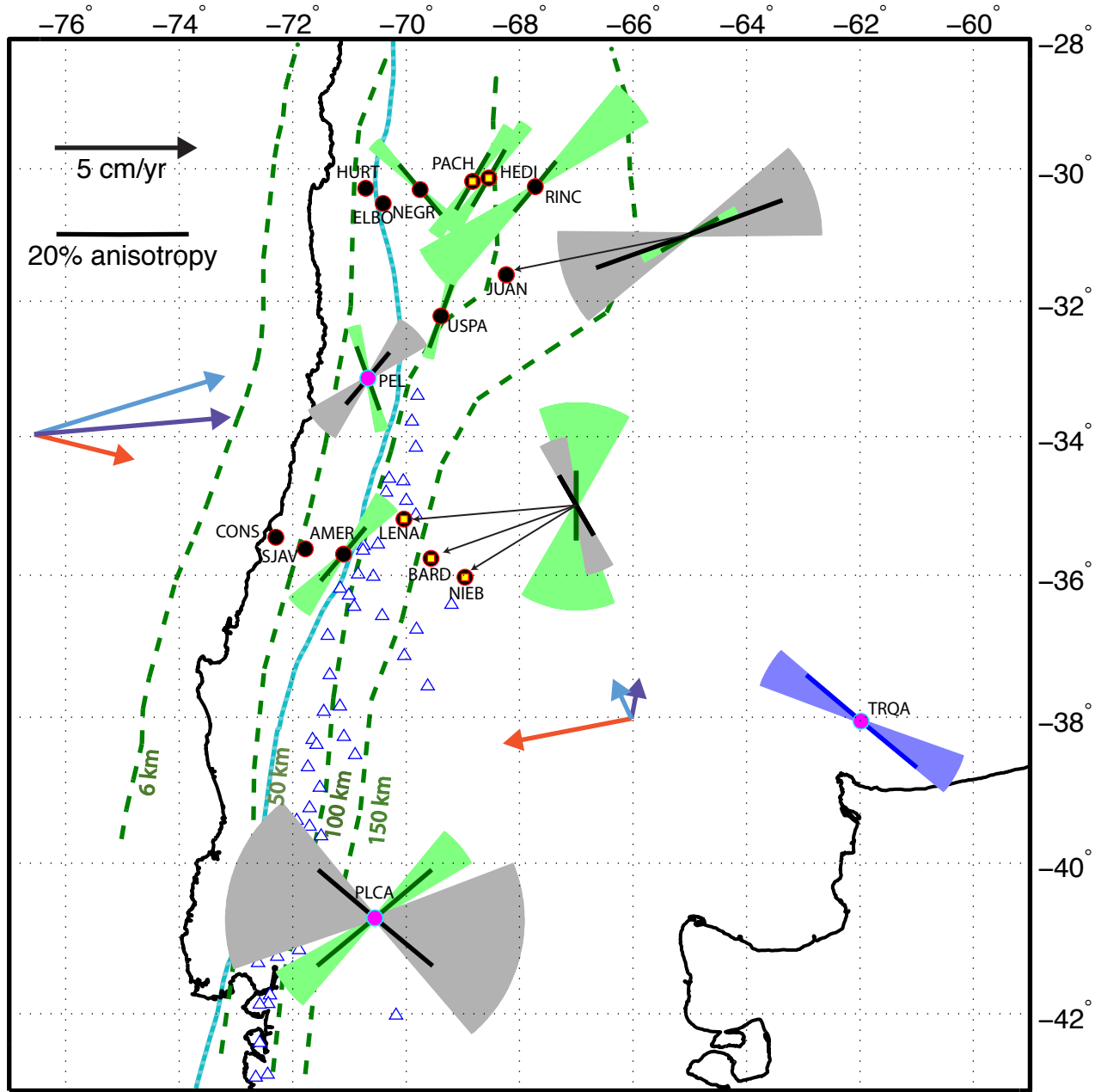


Figure 13. Best-fitting anisotropy models from local S phases at each station (green bars), from the sub-slab models (SK(K)S phases corrected for local S anisotropy; black bars) and for total SK(K)S splitting at TRQA (blue bar). The orientation of each line from North represents the best-fitting a-axis azimuth (θ). The length of the line represents the best-fitting percentage of single crystal anisotropy (α), rather than the total percent of anisotropy. Light green areas represent the ranges of θ and α that are within the 95% confidence limits of the best-fitting model for above-slab models, and light grey areas represent the range of θ and α that are within the 95% confidence limits of the best-fitting model for sub-slab models. For the sub-slab anisotropy models (black bars), θ and α are shown for the cases with a-axes that plunge parallel to the down-dip direction of the slab (Table 1). For all other cases, horizontal a-axes were assumed. Slab contours from Syracuse and Abers (2006) are displayed. Absolute plate motion vectors are red for Pacific hotspot (HS3-NUVEL1a, Gripp and Gordon, 2003), blue for Indo-Australian hotspot (O'Neill et al., 2005; Schellart et al., 2008), and purple for “no net rotation” (Kreemer et al., 2003) reference frames. The light blue line represents the division between fore-arc/arc and back-arc.

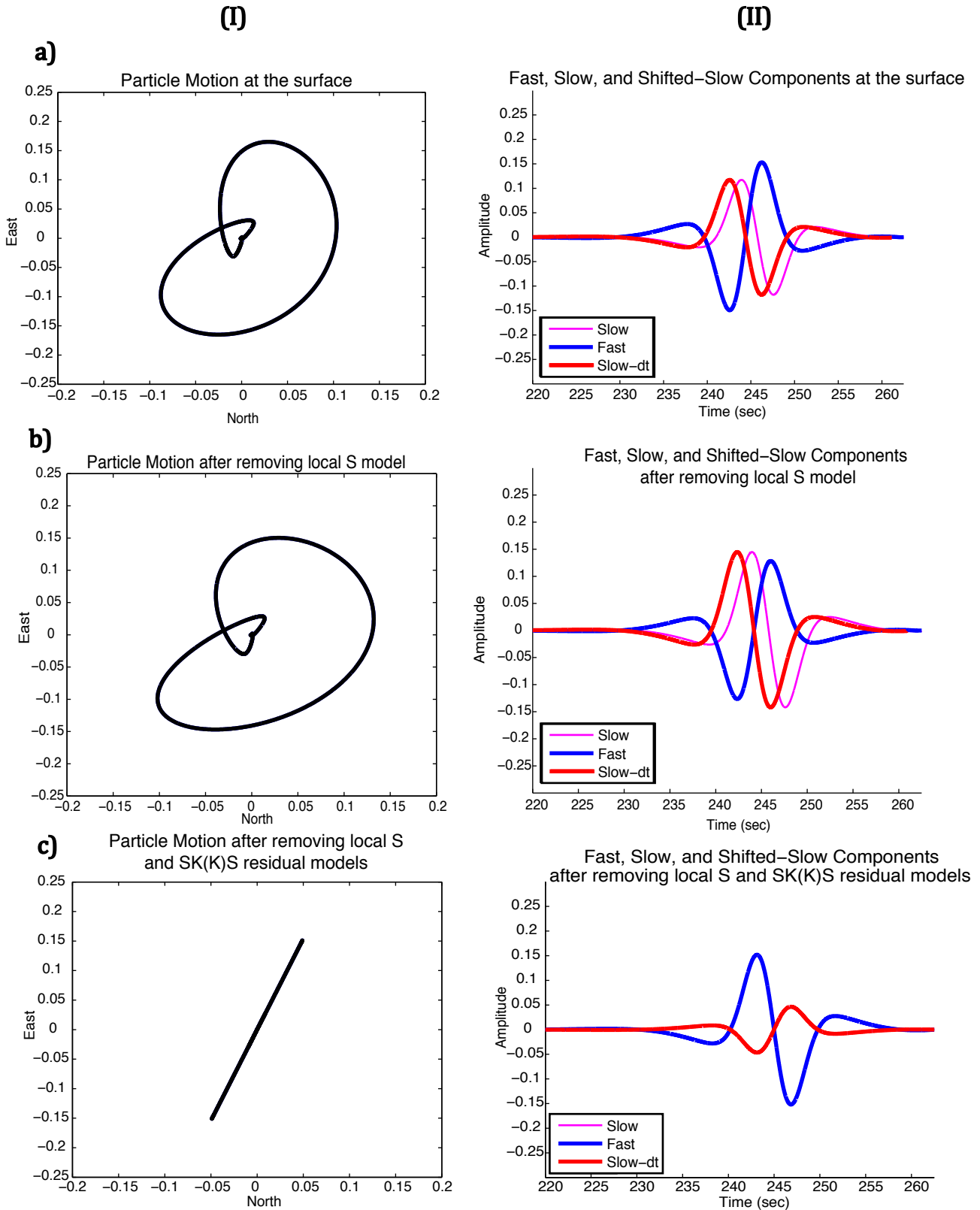


Figure 14. Example of “un-splitting”. A synthetic SKS waveform was propagated along the path for a real phase through a two-layer model for PLCA whose lower layer contains the best-fitting residual SK(K)S model for the mantle below the slab interface ($\theta=130^\circ$, $\alpha=20\%$, $\delta=38^\circ$) and whose upper layer contains the best-fitting local S model for the mantle wedge ($\theta=50\%$, $\alpha=20\%$). Synthetic particle motions (Column I) and waveforms (Column II) are shown for: a) the surface; splitting parameters are $\phi=-72^\circ\pm5^\circ$ and $dt = 1.36 \text{ s} \pm 0.10 \text{ s}$; b) the slab interface, after removal of splitting from the best-fit local S mode; splitting parameters are $\phi=-62^\circ\pm5^\circ$ and $dt = 1.61 \text{ s} \pm 0.10 \text{ s}$; splitting time has increased because the θ values in the two layers are nearly orthogonal; and c) the source, after removal of splitting from both the best-fitting local S and residual SK(K)S models; as expected, particle motion at the source is linear.

Station	Type of model	Best-fit			95% confidence		
		θ (°)	α (%)	δ (°)	θ range (°)	α range (%)	
AMER	Local S		40	10	0 30-50	10-20	
BARD, LENA, NIEB	Grouped Local S		0	10	0 -20-30	10-30	
BARD, LENA, NIEB	Grouped SK(K)S residual		150	10	0 140-180	10-40	
BARD, LENA, NIEB	Grouped SK(K)S residual		-30	10	28 150-170	10-20	
CONS	Local S		50	20	0 -10-90	10-50	
ELBO, HURT	Grouped Local S		0	10	0 0-180	10-50	
HEDI, PACH	Grouped Local S		30	10	0 30-40	10-20	
JUAN	Local S		60	10	0 60-70	10-16	
JUAN	SK(K)S residual		70	30	0 50-80	10-40	
JUAN	SK(K)S residual		70	30	14 50-90	10-40	
NEGR	Local S		140	10	0 130-140	10-20	
PEL	Local S		160	10	0 160-170	10-16	
PEL	SK(K)S residual		50	10	0 40-50	10-20	
PEL	SK(K)S residual		40	10	26 30-60	10-20	
PLCA	Local S		50	20	0 40-60	10-30	
PLCA	SK(K)S residual		100	30	0 80-120	20-60	
PLCA	SK(K)S residual		130	20	38 70-140	10-40	
RINC	Local S		40	10	0 40-60	10-40	
SJAV	Local S		40	10	0 10-110	10-50	
TRQA	SK(K)S		130	20	0 110-130	20-30	
USPA	Local S		20	10	0 12-22	10-13	

Table 1. Best-fit modeling values. θ is the a-axis azimuth, measured from North. α is the strength of anisotropy measured as a percentage of single crystal values. δ is the positive downward plunge in the direction of the best-fitting a-axis azimuth. 'Grouped' indicates that a station did not have enough shear-wave splitting measurements to provide a well-constrained model, and that it was modeled in combination with nearby stations. 'SK(K)S residual' indicates the best-fit sub-slab a-axis orientation and strength of anisotropy after the best-fit above-slab model has been removed. 'SK(K)S' indicates that there were no local S measurements available at that station. All 'SK(K)S residual' cases have best-fit parameters for both the non-plunging ($\delta = 0^\circ$) and plunging cases.

Chapter 2

The Impact of Slab Dip Variations, Gaps, and Rollback on Mantle Wedge Flow: Insights from Fluids Experiments

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Summary

Observed seismic anisotropy and geochemical anomalies indicate the presence of three-dimensional flow around and above subducting slabs. To investigate how slab geometry and velocity affect mantle flow, we conducted a set of experiments using a subduction apparatus in a fluid-filled tank. Our models comprise two independently adjustable, continuous belts to represent discrete sections of subducting slabs that kinematically drive flow in the surrounding glucose syrup that represents the upper mantle. We analyze how slab dip (ranging from 30° to 80°), slab dip difference between slab segments (ranging from 20° to 50°), rates of subduction (4 to 8 cm/yr), and slab/trench rollback (0 or 3 cm/yr) affect mantle flow. Whiskers were used to approximate mineral alignment induced by the flow, as well as to predict directions of seismic anisotropy. We find that dip variations between slab segments generate three-dimensional flow in the mantle wedge, where the path lines of trenchward moving mantle material above the slab are deflected towards the slab segment with the shallower dip. The degree of path-line deflection increases as the difference in slab dip between the segments increases, and, for a fixed dip difference, as slab dip decreases. In cases of slab rollback and large slab-dip differences, we observe intrusion of sub-slab material through the gap and into the wedge. Flow through the gap remains largely horizontal before eventual downward entrainment. Whisker alignment in the wedge flow is largely trench-normal, except near the lateral edges of the slab where toroidal flow dominates. In addition, whisker azimuths located above the slab gap deviate most strongly from trench-normal orientations when

slab rollback does not occur. Such flow field complexities are likely sufficient to affect deep melt production and shallow melt delivery. However, none of the experiments produced flow fields that explain the trench-parallel shear-wave splitting fast directions observed over broad arc and back-arc regions in many subduction zones.

Keywords: Seismic anisotropy, Dynamics of lithosphere and mantle, Subduction zone processes

1. Introduction

Subduction drives plate tectonics and greatly influences mantle flow patterns and melt pathways between the mantle wedge and volcanic arcs. The results of previous geological, geophysical, and geodynamic studies have provided essential insight on subduction zone mantle flow. Early analytical (Tovish *et al.* 1978) and numerical models of subduction defined a mode of plate-induced, two-dimensional (2-D) corner flow in the mantle wedge, including trench-normal flow of material towards the trench and its subsequent entrainment with the subducting slab (e.g. Davies, 1977; Garfunkel *et al.* 1986; Ribe, 1989a). This style of 2-D modeling has revealed important relationships between subduction parameters and slab thermal evolution (Kincaid and Sacks, 1997; Peacock, 1996; van Keken *et al.* 2002; Peacock, 2003; Syracuse *et al.* 2010). Additional 2-D cases consider how corner flow patterns, temperatures, and melt production in the mantle wedge are influenced by slab rollback (Garfunkel *et al.* 1986), back-arc extension (Ribe, 1989a; Kincaid and Hall, 2003; Conder *et al.* 2002), non-Newtonian rheology (van Keken *et al.* 2002; Kelemen *et al.* 2003), and viscosity anisotropy (Lev and Hager, 2011). 2-D models have also been used to define a type of secondary convection in which buoyant diapirs (for example, partially molten hydrated mantle and/or hydrated mantle plus slab crust and sediment rising off of the subducting slab) interact with plate-scale flows within the mantle wedge (e.g. Gerya and Yuen, 2003; Gerya *et al.*, 2004).

Other geochemical, seismic, and geodynamic studies have emphasized the three-dimensional (3-D) nature of subduction zone flow and related processes. Geochemical

data indicate 3-D circulation of distinct mantle material with anomalous compositions around slab edges or through slab gaps (Klein and Karsten, 1995; Karsten *et al.* 1996; Turner and Hawkesworth, 1998; Faccenna *et al.* 2005; Falloon *et al.* 2007; Regelous *et al.* 2008; Gazel *et al.* 2011). A number of observational studies call for entrainment of Samoan plume material along a trench-parallel trajectory from the north to interact with the Tonga subduction zone and the Lau back-arc spreading center (Wendt *et al.* 1997; Turner and Hawkesworth, 1998; Hart *et al.* 2004). Similarly, geochemical models for the Central American subduction zone invoke the opening of a slab window (Abratis and Worner, 2001), toroidal flow around the retreating Nazca plate and into the wedge (Herrstrom *et al.* 1995), and complex entrainment of Galapagos plume material into the fore-arc beneath Central American arc volcanoes (Gazel *et al.* 2011). Spatial-temporal trends in arc geochemical anomalies have also been used to argue for trench-parallel transport in the mantle wedge (Regelous *et al.* 2008; Hoernle *et al.* 2008; Escrig *et al.* 2009).

Seismological constraints on anisotropy in subduction zones also indicate mantle deformation consistent with 3-D flow. B-type olivine fabrics (Jung and Karato, 2001; Jung *et al.* 2006; Karato, 2008; Lassak *et al.* 2006; Long, 2013; Faccenda and Capitanio, 2013) and serpentine fabrics (e.g. Katayama *et al.*, 2009; Jung, 2011) may exist in the cold corner of the mantle wedge (Kneller *et al.* 2005; Kneller *et al.* 2007), and anisotropy in the slab may be influential on certain paths near the trench (Hammond *et al.* 2010; Healy *et al.* 2009). However, outside of these regions, shear-wave splitting observations, including trench-parallel fast polarizations, provide evidence for 3-D flow in the mantle

wedge (Yang *et al.* 1995; Fouch and Fischer, 1996; Audoiné *et al.* 2000; Smith *et al.* 2001; Levin *et al.*, 2004; Long and van der Hilst, 2006; Long and Silver, 2008; Hoernle *et al.*, 2008; Abt *et al.*, 2009; MacDougall *et al.* 2012; Long, 2013), the sub-slab mantle (Russo and Silver, 1994; Fouch and Fischer, 1996; Audoiné *et al.* 2000; Matcham *et al.* 2000; Anderson *et al.* 2004; Civello and Margheriti, 2004; Baccheschi *et al.* 2007; Long and Silver, 2008; Christensen and Abers, 2010; Abt *et al.* 2010; Hicks *et al.* 2012; MacDougall *et al.* 2012; Long, 2013), as well as flow through slab gaps (e.g. Russo *et al.* 2010) and around slab edges (e. g. Peyton *et al.* 2001).

While 2-D models have proven valuable, numerous geodynamical studies have also shown that mantle flow above, below, and around subducting slabs is fundamentally 3-D, thus motivating our experiments with 3-D flow driven by kinematic subduction. Models representing dynamic subduction of discrete slab segments show that rollback occurs naturally and produces a strong toroidal flow of sub-slab mantle around the slab edges and into the mantle wedge (Kincaid and Olson, 1987; Schellart 2004; Funiciello *et al.* 2003; Funiciello *et al.* 2006; Piromallo *et al.* 2006; Honda, 2008; Honda, 2009; Faccenda and Capitanio, 2012; Billen and Jadamec, 2012; Faccenda and Capitanio, 2013). The length scales of toroidal and poloidal flow are positively correlated with slab width (Funiciello *et al.* 2006; Piromallo *et al.* 2006; Schellart 2004; Becker and Faccenna, 2009; Li and Ribe, 2012; Faccenda and Capitanio, 2013) and slab viscosity (Piromallo *et al.* 2006; Jadamec and Billen, 2012; Billen and Jadamec, 2012). Varying dip values and rollback rates can cause time-variable mantle flow patterns both above and below the slab (Kincaid and Griffiths, 2004). These flow patterns control the spatial and

temporal development of anisotropic fabrics in the wedge (e.g. Buttles and Olson, 1998; Druken *et al.* 2011), a fact that we exploit in our experiments in an effort to understand the origin of anisotropy with a trench-parallel fast direction in the mantle wedge.

3-D kinematic models suggest that rollback-induced mantle flow fields deform and stall thermally buoyant upwellings, effectively converting these density and viscosity anomalies into passive features embedded within large scale, plate-driven circulation (Kincaid *et al.* 2013; Druken *et al.* 2013). These models have also shown that changes in subduction style, from down-dip-only motion to rollback sinking, can produce strong variations in slab thermal evolution, buoyant upwelling rates, and temperatures throughout the wedge (Kincaid and Griffiths, 2003; Kincaid and Griffiths, 2004; Kincaid *et al.* 2013). In addition, 3-D slab morphologies yield significantly 3-D local flow patterns (Kneller and van Keken, 2007; Kneller and van Keken, 2008; Jadamec and Billen, 2012), which affect the trajectories of buoyant diapirs or downwellings interacting with subduction-driven flow in the wedge (Hall and Kincaid, 2001; Behn *et al.*, 2007; Zhu *et al.* 2009; Hasenclever *et al.* 2011). In these modeling studies, however, mantle wedge deformation capable of producing trench-parallel fast polarizations is relatively localized near the causative slab dip variation or zone of upwelling/downwelling, and does not explain the wider geographic range of observed trench-parallel shear-wave splitting fast polarizations.

In our laboratory experiments, we explore how slab dip variations, slab gaps, subduction rate, and slab rollback influence 3-D patterns of mantle wedge flow. Our

results show that along-trench dip variations lead to significant 3-D complexity in wedge flow. Both the strength of the 3-D flow and the flux of sub-slab mantle through the gap into the wedge increase as the slab gap size grows and the shallower slab dip decreases. We also quantify the effects of 3-D mantle flow, including cases with rollback, on both wedge anisotropy and on the trajectories of buoyant diapirs rising off of the subducting slab. While we are unable to reproduce the wide extent of observed trench-parallel shear-wave splitting fast polarizations (e.g. Fouch and Fischer, 1996; Long and Silver, 2008; Abt *et al.*, 2009), our results highlight strong 3-D time-dependence in mantle flow and anisotropy, indicating that the interpretation of seismic and geochemical observations requires a 3-D geodynamic context.

2. Apparatus and Experimental Procedure

A laboratory apparatus is used to model upper mantle flow in response to subduction (Figs 1c and 1d). The working fluid is a concentrated glucose solution contained within a 150 cm x 75 cm x 45 cm plexiglass tank. The volume of fluid ($4.5 \times 10^5 \text{ cm}^3$) is maintained at constant temperature, resulting in isothermal density and dynamic viscosity values of 1.42 g/cm^3 and 150 Pa s at 22°C , respectively (Kincaid and Griffiths, 2003).

Kinematic or dynamic forcing may be used to generate slab movement, and in our experiments, subduction is represented kinematically. We employ kinematic subduction for a number of reasons: 1) precise control of relative percentages of down-dip and

rollback slab motion, 2) control over slab dip and changes in slab dip along the trench, 3) the ability to repeat experiments while re-focusing data collection strategies, and 4) characterizing wedge response to a systematic variation in a key parameter, while holding others parameters fixed. The dominant mode of mantle flow in subduction zones is driven by the super-adiabatic temperature gradient, bound up in the viscous, tabular subducting slab. Slab motion therefore drives a mode of forced convection in an upper mantle that, to first order, is free of competing super-adiabatic gradients capable of driving flows of a similar magnitude.

Down-dip slab motion is produced using reinforced rubber belts moving around rollers, similar to the method of Hall and Kincaid (2001). The rollers are located at depths equivalent to the upper-lower mantle transition (D_{670}) and the core-mantle boundary (D_{CMB}) (at the base of the tank) (Fig. 1d). Dip changes are accommodated by attaching the rollers to two metal frames that move horizontally; independent translation of the metal frames allows for different dip angles between the two slabs (i.e. belts), creating a slab gap. Rollback motion of the dipping slabs is produced when the large metal frame supporting all belt rollers is drawn backwards at a prescribed, adjustable rate by a low speed, high torque electric motor (e.g. following the method of Buttles and Olson, 1998).

Our experiments follow the design of the Kincaid and Griffiths (2003, 2004) kinematic models, except that the fluid in our experiments is isothermal, whereas Kincaid and Griffiths (2003, 2004) employ surface cooling (i.e. upper thermal boundary layers) to relate subduction style to slab thermal evolution. In these cases, plate rates and modes of

downward motion mimic the range of styles observed in time-evolving, 3-D dynamic subduction models (Kincaid and Olson, 1987; Griffiths *et al.* 1995; Schellart *et al.* 2004; Funiciello *et al.* 2003; Funiciello *et al.* 2006). The kinematic models are scaled to the mantle using the dimensionless ratio of advection to diffusion, written as the Péclet number,

$$Pe = \frac{U_D D}{\kappa} \quad (1)$$

where U_D is the down-dip subduction velocity, D is the upper mantle depth scale, and κ is thermal diffusivity. Time and velocity scales are determined by relating mantle and lab depth scales, including the deepest extent of the dipping part of the slabs at 670 km ($D_{670} = 670 \text{ km} \approx D_L = 14 \text{ cm}$), and known values for thermal diffusivity ($\kappa_{\text{mantle}} = 10^{-2} \text{ cm}^2/\text{s}$; $\kappa_{\text{lab}} = 10^{-3} \text{ cm}^2/\text{s}$; Kincaid and Griffiths, 2004). One minute of lab time is equivalent to 5 my of geologic time, and a lab plate rate of 1 cm/min equals a real subducting plate rate of 1 cm/yr.

While we match the length scales and plate rates of Kincaid and Griffiths (2003), because of the added complexity of variable slab morphology, we choose to begin more simply with the assumption of an isothermal fluid. Motion of the belt generates a velocity boundary layer. However, without thermal gradients, these models do not include the chilled margin of the slab or a cold viscous blanket that characteristically thickens with depth. The lack of thermal gradients, while resulting in a smaller slab thickness, is not expected to significantly alter far-field patterns, because they scale with the half width of

the plate, measured in the trench-parallel direction (Buttles and Olson, 1998; Kincaid and Griffiths, 2003). A more immediate effect is that, with trench-parallel dip variations, the lack of a thermal boundary layer produces a larger slab gap size than would be present with greater slab thicknesses.

Down-dip plate velocity is modeled with independent control of speed (U_D) and dip angle (Ω) for two separate belt systems. The speed is controlled by high torque, low speed motors driving the belts around a series of rollers at either 4 or 8 cm/min (4 or 8 cm/yr). In rollback cases, a translation rate of 3 cm/min (3 cm/yr) is held constant over a total horizontal distance of 25 cm (1250 km). The total trench-parallel width (W) of the plate is 25 cm (1250 km) (Fig. 1c). Experimental slab dip values range from 30° to 80° , and slab gap sizes, measured as differential dip angle between belts ($\Delta\Omega$), vary from 20° - 50° .

The slab edges are 25 cm (1250 km) from the tank sidewalls, or roughly twice the length scale of the toroidal flow (Kincaid *et al.*, 2013). As defined in Funiciello *et al.* (2003), the ratio between the tank width and the slab width is large enough for the flow to be ‘volumetrically unconstrained’; the distance between the sidewalls of the tank and the belts is great enough that flow fields, both around the sides of the plate and in the mantle wedge, are not distorted by edge effects due to interaction of the glucose syrup with the walls of the tank (Funiciello *et al.*, 2006; Schellart *et al.*, 2007; Schellart *et al.*, 2011).

Passive Lagrangian markers (whiskers, beads, and micro-bubbles) distributed throughout and moving within the fluid are tracked to provide velocity information in both time and space. Small diameter (~ 3 mm), nearly passive Delrin beads (rise rates of the beads through the fluid are $\sim 1\%$ of U_D) are used to track flow rates and patterns within the wedge and to visualize sub-slab flow through the gap and into the mantle wedge. Whiskers measuring roughly 5 mm in length and 0.1 mm in diameter provide information on velocity and potential olivine lattice preferred orientation (LPO) within the wedge. Large aspect ratio cylinders (whiskers) rotating in viscous shear flows have been shown to follow orientation distribution functions with integrated strain (Jeffery, 1922; Buttle and Olson, 1998) that closely match theoretical orientation distribution functions derived for the a -axis of olivine (Ribe, 1989b). Whiskers cannot reproduce the potentially re-orienting effects of dynamic recrystallization, but do provide a proxy for the amount and direction of olivine a -axis alignment in the absence of dynamic recrystallization. Microbubbles (~ 1 mm in diameter) introduced into the fluid during whisker emplacement act as velocity markers, and are completely passive (i.e. do not rise buoyantly) on the timescales of these experiments.

Wedge flow and whisker alignments are tracked within three 1 cm thick horizontal light sheets centered at depths below the fluid surface corresponding to: $z_1 = 2$ cm (100 km), $z_2 = 4$ cm (200 km), $z_3 = 7$ cm (350 km) (Fig. 1). A single vertical light sheet (e_1) oriented along the slab centerline (the trench-normal line that intersects the trench at the gap between the two belts) provides information on vertical velocity (Fig. 1b). Light sheets are produced by light from a slide projector shining through thin

horizontal or vertical slits. Covers on the slits are sequentially opened and closed to illuminate individual light sheets. Images are captured at five-second intervals using high-resolution cameras oriented normal to each light sheet (Figs 1a and 1b). Continuous cycling allows data to be collected on each sheet at 20-second intervals. Velocities for individual markers are calculated by dividing digitized distance changes between images by the known time interval. Flow patterns are represented as material path lines, or the tracks of parcels of wedge fluid. In time-evolving flows, path lines are distinct from streamlines, the latter of which represent flow at a specific instant in time. The majority of cases employ a stationary trench; however, in an experiment where we add slab rollback (e.g. Exp. 16, Table 1) the top-view camera is coupled to the frame, which translates at the rollback rate. For comparison to the experiments with a stationary trench, position data from the rollback experiment are converted into position data in a fixed reference frame by adding distance for a constant rollback rate of 3 cm/min.

3. Experimental Results

In a set of seventeen experiments (Table 1), we explore the relationship between complex slab morphology and mantle wedge flow patterns. We begin by describing wedge flow for reference cases where both belts have the same dip, progress to cases where slab dip varies between belts, and finally include the effects of trench rollback. Whisker orientations are described at the end of this section.

To describe the experiments, flow patterns are summarized by dividing the wedge into source zones that define the starting points of groups of markers (i.e. whiskers, bubbles, or beads) with similar path lines. In addition, we categorize the normalized trench-normal distance (x) as the fore-arc ($x = 0-0.25$), arc ($x = 0.25-0.5$), and back-arc ($x > 0.5$). Velocities are described by the dimensionless marker speed $u^* = \mathbf{u}(\mathbf{x},t) / U_D$, the measured velocity along the path line ($\mathbf{u}(\mathbf{x},t)$) normalized by the subduction rate (U_D); $u_{\perp}^*$ and $u_{\parallel}^*$ are the trench-normal and trench-parallel components of the marker speed, respectively. The approach angle (A_w) is the angle of the path line relative to trench-normal (trench-normal: $= 0^\circ$ and 180° , trench-parallel: $= +/- 90^\circ$); the normalized approach angle A_{norm} is equal to A_w divided by 180° . These parameters were measured over a range of normalized distances from the trench, where s_{norm} is the (trench-normal) distance from the trench divided by the total slab width of 25 cm. Plots of $u_{\perp}^*$, $u_{\parallel}^*$, and A_{norm} with distance from the trench for select experiments are shown in Appendix A. The slab gap, when present, is located along the centerline, and in the figures, the steeper slab is always on the right, and the shallow slab is always on the left.

3.1 *Wedge flow: Down-dip reference cases*

Results are first presented for reference cases simulating subduction of a single, constant-dip plate segment, without rollback. For example, we consider an experiment with an intermediate slab dip (50°) at two different depths (Figs 2a and 2b, Exp. 1). In the shallowest (z_1) light sheet, equivalent to a depth of 100 km from the top of the mantle wedge, subduction drives flow towards the trench through the central portion of the

wedge (Zones 2, 3; Fig. 2a) along path lines that are mostly trench-normal ($A_w \sim 0$ to $\pm 10^\circ$) before markers are entrained with the subducting slab. This flow pattern is similar to those of 2-D corner flow models. Wedge circulation differs near the slab edges (Zones 1 and 4), where fluid moves along trench-parallel approach angles ($A_w \sim \pm 80$ - 90°) before turning downward with entrained flow above the slab. Some of the fluid is supplied from behind the plane of the slab surface ($A_w > 90^\circ$), as shown in Fig. 2a. These patterns are replicated at depth (for example at z3, an equivalent depth of 350 km, in Fig. 2b), but with reduced marker speeds. Maximum marker speeds at z1 are nearly three times those at z3, with values of $u^* = 0.35$ - 0.42 and $u^* = 0.15$, respectively. Because of the dipping slab, path lines in z3 turn downward with entrained flow at larger values of x , causing some deeper path lines to be orthogonal to those at shallower depths. These vertical gradients in flow direction and speed can create zones of high shear.

Wedge flow varies with slab dip (Figs 2c and 2d). For example, a steeper slab dip (70° , Exp. 3) produces flow that is more 2-D (Fig. 2c) than the path lines for a dip of 50° (Exp. 1, Fig. 2a). For a dip of 70° , material is drawn towards the slab along primarily trench-normal flow lines originating from Zones 2 and 3, although near the slab edges (Zones 1 and 4), marker trajectories are oblique to the trench (Figs 2c, A1a and A1f). Shallower dips (30° , Exp. 4, Supplementary Video 1) result in strongly 3-D flow (Figs 2d and A1b). In contrast to Exp. 3, the fastest flow in this experiment enters the wedge along a trench-parallel trajectory near the slab edges (Figs 2d, A1b, A1c, A1e, and A1f), such that both the fore-arc and arc are supplied along path lines emanating from outside the wedge (beyond the edge of the plate or $|y| > 0.5$). Trench-normal flow near the slab

centerline (Zones 2 and 3) is slower and supplies material only to the central third of the slab surface (Figs 2d, A1b, A1c, and A1d). Maximum marker speed (u^*) in the z1 layer increases systematically with slab dip; $u^* = 0.39, 0.45$, and 0.52 for slab dips of 30° , 50° , and 70° , respectively (Fig. A4).

3.2 *Wedge flow: Trench-parallel slab dip variations*

A subset of twelve experiments is used to characterize how trench-parallel changes in slab dip influence wedge flow and the flux of sub-slab material through a slab gap and into the wedge.

Introducing a difference in the dip of the two slab segments produces asymmetry in wedge circulation in which mantle flow towards the trench is deflected towards the slab segment with the smaller dip. This asymmetry occurs even when the change in dip is relatively small ($\Delta\Omega=20^\circ$, Exp. 8, Table 1) (Fig. 3). While most of the material originating in Zone 3 is entrained downward above the steeper slab, some wedge fluid closest to the centerline crosses the centerline and is entrained downward above the shallowly-dipping slab. Flow asymmetry also occurs at greater (z3) (Fig. 3b) depths. As in the constant dip cases, deeper marker speeds are lower, and some z3 path lines are roughly orthogonal to the overlying path lines in z1, indicating significant vertical shear. Path line patterns, including flow asymmetry created by trench-parallel dip variations, show little change as a function of down-dip subduction rate (4 cm/min vs 8 cm/min).

Larger trench-parallel dip variations generate stronger flow asymmetry with greater trench-parallel speeds. For example, at z_1 in Exp. 10 (slab dips of $80^\circ/60^\circ$) (Fig. 4a), only mild flow asymmetry is observed, but in Exp. 7 (slab dips of $80^\circ/30^\circ$) (Fig. 4b, Supplementary Video 2, Table 1), path lines are strongly deflected in the trench-parallel direction towards the slab segment with the shallower dip, and significant amounts of Zone 3 wedge material are drawn across the slab centerline before downward entrainment. This deflection is reflected by the larger $u_{||}^*$ values (Fig. A2e) and Zone 3 A_{norm} values for $s_{\text{norm}} < 0.7$ (Fig. A2f) in Exp. 7 relative to Exp. 10.

The wedge flow asymmetry created by trench-parallel slab dip changes is consistent with 2-D analytical pressure solutions (Fig. A3; Appendix B) and numerical results (Hall *et al.* 2000; Kneller and van Keken, 2007; Kneller and van Keken, 2008). Non-hydrostatic pressures from 2-D corner flow in the mantle wedge are more negative (Fig. A3, top row) and trench-parallel pressure gradients are higher above the shallowly-dipping slab (Fig. A3, middle row). In cases where a slab gap is present, markers in Zones 3 and 4 are responding to three different drivers: the trench-parallel pressure gradient, which is strongest in the fore-arc and pulls wedge material from the steeply-dipping slab to the shallowly-dipping slab (Fig. A3), the subduction velocity boundary condition imposed by encountering the part of the shallow belt close to the slab centerline, and the normal entrainment flow of the steeply dipping slab.

Maintaining a similar slab dip variation, but steepening both plate dips leads to weaker trench-parallel flow close to the slab and the slab centerline (Fig. 5). This trend is

evident when comparing Exp. 10 (dips of 80°/60°) (Fig. 4a), Exp. 8 (dips of 65°/45°) (Fig. 3a), and Exp. 5 (dips of 50°/30°), which have maximum $u_{||}^*$ values of 0.17, 0.21, and 0.27, respectively (Figs 5a and 5b, Table 1). The experiments with the same dips but a down-dip velocity of 4 cm/min (Exps 11, 9, and 14) produce a similar $u_{||}^*$ trend (0.12, 0.22, and 0.32, respectively). However, steepening both dips does not yield clear trends in total maximum marker speed (u^*) (Figs A4a and A4b, Table 1), as seen in Exps 10, 8, and 5 ($U_D = 8$ cm/min), where u^* values are 0.61, 0.61 and 0.51, and for Exps 11, 9, and 14 ($U_D = 4$ cm/min), where u^* values are 0.63, 0.53 and 0.57.

3.3 *Wedge flow: Slab rollback*

Trench/slab rollback significantly modifies the 3-D asymmetric flows generated by the presence of trench-parallel dip variations. In Exp. 16 (Supplementary Video 3, Table 1, Fig. 6), 3 cm/min of rollback is added to the case with dips of 80°/30° and a U_D of 8 cm/min (Exp. 7, Table 1). With slab rollback, shallow (z_1) path lines (Fig. 6a) are still highly asymmetric, with path lines from Zones 3 and 4 crossing the slab centerline before turning downward with entrainment above the more shallowly-dipping slab. The curvature of path lines in Zone 3 moving towards the shallow slab in Exp. 16 (clockwise in map view) has the opposite sense from Zone 3 transport paths in the similar, but non-rollback case of Exp. 7 (counterclockwise in Fig. 4b). This difference in curvature is reflected in normalized approach angle values for Zone 3 in both experiments (Fig. A5f). Both trench-parallel (Fig. A5e) and trench-normal (Fig. A5d) flow rates are significantly higher in the case with rollback.

Rollback also drives vigorous toroidal flow around the edges of the slab at greater depths in the wedge (e.g. level z3 in Fig. 6b). Fluid parcels close to the retreating slab in Zone 1 at level z3 first move out from the wedge with strong trench-parallel motion and then back into the wedge after passage of the slab. Once the slab passes, deeper path line trajectories (z3) are more similar to those at z1. As in previous cases, marker speeds at z3 are much lower than those at z1 (e.g. marker 3a in Fig. A5c).

3.4 *Wedge flow: Summary of flow velocities and drivers*

In summary, $u_{||}^*$ values systematically vary as a function of trench-parallel dip variation ($\Delta\Omega$ or slab gap size) (Table 1 and Fig. 5), but u^* behavior is more complex. As previously mentioned, with no slab gap ($\Delta\Omega = 0^\circ$), steeper dips drive greater maximum u^* values in the wedge (Fig. A4a). When trench-parallel slab dip variations exist, maximum u^* is strongly correlated with greater $\Delta\Omega$ (Fig. A4b), but does not systematically increase with greater values of the steepest slab dip (e.g. the u^* values for $\Delta\Omega = 20^\circ$ in Fig. A4a). As for maximum $u_{||}^*$ values for the central wedge closest to the trench (defined by $x \leq 0.25$ and $-0.5 \leq y \leq 0.5$), larger trench-parallel slab dip variations generate greater $u_{||}^*$ values and stronger flow asymmetries (Fig. 5a). When a slab gap is present, and for a given value of $\Delta\Omega$, $u_{||}^*$ increases as the dips of both slabs decrease (see the points for $\Delta\Omega = 20^\circ$ in Figs 5a and 5b). This result is consistent with the increase in trench-parallel pressure gradient predicted for the mantle wedge above more shallowly-dipping slabs (Fig. A3). Finally, the addition of slab rollback increases wedge flow

velocities, and the highest measured u^* value in all of our experiments is found in Zone 4 of Exp. 16 ($u^* = 1.08$, labeled ‘R’ in Figs A4a and A4b), above the steeply-dipping slab. Rollback does not significantly change the maximum u_{II}^* value (Figs. 5a and 5b), indicating that the increase in marker speed close to the retreating slab ($x \leq 0.5$) partitions primarily into trench-normal velocity.

3.5 *Flow through slab gaps*

We measured flow from beneath the slab, through the gap, and into the mantle wedge in order to evaluate the implications of this flow for shear-wave splitting patterns and geochemical data in subduction zone mantle wedges near hypothesized slab tears or gaps (e.g. Russo *et al.* 2010). By seeding the sub-slab mantle material with bubbles and red beads, we observed sub-slab flow penetrating the mantle wedge in all cases where $\Delta\Omega \geq 35^\circ$, and in only one case where $\Delta\Omega = 20^\circ$ (Table 1). We observed three different trajectories of sub-slab material penetrating the mantle wedge (Fig. 7). A number of tracers emerge from beneath the shallow slab and enter the mantle wedge above the shallowly-dipping slab (purple path in Fig. 7). Other markers escape into the mantle wedge above the steeply-dipping slab (blue path in Fig. 7; also Fig. A6a), while others barely enter the mantle wedge before being entrained either above or along the edge of the steeper slab (red path in Fig. 7). Larger gaps increase the likelihood of sub-slab flow through the gap, but do not necessarily cause higher gap flow rates. For example, the highest measured through-gap speed without rollback occurred in Exp. 8, with $\Delta\Omega = 20^\circ$.

Rollback increases rates of mantle flow through the slab gap, as expected when the translating plate displaces sub-slab fluid. In cases with rollback, marker speeds through the gap range from $u^* = 0.19\text{-}0.46$; these values are equivalent to $1.5\text{-}3.7\text{ cm/min} = 1.5\text{-}3.7\text{ cm/year}$ (e.g. Fig. A6b). In cases without rollback, marker speeds through the gap are $u^* = 0.05\text{-}0.36$ (equivalent to $0.2\text{-}2.9\text{ cm/min} = 0.2\text{-}2.9\text{ cm/year}$). Similarly, cases with rollback yield significantly higher volume fluxes (β in Table 1) through the gap ($4.9\text{-}12.0\text{ cm}^3/\text{min} \approx 0.12\text{-}0.30\text{ km}^3/\text{year}$) than those cases without rollback ($0.7\text{-}9.4\text{ cm}^3/\text{min} \approx 0.02\text{-}0.24\text{ km}^3/\text{year}$) (method described in Appendix C).

Larger gap sizes lead to increased intrusion of sub-slab tracers into the wedge. The furthest penetration of sub-slab markers into the mantle wedge prior to downward entrainment (up to 17 cm or $\sim 850\text{ km}$ away from the trench) occurred in a large gap case (Exp. 17) where a bead entered the wedge above the more steeply-dipping plate. We do not observe an upward component of motion for markers traversing the gap, suggesting that other conditions (e.g. possibly the onset of back-arc spreading) are required for gap material to be sampled by decompression melting processes. Contrary to the geochemical observations of Klein and Karsten (1995) and Karsten *et al.* (1996), we find no evidence for penetration of mantle wedge material into the sub-slab mantle.

3.6 *Whisker orientations*

The evolution of whisker azimuths from an initially random distribution provides information on potential seismic anisotropy. As discussed previously, whiskers are good

first-order approximations for LPO formation from olivine a -axis alignments, but do not include the effects of dynamic recrystallization.

With a constant slab dip (Exp. 4, Fig. 8a), whiskers generally rotate to orientations that are parallel to sub-parallel to their path lines. Exceptions to this rule occur close to the slab centerline, where path lines are in general trench-normal and little rotation in the x-y plane occurs, and on select paths that involve toroidal flow close to the edges of the slab (e.g. Zone 1 nearest to the trench) (Fig. 8a). In the case of a large slab gap and no rollback (Exp. 7; Fig. 8b), the zone near the slab centerline is no longer a zone of low whisker rotation, due to the high rates of x-y plane shear involved in the flow drawn over to the shallowly-dip slab as it approaches the trench. With the addition of rollback (Exp. 16; Fig. 8c), the strongest rotation of whisker alignments towards their path lines again occurs in the group of paths that converges above the shallowly-dipping slab (Zones 2 and 3).

Trench-oblique and trench-parallel whisker orientations are generally found on two types of path lines: 1) path lines that are the most deflected from trench-normal, across the slab centerline from the steeply-dipping slab to the shallowly-dipping slab, with rapid rotation along the path, or 2) path lines that bring flow into the wedge from around the edge of the slab. To quantify these patterns, we measured the orientations of every whisker within each horizontal light sheet (z_1 , z_2 , z_3) in one late-stage (time elapsed ≈ 8 min [≈ 40 my]) photograph for two similar experiments, where the only difference was the absence (Exp. 17) or presence (Exps 16 and 20, Table 1) of trench

rollback. Exps 16 and 20 (Exp. 16/20 hereafter) have identical subduction parameters (Table 1), and we have combined their results to obtain more robust whisker orientation distributions.

Whisker azimuths at all three depths of Exp. 17 (Fig. 9, Column I) and Exp. 16/20 (Fig. 9, Column II) are displayed in rose diagrams placed in their corresponding model blocks. Each block measures 5 cm (250 km) in the trench-parallel direction and 6.25 cm (312.5 km) in the trench-normal direction (Fig. 9). Rose diagrams are only plotted for blocks that contain four or more whiskers, with the exception of model block 10 in Exp. 16/20 at z_3 . While this block contained only two whiskers, it was retained to enable comparison with Exp. 17 at a full range of depths.

Most mean whisker orientations lie within 45° or less of trench-normal. More trench-parallel mean whisker alignments occur near the slab edges, where toroidal flow is present (e.g. block 20 in z_1 of Fig. 9, Column II), or close to the slab gap above the steeply-dipping slab where flow is deflected towards the shallowly-dipping slab (e.g. block 14 of Exp. 17, Figs 9 and 10). The observed trench-parallel deflection of whisker orientations near the slab centerline occurs on roughly the same length scale (one block with dimensions of 5 cm by 6.25 cm [250 km by 312.5 km]) as that seen in the ‘trench-parallel stretching’ of Kneller and van Keken (2007, 2008), although our whisker orientations are more trench-oblique than truly trench-parallel. In contrast, the addition of rollback (Exp. 16/20) causes the whiskers near the centerline to have trench-normal alignments, due to the stronger trench-normal flow near the center of the model produced

by toroidal flow (Fig. 9, Column II), in agreement with previous studies (e.g. Funicello *et al.* 2003; Kincaid and Griffiths, 2003; Funicello *et al.* 2006; Druken *et al.* 2011).

To illustrate how whisker orientations at the steep-to-shallow slab transition are affected by rollback, we examined whiskers azimuths in both the back-arc (model block 10) and the arc (model block 14) at the slab centerline for cases both with (Exp. 16/20) and without (Exp. 17) rollback (Fig. 10). Model block 14 is close to the area that experiences the highest marker speeds due to the presence of a slab dip variation (e.g. Fig. 4b); it is representative of flow near the slab gap. Model block 10 is representative of flow further from the slab gap. At all three depths in block 14, the whisker azimuth averages in the case without rollback (Exp. 17) are deflected further away from trench-normal and have larger standard deviations, which represent the wider range of observed whisker azimuths (Fig. 10). The effect of rollback on whisker alignment is relatively localized. Whiskers in the back-arc (block 10) in both experiments are similar, maintaining an orientation just counter-clockwise of trench-normal orientations at all depths, although the whiskers in the case with rollback (Exp. 16/20) have less scatter (Figs 9 and 10). In addition, block 17 whiskers, further away from the gap, in fact become more trench-parallel with rollback (Fig. 9).

Near the slab gap (block 14) in the case with no rollback, whisker alignments rotate significantly with depth in the counter-clockwise direction towards the shallowly-dipping slab (Fig. 10). (The sense of rotation was determined by observing the counter-clockwise rotation of whiskers at the centerline near the slab gap in real time.) In

contrast, with rollback (Exp. 16/20), whisker azimuths in block 14 rotate clockwise with depth (Fig. 10). The implications of these results for shear-wave splitting in subduction zone mantle wedges are explored in the next section.

4. Implications

4.1 *Geochemical anomalies and melt migration*

In all of our experiments with trench-parallel slab dip variations, three types of mantle wedge flow are present: strong toroidal flow, trench-oblique flow towards the shallowly-dipping slab, and slab gap through-flow into the mantle wedge. Geochemical data provide evidence for such flows in real subduction zones. In the Tonga subduction zone, Samoan hotspot magmatic signatures, originating from beneath the subducting Pacific Plate, have been recorded in magmas in the Northern Lau Basin, a back-arc spreading center on the overriding Australian Plate (e.g. Turner and Hawkesworth, 1998). A trench-parallel component of toroidal flow, propagating around the slab edge and laterally into the wedge, is also supported by evidence of mixing between mantle wedge material and Samoan hotspot material further south along the trench (Falloon *et al.* 2007; Regelous *et al.* 2008). Geochemical evidence for trench-parallel flow in the mantle wedge has also been found in the Calabrian (e.g. Faccenna *et al.* 2005) and Central American (e.g. Hoernle *et al.* 2008; Gazel *et al.* 2011) subduction zones. Flow through a slab gap or slab window is also detectable with geochemistry, whether it be the influx of sub-slab mantle material into the mantle wedge after slab detachment (Ferrari, 2004;

Macera *et al.* 2008; Guo *et al.* 2011; Pan *et al.* 2012), or the subduction of a spreading center (D’Orazio *et al.* 2000; Abratis and Worner, 2001; Gutierrez *et al.* 2005; Guivel *et al.* 2006).

We use our experimental results to demonstrate the importance of 3-D flow fields in controlling surface geochemical patterns at subduction zones. We track theoretical diapirs (i.e. material rising vertically off of the slab) moving with imposed buoyant rise rates through our highly heterogeneous wedge flow fields. Exps 3, 8, and 16 (Table 1) are used to develop the diapir trajectory models because they provide sufficient path line data at all three depths. After finding two places in Exps 3, 8, and 16 where we have (close-to) overlapping path lines at all three depths, we place a theoretical diapir at a depth of 375 km and, for three different rise rates ($V_p = 1.5$ cm/yr, 2.5 cm/yr, or 8 cm/yr, equation (2)),

$$V_p = \frac{B}{n_{rise}} \quad (2)$$

we calculate the propagation distance of the diapir along the path line at each depth (3).

$$q_h = q \Big|_{o+n_{rise}} \quad (3)$$

B is the known distance from the bottom of one layer to the next, n_{rise} is the number of time steps it takes for the diapir to rise from one layer to the next (in this analysis, $n_{rise} = 1, 3, \text{ or } 5$), q is the marker path at a given time step, q_h is the horizontal position of the diapir along that path line, and o is the original time step of the marker located closest to the position of the diapir in the previous layer or at the source. As seen in Fig. 11, we find that the trench-parallel (y-direction) offsets between the initial and final diapir locations are as small as ~ 30 km in the constant-dip case with a steeply-dipping slab (Fig. 11a), and

as large as ~ 680 km in the case with rollback (Fig. 11c). We estimate uncertainties in these calculations to be ± 20 km.

These results demonstrate that material path lines between the top of the subducting slab and surface volcanic centers are not directly vertical, and that trench-parallel offsets increase with greater subduction complexity, especially in the experiment with slab rollback. However, diapirs or fluids rising at rates much faster than the down-dip subduction rate (not shown) would experience smaller offsets (i.e. have much more vertical paths) than those seen in Fig. 11 (Kincaid *et al.* 2013). These findings have implications for geochemical studies that look for melt signatures directly above (e.g. Smith and Price, 2006) or directly trenchward (e.g. Green *et al.* 1999; Leeman *et al.* 2005; Pearce *et al.* 2005) of the hypothesized melting source on the subducting slab; these melt signatures may actually be found much closer to the slab centerline or on the shallowly-dipping slab (if one is present) than their origin points alone suggest, especially when slab rollback or an unusual slab geometry is present. The flow fields strongly vary with changing dip values and the imposition of rollback; the response of vertically rising material through these flow fields is similarly complex.

4.2 *Mantle anisotropy and shear-wave splitting*

To assess how the whisker azimuths observed in the Exps 16/20 and 17 might translate into potential shear-wave splitting in real subduction zones, shear-wave splitting values were calculated by propagating synthetic local S and SK(K)S waves through an

anisotropic model based on the observed whisker orientations. Although development of real olivine LPO and anisotropy in the mantle will no doubt differ in some respects from the simple whisker rotations (for example, the effects of recrystallization), these calculations provide a first order estimate of the shear-wave splitting implied by the experiments in this study.

Our model of anisotropy is divided into three layers in depth, each with a thickness of 125 km (including both the width of the light sheet and the gap above it to the next light sheet or to the surface of the fluid). Horizontal variation of anisotropy is specified by the whisker orientations of each block that falls within a given model layer (e.g. Figs 9 and 10). Elastic constants are specified by a blend of 70% olivine (the averaged forsterite elastic constants of Anderson and Isaak (1995) and Abramson *et al.* (1997)) and 30% orthopyroxene (the bronzite elastic constants of Frisillo and Barsch (1972)). For each whisker in a block, these constants are rotated so that the olivine *a*-axis orientation is parallel to the whisker azimuth; aggregate block coefficients are the result of averaging over all of the rotated coefficients (i.e. whisker orientations) within the block. Whisker alignments observed on or near the shallow slab (z3 level) are sub-parallel to slab dip; we vertically rotate the averaged elastic constants for these whisker azimuths to match the 30° dip of the shallowly-dipping slab. Elastic constants in the z2 and z1 layers are left horizontal; the whiskers at these depths do not show strong enough down-dip alignment to warrant a vertical rotation.

Once the elastic constants for a model block are calculated, we determine the ray path through each model block. To model splitting for local S waves from intermediate

depth earthquakes, we propagate synthetic local S waves through only the upper two depth slices (from 250 km to 0 km depth) (Fig. 10). Assuming an initial incidence angle of 30° in the z2 layer, we calculate the incidence angle in the z1 layer that would conserve phase ray parameter, assuming that the phase propagates at the average of the fast and slow shear wave velocities. Ray paths are determined in a similar manner for the synthetic SK(K)S waves, except that all three layers are included (Fig. 10), and the initial incidence angle of the SK(K)S wave in the z3 layer is 15° . Local S and SK(K)S ray paths are calculated over a range of backazimuths from 0° to 180° (with respect to the orientation of the trench).

Splitting is predicted for local S and SK(K)S waves with periods of roughly 2 seconds and 10 seconds, respectively. As SKS and SKKS waves emerge from the core as purely SV motion, the initial 10-second Gaussian wavelet for these phases is polarized in the plane of the phase path. The initial 2-second Gaussian wavelets for the local S phases are assigned equal SV and SH motion. For the incidence angle and backazimuth in each layer, the Christoffel equation is used to determine the particle motion and phase velocities of the fast and slow shear waves. For each layer, the wavelet is rotated into the fast and slow shear wave polarizations, the fast and slow polarizations are offset by the shear-wave splitting time for the layer, and the phase is then rotated back into its original coordinate system. Using the particle motions calculated at the surface, which integrate the effects of anisotropy in all layers, shear-wave splitting parameters (fast direction, ϕ , and splitting time, dt) are calculated using the eigenvalue minimization of Silver and Chan (1991).

The strength of anisotropy is adjusted to match the estimate of splitting time per kilometer of mantle wedge (2.2×10^{-3} s/km) calculated for the South American subduction zone by MacDougall *et al.* (2012). Anisotropic strength is reduced from single crystal values by applying a uniform dilution factor (i.e. the anisotropy percentage) to the elastic coefficients for each whisker. This assumption produces backazimuth-averaged splitting times of 0.66-0.67 seconds for the SK(K)S waves and 0.38-0.39 seconds for the local S waves. Averaged predicted shear-wave splitting parameters for local S and SK(K)S phases in model blocks 10 and 14 for Exps 16/20 and 17 are given in Fig. 10. The predicted splitting values, as a function of backazimuth, are in Supplementary Table 1.

Predicted shear-wave splitting fast polarizations for model blocks 10 and 14 in Exps 16/20 and 17 yield averages that range from mildly (-10°) to moderately (-30°) deflected from trench-normal, respectively. In the back-arc (block 10) of both experiments, whisker orientation rose diagrams show little variation in depth (Fig. 10); averaged predicted local S and SK(K)S shear-wave splitting fast polarizations are close to the mean whisker orientations observed in each model layer and range from -10° to -15° (Fig. 10). At the slab gap in the case without rollback (block 14, Exp. 17), however, trench-parallel flow near the slab gap causes average whisker azimuths to be rotated counter-clockwise in depth (Figs 9 and 10); averaged predicted local S and SK(K)S fast directions for this case are -29° and -33° , respectively. In comparison, rollback counteracts much of the trench-parallel flow produced by the slab gap (block 14, Exp. 16/20), yielding more trench-normal averaged fast polarizations of -12° and -15° .

While some portions of the mantle wedge near the slab gap manifest trench-parallel flow and whisker alignments indicative of the ‘trench-parallel stretching’ modeled by Kneller and van Keken (2007), none of these flow fields produce trench-parallel whisker alignments over a large enough depth range to generate trench-parallel shear-wave splitting fast polarizations. Thus the flow patterns in these experiments do not explain the trench-parallel fast polarizations observed in many subduction zones, including those from local S phases representing anisotropy within the mantle wedge (e.g. Yang *et al.* 1995; Fouch and Fischer, 1996; Audoine *et al.* 2000; Polet *et al.* 2000; Anderson *et al.* 2005; Long and van der Hilst, 2006; Long and Silver, 2008; Abt *et al.* 2009; MacDougall *et al.* 2012; Long, 2013). This result raises several possibilities: 1) other geodynamic factors may be at work (resulting in different shear patterns), 2) factors other than those represented by whiskers control olivine a-axis orientations (i.e. dynamic recrystallization, non A-type olivine, etc.) or 3) other mineralogies or sources of anisotropy (e.g. melt) strongly influence the anisotropy observed in subduction zones.

Comprehensively considering all alternative possibilities to explain trench-parallel shear-wave splitting fast polarizations goes well beyond the scope of this study, but a few may be relatively easily evaluated, including anisotropy in the subducting slab and the presence of B-type olivine fabric in the mantle wedge. While anisotropy within the subducting slab may be present and can produce significant shear-wave splitting (e.g. Faccenda *et al.* 2008; Healy *et al.* 2009; Hammond *et al.* 2010), these effects will be the largest at fore-arc stations, where phase path lengths through the slab can be longer than

those that sample the mantle wedge. With B-type olivine fabrics, the olivine *a*-axis aligns perpendicular to the flow direction; trench-normal flow in the mantle wedge would produce trench-parallel shear-wave splitting observations at the surface (e.g. Jung and Karato, 2001). However, B-type olivine is generated in water-rich, high stress, low temperature environments (Jung and Karato, 2001; Jung *et al.* 2006; Katayama and Karato, 2006); numerical modeling indicates that these conditions only occur in the fore-arc of the mantle wedge (Kneller *et al.* 2005; Kneller *et al.* 2007). Thus neither slab anisotropy nor B-type fabrics provide sufficient explanations for the abundant trench-parallel fast directions from local S paths that sample the hotter regions of the mantle wedge beneath arc and back-arc regions of many subduction zones.

The influence of slab rollback is a key issue in the interpretation of seismic anisotropy in subduction zones. Slab rollback has long been invoked to explain observed trench-parallel fast polarizations from SK(K)S phases that sample anisotropy below the slab (e.g. Russo and Silver, 1994), and experimental and numerical studies have shown that rollback has the potential to produce a zone of trench-parallel olivine *a*-axes below the slab (e.g. Buttles and Olson, 1998; Kincaid and Griffiths, 2003; Kincaid and Griffiths, 2004; Schellart 2004; Funicello *et al.* 2006; Jadamec and Billen, 2010; Jadamec and Billen, 2012; Faccenda and Capitanio, 2013). Rollback has also been thought to drive flow that would create olivine alignments consistent with trench-parallel splitting fast polarizations within the mantle wedge (e.g. Long and Silver, 2008; Abt *et al.* 2009). However, recent experimental and numerical models (including this one) have shown that rollback will instead tend to produce trench-normal olivine *a*-axes above central regions

of the slab away from toroidal flow around slab edges (Druken *et al.* 2011; Faccenda and Capitanio, 2012; Faccenda and Capitanio, 2013). We observe that rollback modifies flow patterns so that trench-normal flow at least partially erases trench-parallel flow near the slab gap. However, whisker alignments in our experiment with rollback are also not purely trench-normal, indicating that the interaction of slab gaps and rollback can produce a range of olivine *a*-axis orientations that vary on small length scales, as has been observed in some subduction zone segments (e.g. MacDougall *et al.* 2012).

5. Conclusions

Our laboratory experiments exhibit time varying, 3-D circulation patterns in response to different styles of subduction with complex plate geometries. In our reference cases, which have a constant slab dip value, flow in the mantle wedge is roughly symmetric about the slab centerline. Marker speeds increase with larger dip angles, as does the two-dimensionality of the path lines. 3-D flow is more obvious at shallower dip angles, where path lines have increasingly trench-oblique to trench-parallel orientations.

Slab dip variations generate even greater 3-D flow in the wedge; trench-parallel flow perturbs path lines from above the steeper slab towards the fore-arc above the shallower slab. This flow is consistent with analytical (Hall *et al.* 2000) and numerical (Kneller and van Keken, 2007; Kneller and van Keken, 2008) models that indicate the same sense of flow produced by reduced pressures above the shallowly-dipping slab. The rate of trench-parallel flow increases with the magnitude of the trench-parallel slab dip

variation, and, for a fixed slab dip difference, as the dip of the shallowly-dipping slab segment decreases. Total flow rates also increase with slab dip difference.

Rollback enhances flow rates in all regions of the wedge, particularly in the region above the steeper plate. Rollback in the presence of a slab gap also tends to enhance trench-parallel flow across the slab centerline, and results in stronger penetration of sub-slab mantle into the wedge. In our models, this through-flow is largely horizontal, and is therefore not expected to passively or actively feed into decompression melting.

These results also allow us to draw conclusions about arc melt paths and mantle wedge seismic anisotropy. Path line asymmetry in cases with complex slab morphology produces significant spatial offsets between the deep source location and the near-surface delivery of arc melts. Despite causing trench-parallel or trench-oblique path line deflections, near the slab gap rollback produces trench-normal whiskers alignments that counter the tendency of the slab dip change to drive whisker rotation towards the shallowly-dipping slab. The production of trench-parallel flow due to subduction of a slab gap causes whisker rotations near the slab gap of up to 30° away from trench-normal. Rollback drives whisker rotation near the slab gap into orientations that are only 10° away from trench-normal.

These experiments show that a slab gap can generate significant perturbations in wedge flow and both seismic and geochemical signals recorded at the surface in convergent margins. However, even in cases that maximize the amount of trench-parallel

wedge flow (no rollback, large slab gap with one very shallowly-dipping slab), whiskers in the central mantle wedge never become strongly aligned in the trench-parallel direction. While slab anisotropy and B-type olivine fabric may contribute to trench-parallel shear-wave splitting observations in the fore-arc, other mechanisms are still needed to explain the extensive zones of trench-parallel shear-wave splitting fast polarizations observed in the arc and back-arc regions of many subduction zones.

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8. Appendices

Appendix A: Additional analyses of u^ , $u_{\perp}^*$, $u_{\parallel}^*$, and A_{norm}*

These figures show u^* , $u_{\perp}^*$, $u_{\parallel}^*$, and A_{norm} as a function of normalized distance from the trench (s_{norm}) for selected path line plots from Exps 3, 4, 7, 10, and 16. Fig. A1 is the expanded version of Figs 2c and 2d, Fig. A2 is the expanded version of Figs 4a and b, and Fig. A5 is the expanded version of Figs 6a and 6b. A summary of the maximum u^* values for the $z1$ level of each experiment is shown in Fig. A4.

Appendix B: 2-D corner flow solutions

Using the methods described in Hall *et al.*, (2000) and Turcotte and Schubert (2002), we calculate 2-D pressure, pressure gradient, and velocity values for slab dips of 30° and 60° (Fig. A3), using the material properties of the glucose syrup. With a smaller slab dip, non-hydrostatic pressures in the mantle wedge are more negative (Fig. A3, top row), and non-hydrostatic pressure gradients are larger (Fig. A3, middle row). These conditions cause stronger and more rapid wedge flow towards the shallowly-dipping slab and away from the steeply-dipping slab.

Appendix C: Flow and volume flux estimates through the slab gap

A path similar to the blue path in Fig. 7 is highlighted in Fig. A6a, where a tracer starts out beneath the shallowly-dipping slab, moves into the plane of the figure and enters the wedge above the steeply-dipping slab, eventually subducting after 7 minutes of transport. Fig. A6b represents faster gap through-flow in the presence of slab rollback.

To obtain order-of-magnitude estimates for volume flux through a slab gap, we assume perfectly horizontal flow across a cross-section whose width equals the separation of the two belts measured in the trench-parallel direction, and whose triangular, cross-sectional shape is the one prescribed by the difference in dip between the slabs, down to the roller at 670 km ($D_L = 14$ cm; the area between the green line and the blue line in Figs A6a and A6b). These volume flux estimates are therefore maximum possible values.

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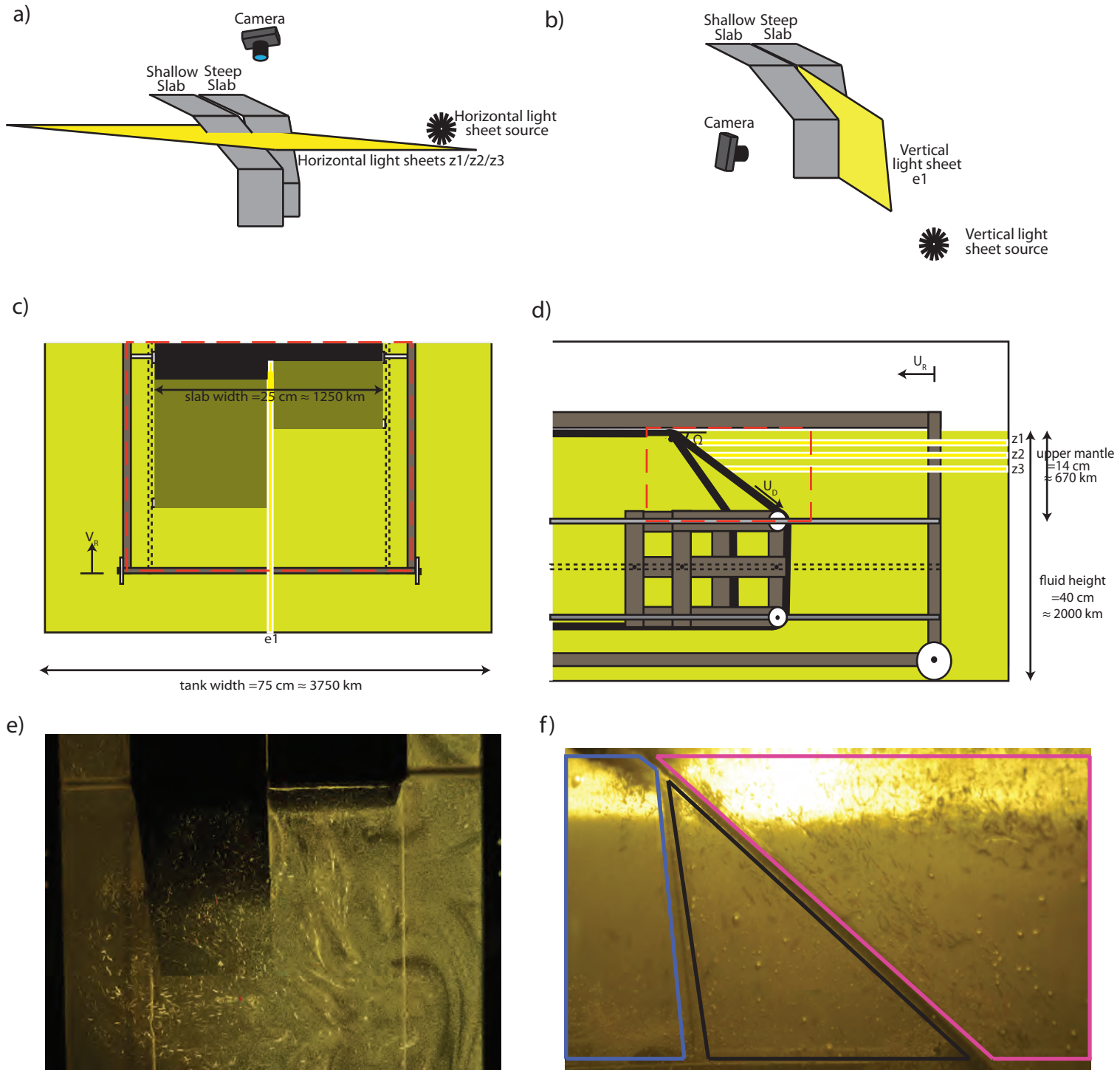


Figure 1. The two different types of light sheet used in our experiments. a) The horizontal light sheet (z_1 , z_2 , or z_3), where the entire mantle wedge is illuminated at a constant depth. b) The vertical light sheet (e_1), where a sliver of mantle wedge is illuminated at all depths. c-d) Diagram of apparatus in map view (c) and cross-section (d). U_d is the down-dip subduction rate, U_r is the rollback rate, Ω is the slab dip measured downward from the horizontal, z_1 , z_2 , and z_3 are horizontal light sheets, and e_1 is the vertical light sheet. Red dashed boxes outline the insets shown in Figs 1e and 1f. e-f) Actual images using the corresponding light sheets in (a-b) and the area inside the red dashed box in (c-d). f) is a side view of the slabs, with a horizontal light sheet used to illuminate the sub-slab mantle wedge beneath both slabs (blue shape), the material that is either above the steeply-dipping slab or below the shallowly-dipping slab (black shape), and the area in the mantle wedge that is above both dipping slabs (magenta shape).

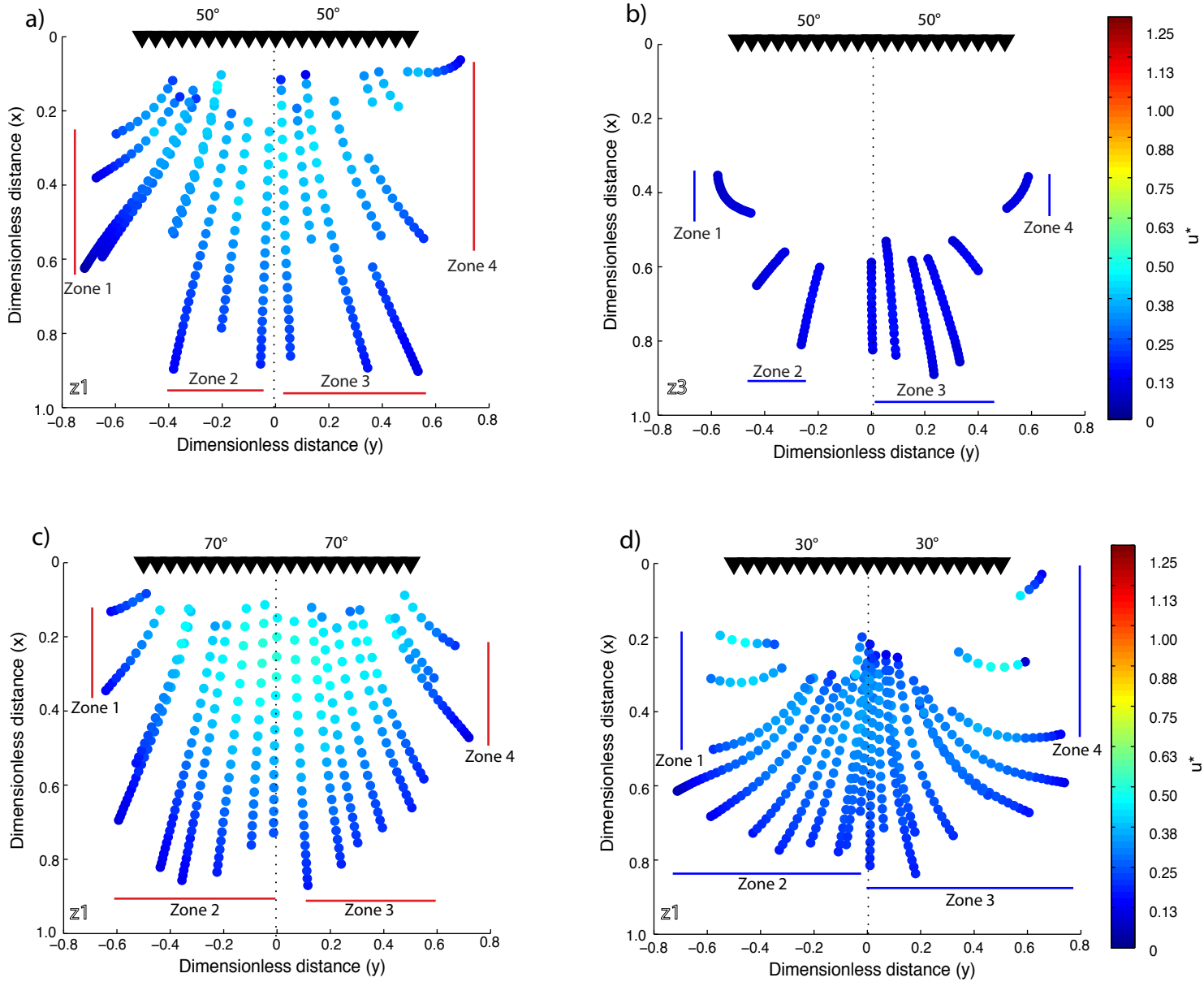


Figure 2. Comparison of Exp. 1 ($U_D = 8$ cm/min, dip = 50°) at depths z_1 and z_3 , and Exps 3 ($U_D = 8$ cm/min, dip = 70°) and 4 ($U_D = 8$ cm/min, dip = 30°) at depth z_1 . Map view plots at depths (a, c-d) z_1 (=2 cm $\approx$ 100 km) and (b) z_3 (=7 cm $\approx$ 350 km) are shown. Individual circles represent motion of a single marker through time. Colors represent dimensionless flow rate ($= u^*$). Black lines with triangles show the location of the trench. Slab segment dip values are displayed above their corresponding halves of the trench. The dotted line represents the slab centerline. Locations are given as distance in trench-parallel (y) and trench-normal (x) directions, normalized by total plate width ($W = 25$ cm) relative to an origin along the slab centerline at the trench. a) Markers moving within the shallow wedge at z_1 (depth of 2 cm $\approx$ 100 km) show four distinct zones of flow and resupply to the slab surface (red lines): **1.** Material originating in the mantle wedge at the outer edge of the left-hand slab. **2.** Material originating in the mantle wedge between $-0.4 < y < 0$. **3.** Material originating in the mantle wedge between $0 < y < 0.55$. **4.** Material originating in the mantle wedge above the outer edge of the right-hand slab. b) Similar plot to (a), but at z_3 (depth of 7 cm $\approx$ 350 km). c) Path line plot of Exp. 3 at depth z_1 . d) Path line plot of Exp. 4 at depth z_1 .

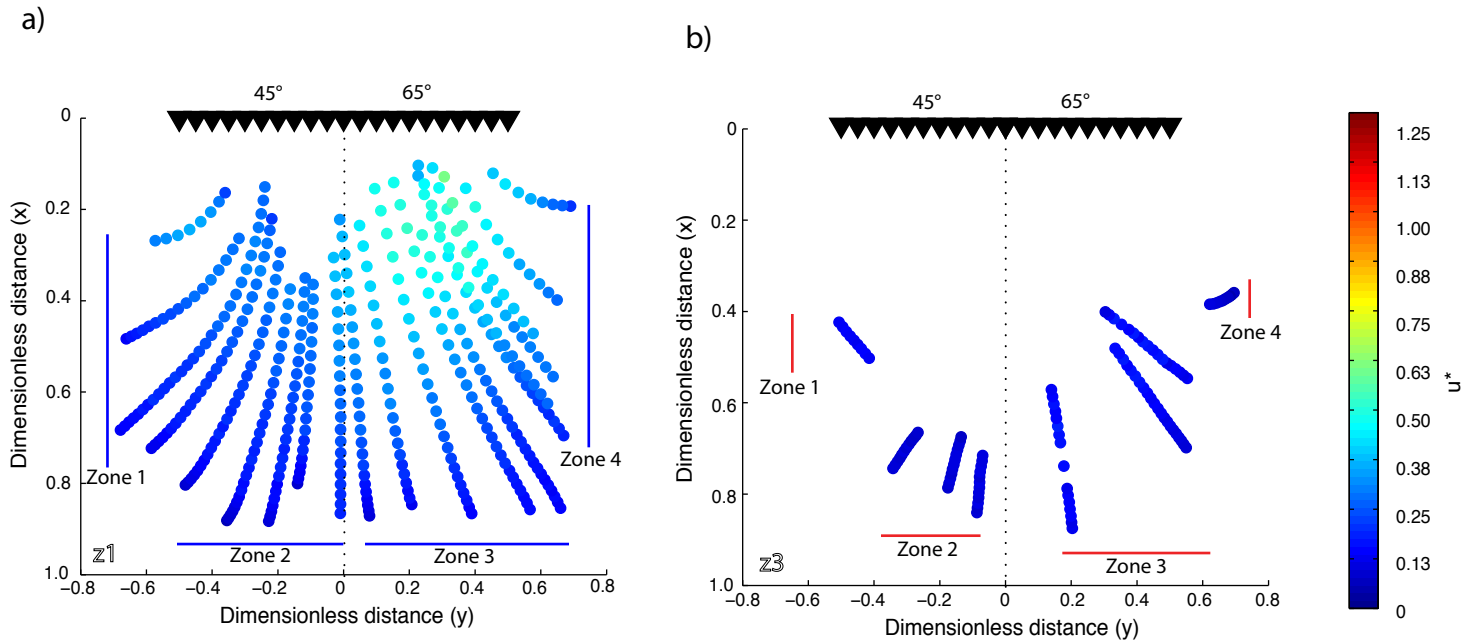


Figure 3. Comparison of Exp. 8 ($U_D = 8$ cm/min) at depths z_1 and z_3 . Exp. 8 has a steep dip of 65°, a shallow dip of 45°, and is plotted at depths a) z_1 ($= 2$ cm ≈ 100 km) and b) z_3 ($= 7$ cm ≈ 350 km). Individual circles represent motion of a single marker through time. Colors represent dimensionless marker speed. Black lines with triangles show the location of the trench. The black dotted line represents the slab centerline. Slab segment dip values are shown above their corresponding halves of the trench. Locations are given as distance in trench-parallel (y) and trench-normal (x) directions, normalized by total plate width ($W = 25$ cm) relative to an origin along the slab centerline at the slab surface. The discontinuous path line in Zone 3 of Fig. 3b represents a whisker that was just on the boundary of the z_3 light sheet; it would sometimes sink below and then re-enter the z_3 layer.

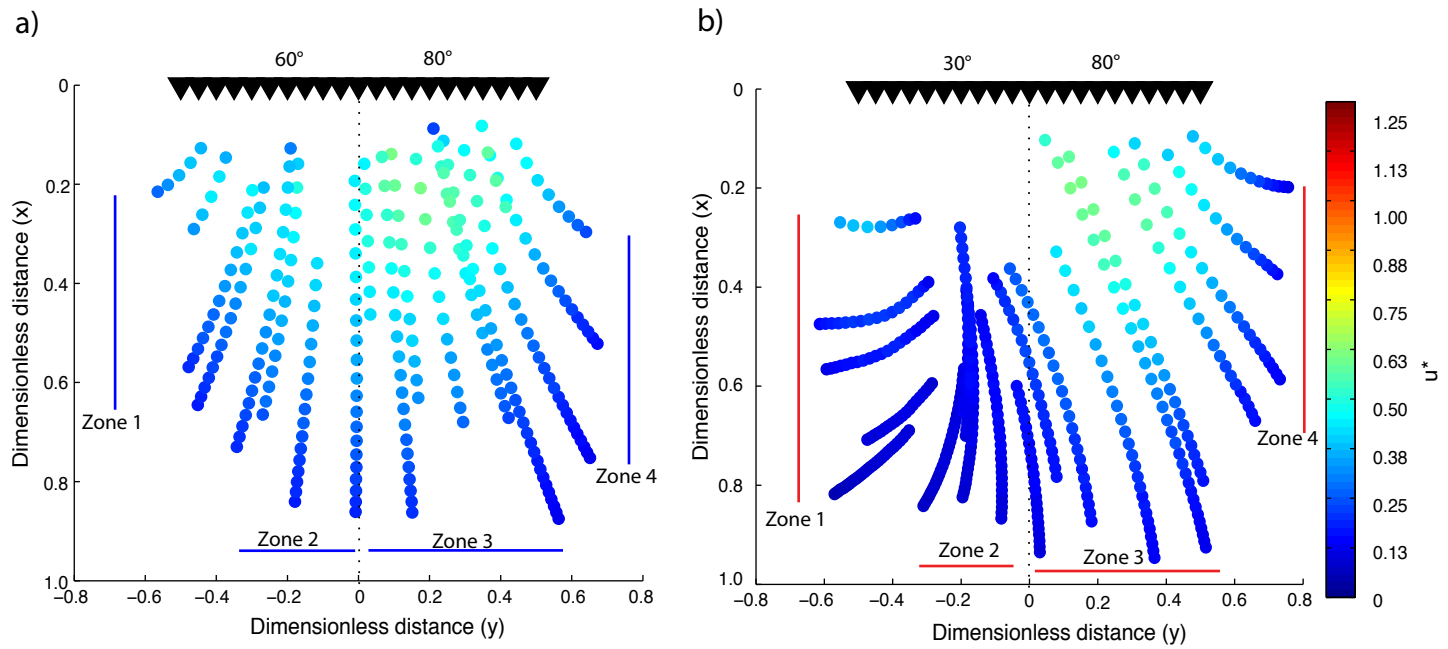


Figure 4. Comparison of Exps 10 ($U_D = 8$ cm/min) and 7 ($U_D = 8$ cm/min) at depth z_1 . Path line plots for (a) smallest gap width at the steepest possible dips (60°/80°, Exp. 10) and (b) the largest gap area (30°/80°, Exp. 7) are shown. Black lines with triangles show the location of the trench. The black dotted line represents the slab centerline. Slab segment dip values are displayed above their corresponding halves of the trench.

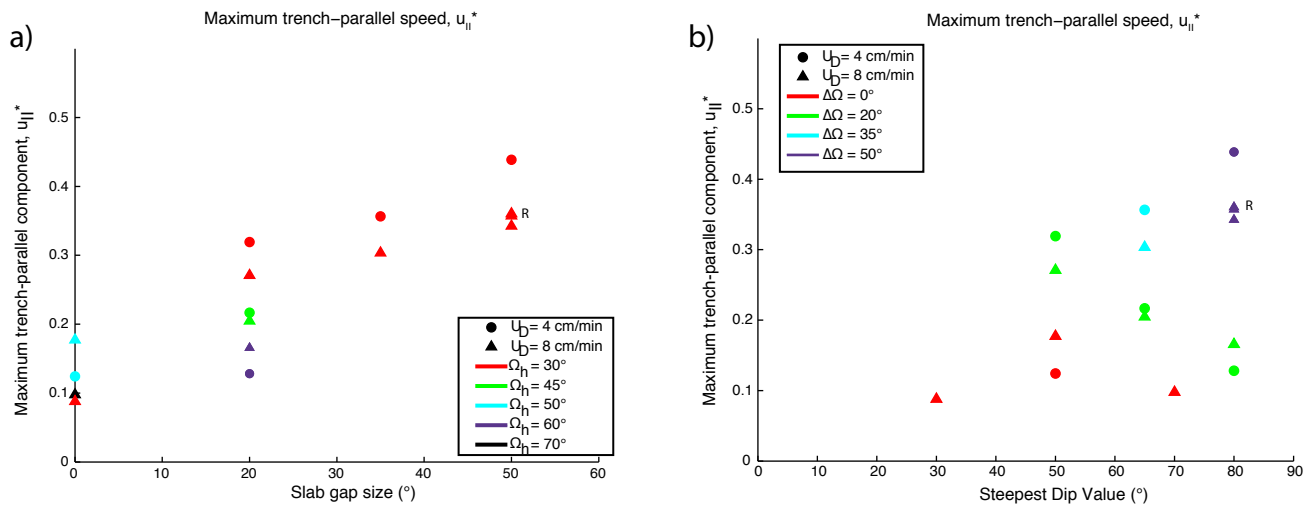


Figure 5. Summary of maximum trench-parallel speeds ($u_{\parallel}^*$), plotted according to both slab gap size ($\Delta\Omega$) and steepest dip value. a) Maximum trench-parallel speed $u_{\parallel}^*$ in each experiment, calculated over each path that had points within a box with width $W/2$ ($= 12.5$ cm) along the trench, centered on the slab centerline, and extending $W/4$ ($= 6.25$ cm) from the trench into the mantle wedge. With rollback, we change the box size to extend $W/2$ ($= 12.5$ cm) from the trench into the mantle wedge to account for the greater spatial extent inherent in the rollback case. $\Delta\Omega$ represents the slab gap size, and Ω_h represents the shallowest dipping slab in a given experiment. The 'R' in both (a) and (b) represents the values for Exp. 16, which is the only experiment where we added slab rollback. (b) represents the same data as in parts (a); we have changed the x-axis in this diagram to illustrate that the gap size, rather than the steepest dip value, drives the increase in $u_{\parallel}^*$.

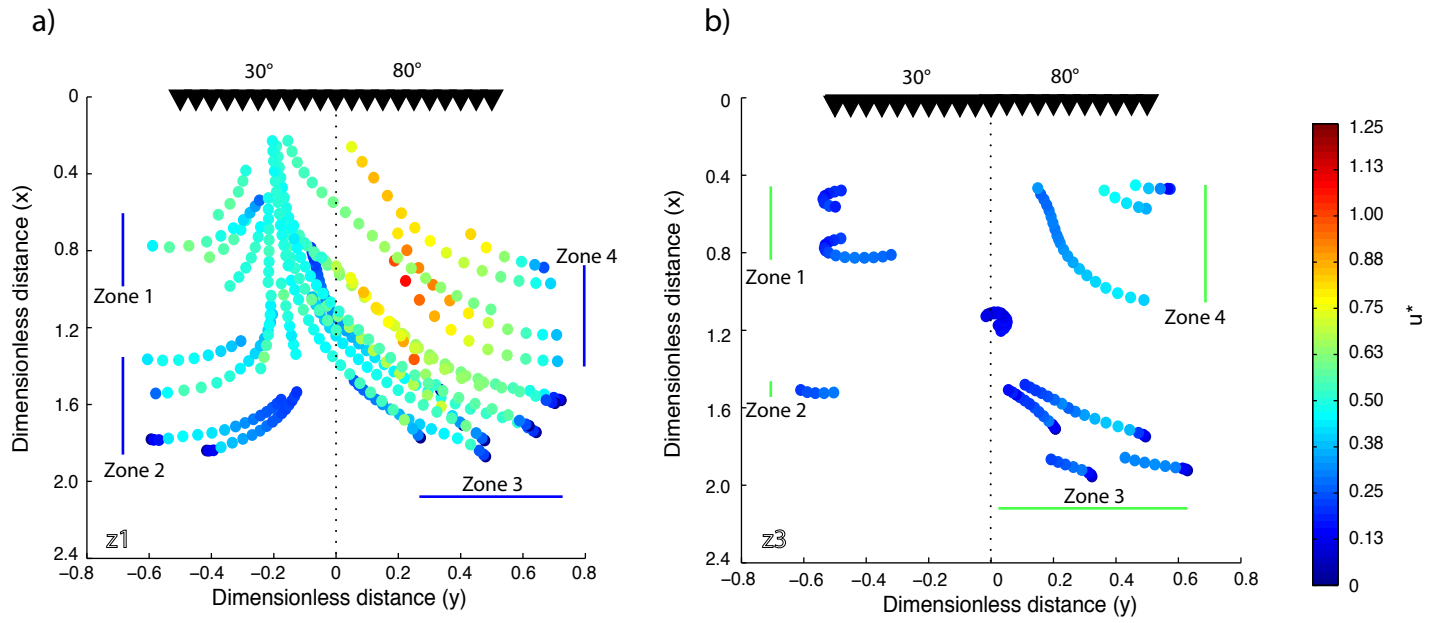


Figure 6. Comparison of Exp. 16 ($U_D = 8$ cm/min, $U_R = 3$ cm/min) at depths z1 and z3. The plots for an experiment with both the maximum slab gap size and rollback (Exp. 16) at depths of a) z1 (= 2 cm ≈ 100 km) and b) z3 (= 7 cm ≈ 350 km) are shown. Black lines with triangles show the location of the trench. The dotted line represents the slab centerline. Slab segment dip values are shown above their corresponding halves of the trench.

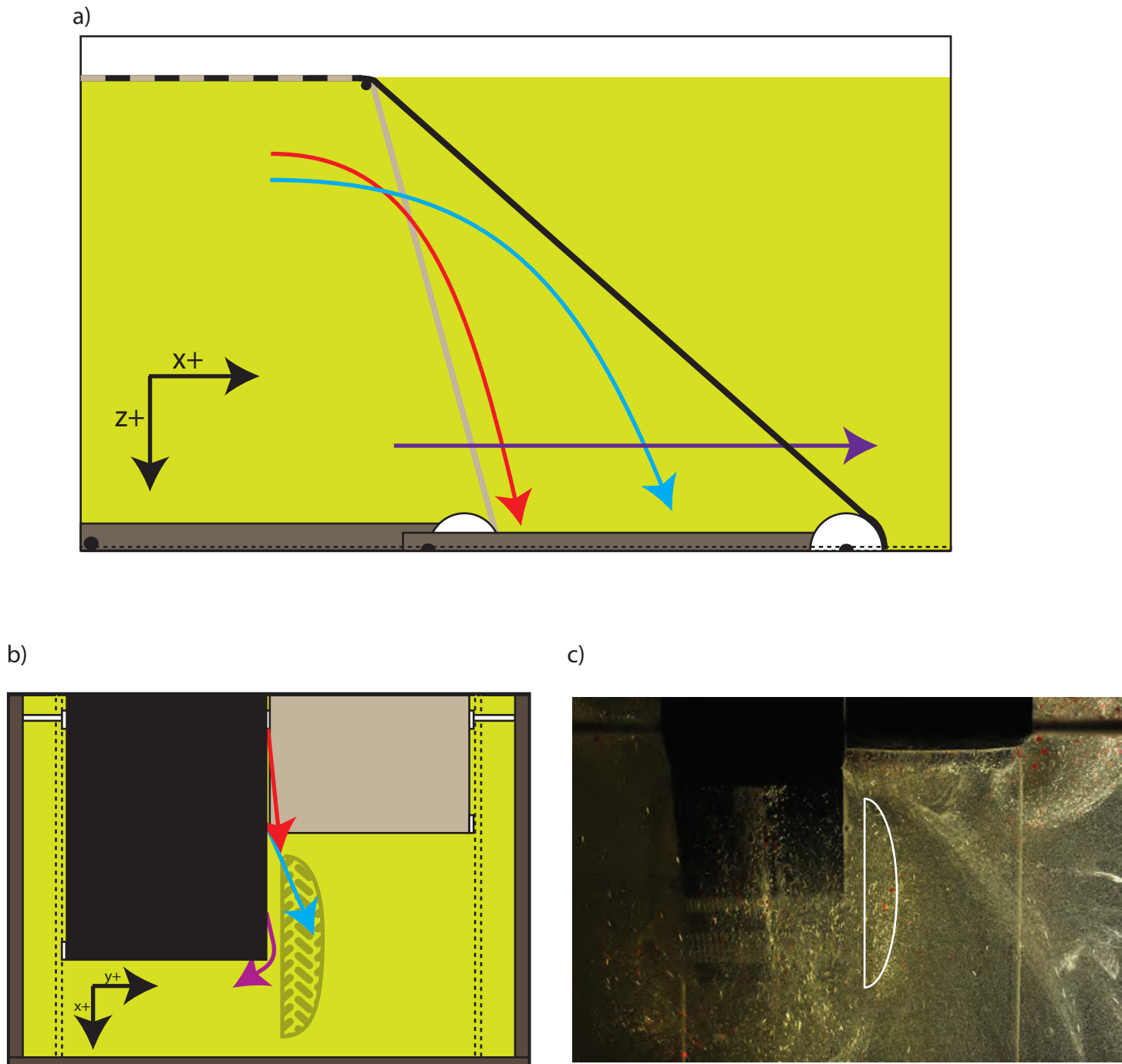


Figure 7. Schematic diagrams of observed paths of sub-slab mantle flowing through the gap and into the mantle wedge in a) cross-section and b) map view. The zone in the mantle wedge where sub-slab material, following the trajectory through the slab gap indicated by the blue line, can reside before being subducted is denoted by the shaded shape. c) Photo from an experiment with the maximum slab gap area and subduction rate (Exp. 17), where the shaded zone in (b) is outlined in white.

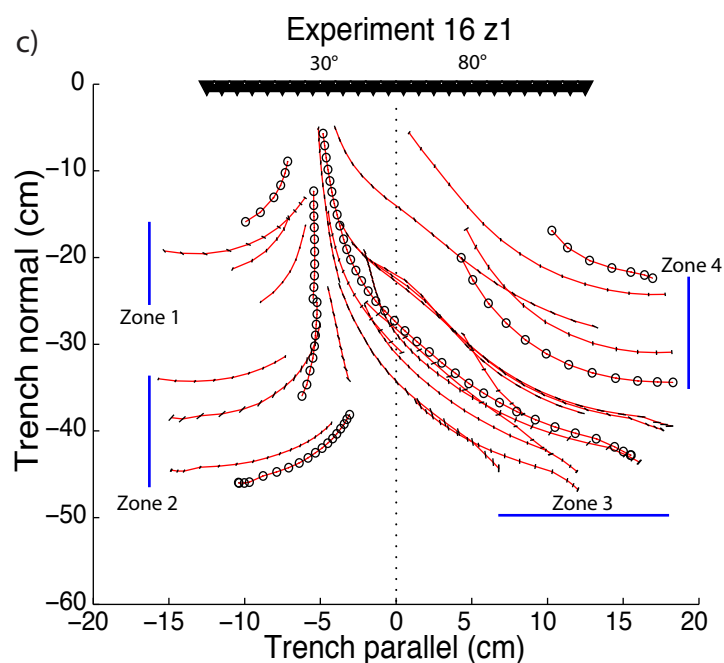
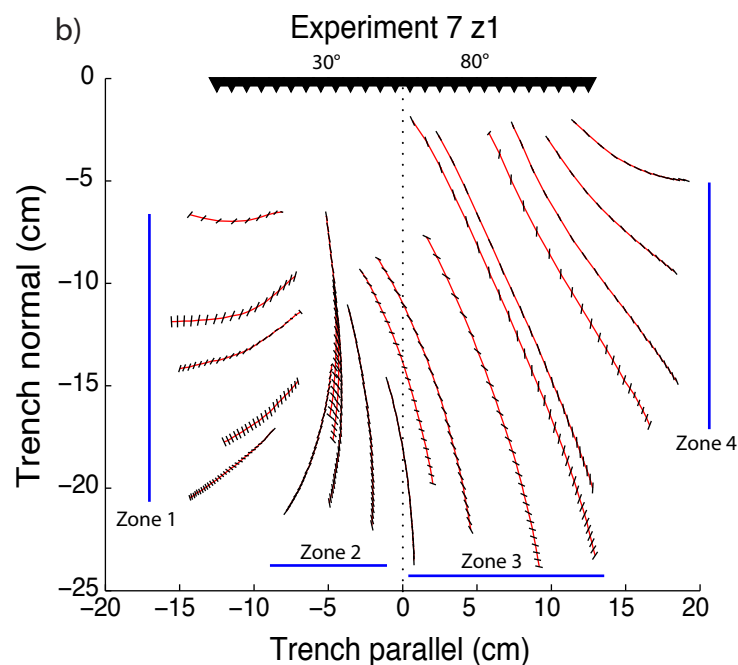
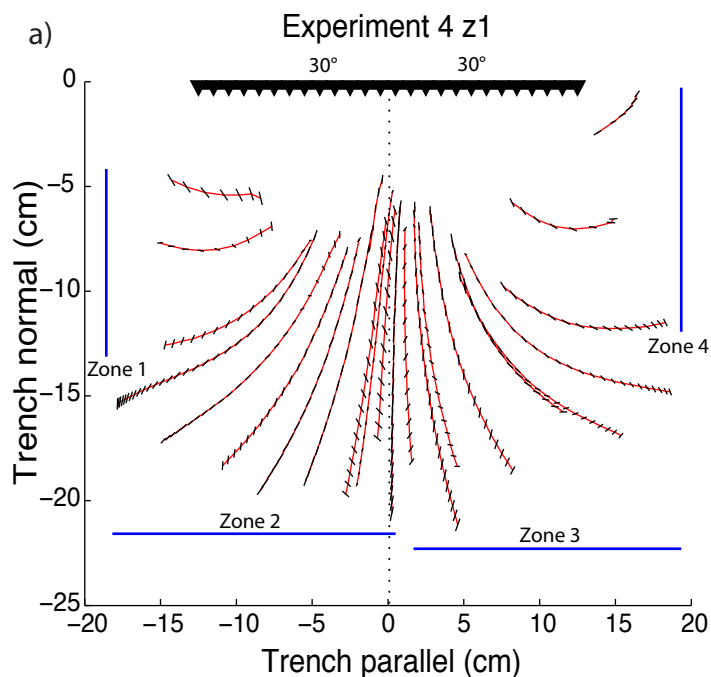


Figure 8. Path lines and whisker orientations for (a) Experiment 4, (b) Experiment 7, and (c) Experiment 16. Red lines represent the path line vector orientation from one time step to the next, and the black lines represent whisker orientations at a time step. The black dotted line represents the slab centerline. Zones from previous figures are denoted by the blue lines. Black circles indicate a bubble or bead, rather than a whisker, and thus provide us with no information about the rotation along a path line. Slab segment dip values are displayed above their corresponding halves of the trench.

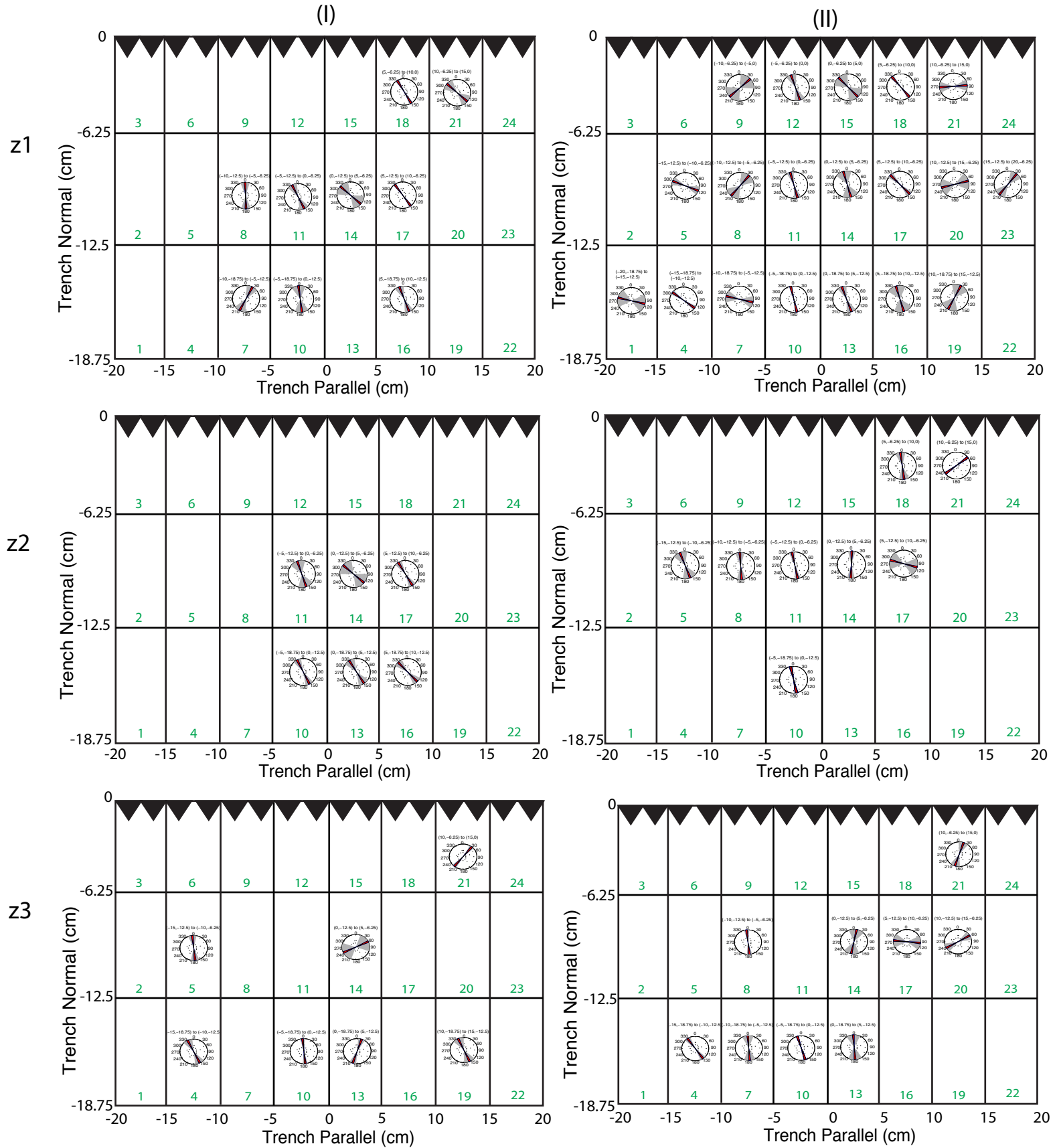


Figure 9. Rose diagrams of whisker orientations within a given model block for (I) an experiment with the maximum slab gap size and maximum down-dip subduction rate (Exp. 17) and (II) an experiment with the maximum slab gap size, maximum down-dip subduction rate, and a rollback rate of 3 cm/min (Exp. 16/20). The three light sheets are designated by z1, z2, and z3. Red lines indicate the average whisker azimuth for that model block, and the grey zone surrounding the red indicates the standard deviation of the whisker orientations. The numbers in green are the model block numbers referred to in the text.

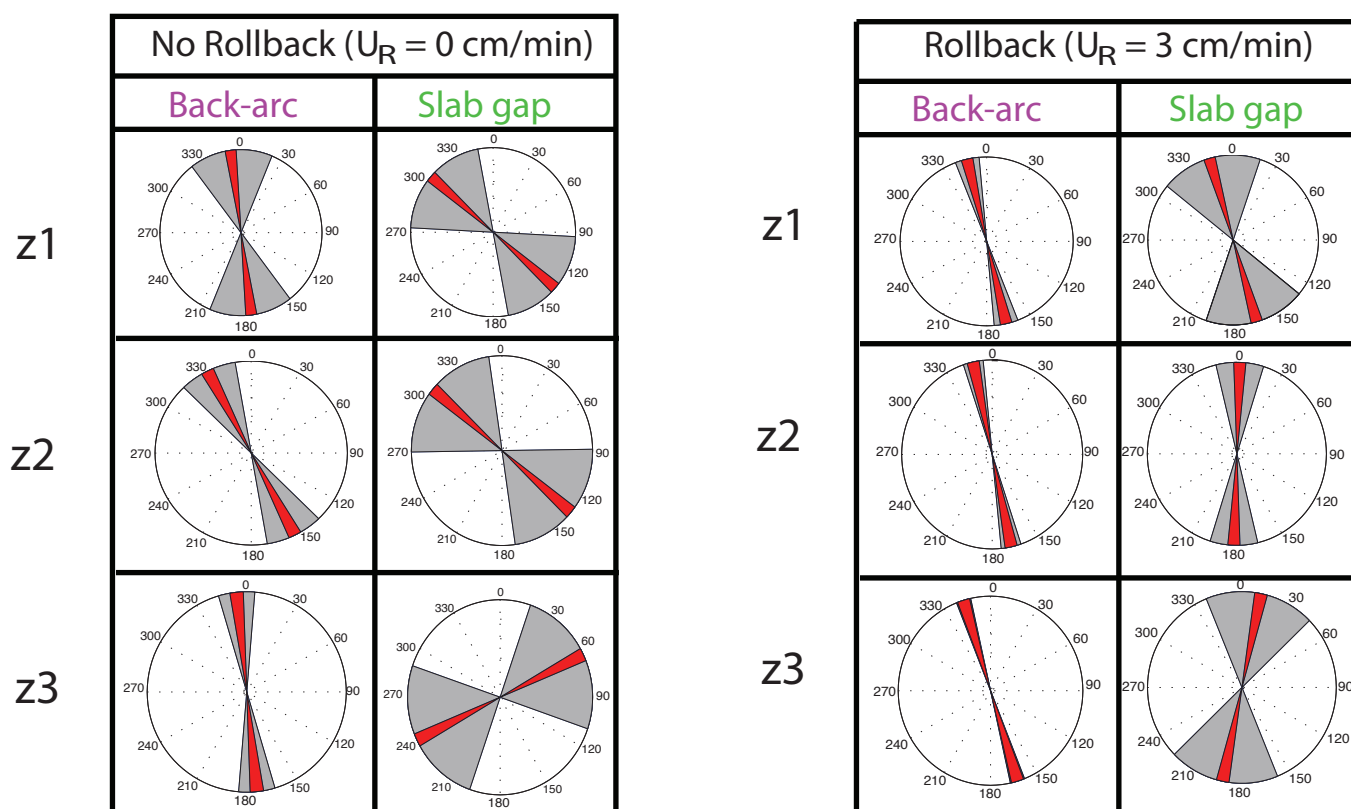
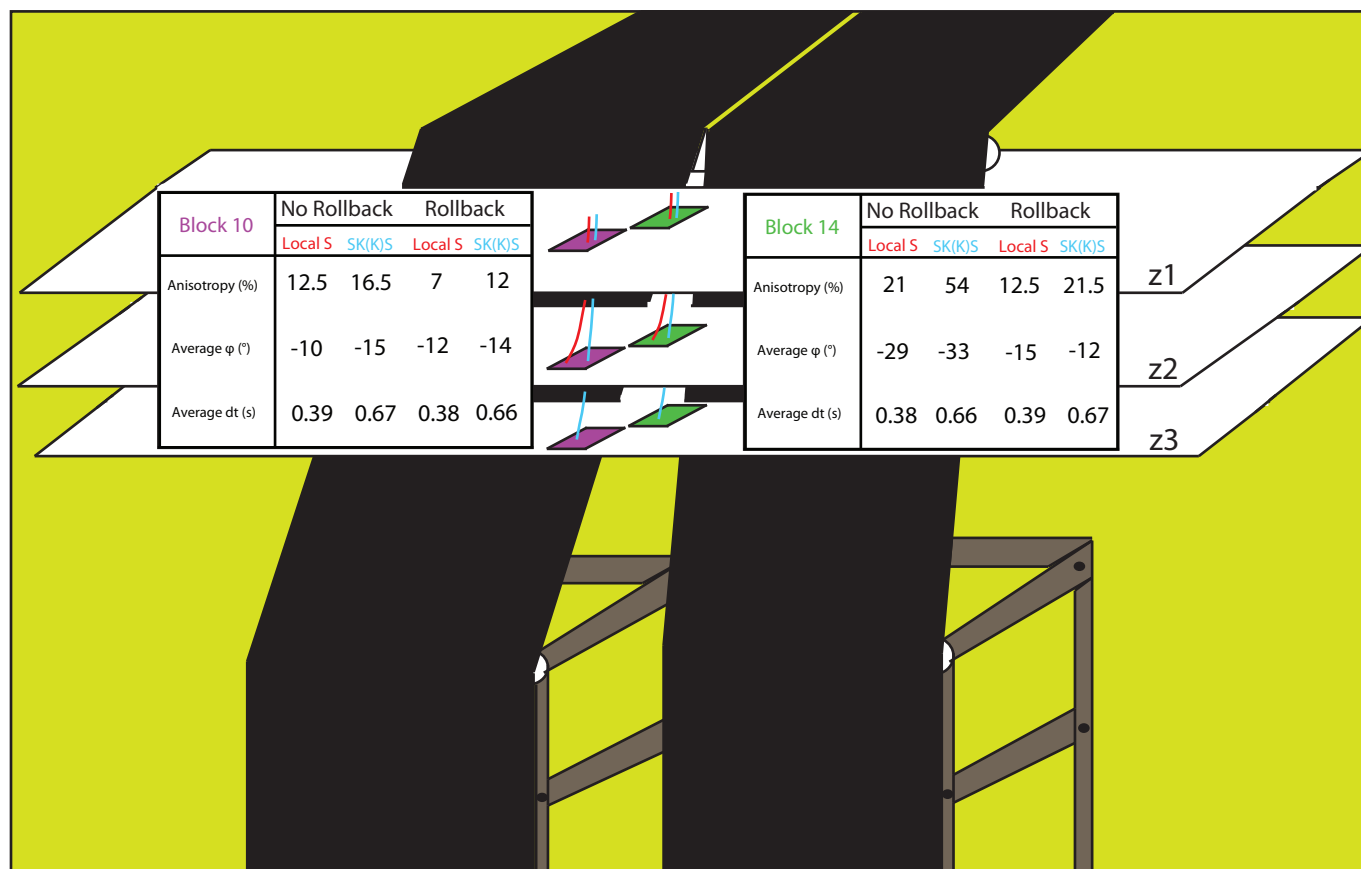


Figure 10. Whisker azimuth rose diagram comparison between two experiments with the same maximum slab gap size and maximum down-dip subduction rate; the only difference is the lack (Exp. 17 [“No Rollback”]) or presence (Exp. 16/20 [“Rollback”]) of rollback. We compare a model block located near the slab gap over the steeply-dipping slab (block 14, at the slab gap, green) and a block further in the back-arc over the shallowly-dipping slab (block 10, in the back-arc, purple). Averaged shear-wave splitting fast directions (ϕ) and splitting times (dt) are included for blocks 10 and 14 for local S (red ray paths) and SK(K)S (blue ray paths) phases. Anisotropy percentages are those that generate the average dt values that most closely match the benchmark values discussed in the text. Original synthetic results are displayed in Supplementary Table 1.

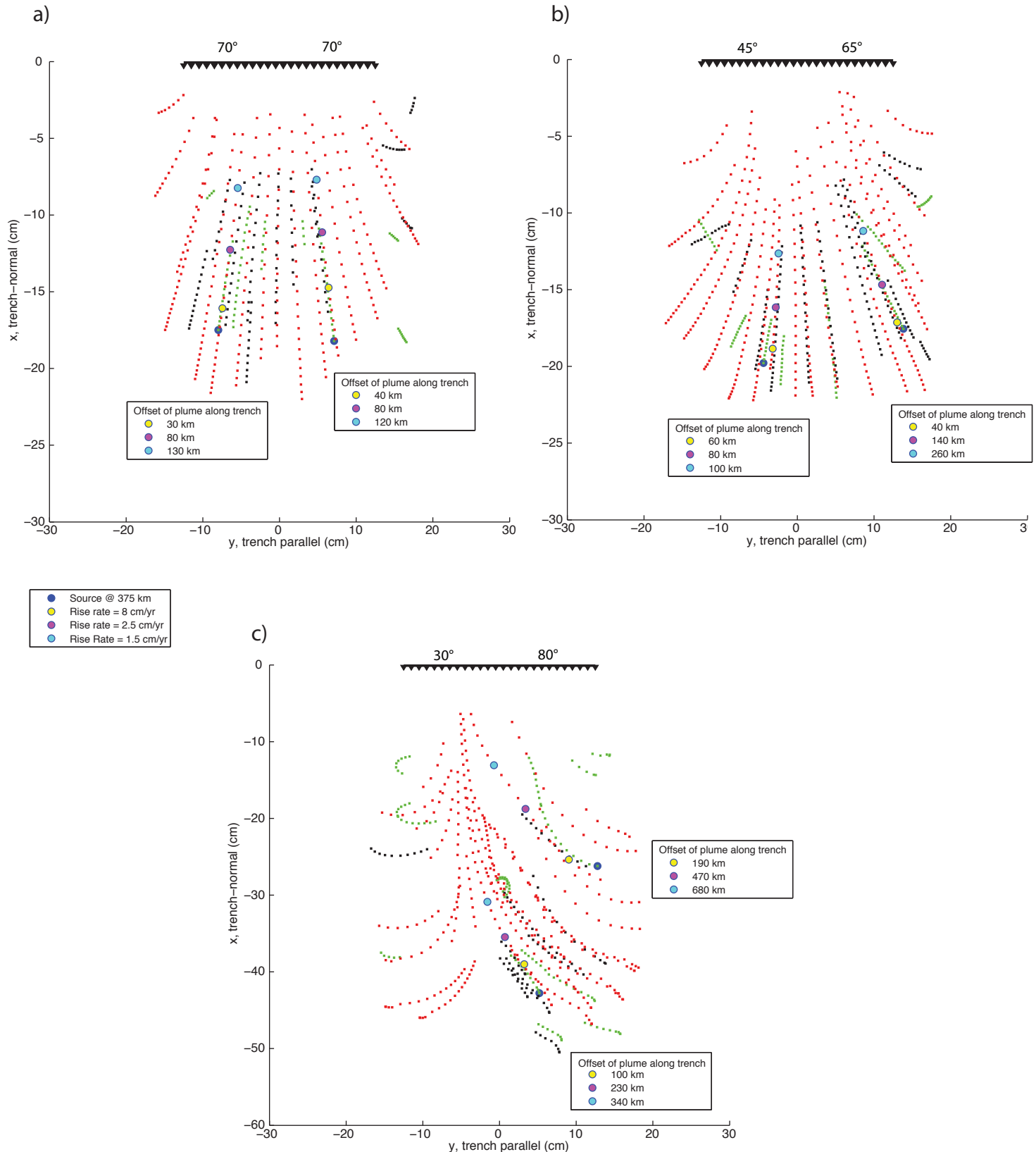


Figure 11. Estimates of diapir locations for a) a constant, steep-dip slab (Exp. 3), b) a small slab gap size with intermediate slab dips (Exp. 8), and c) the maximum slab gap size and rollback (Exp. 16). The legend beneath (a) holds for all three figures. Based on the different rise rates, the diapirs are propagated along (close to) overlapping marker paths first in z3 (green points), then in z2 (black points), then in z1 (red points). The trench-parallel distance (i.e. in y) is calculated and translated into the real Earth system by our conversion of 1 cm in the lab equals 50 km in the Earth reference frame. Slab segment dip values are shown above their corresponding halves of the trench.

Exp	Ω_N (°)	Ω_S (°)	U_D (cm/min)	U_R (cm/min)	$u(x,t)$ Max (cm/min)	u^* Max (unitless)	$u_{ }^*$ Max (unitless)	β (cm ³ /min)
1	50	50	8	0	3.6128	0.4516	0.1772	--
2	50	50	4	0	1.7648	0.4412	0.1241	--
3	70	70	8	0	4.1242	0.5155	0.0978	--
4	30	30	8	0	3.1181*	0.3898*	0.0879*	--
5	50	30	8	0	4.0568	0.5071	0.2710	--
6	65	30	8	0	4.4444	0.5555	0.3036	--
7	80	30	8	0	4.9093	0.6137	0.3604	0.7-8.1
8	65	45	8	0	4.8905	0.6113	0.2048	5.3-9.4
9	65	45	4	0	2.1237	0.5309	0.2168	--
10	80	60	8	0	4.9021	0.6128	0.1657	--
11	80	60	4	0	2.5153	0.6288	0.1282	--
12	80	30	4	0	3.1848	0.7962	0.4388	0.7-4.9
14	50	30	4	0	2.279	0.5698	0.3193	--
15	65	30	4	0	2.5592	0.6398	0.3566	0.7-2.6
16, 20	80	30	8	3	8.6228	1.0779	0.3574	4.9-12.0
17	80	30	8	0	5.6264	0.7033	0.3426	--

Table 1. The subduction parameters and analysis results for each experiment. Ω_N is the dip of the belt on the right side in Fig. 1c, Ω_S is the dip of the belt on the left, U_D is the down-dip subduction velocity, U_R is the rollback rate, $u(x,t)$ Max is the maximum horizontal velocity, u^* is the maximum unitless marker speed (maximum marker velocity $u(x,t)$ Max normalized by U_D), $u_{||}^*$ Max is the maximum trench-parallel component of marker speed (u^*) in each experiment in the area covering the middle of the left-hand belt to the middle of the right-hand belt ($= 12.5$ cm = total belt width/2), and extending 6.25 cm (=total belt width/4) away from the trench and into the mantle wedge. The only exception for this boundary box is for Exp. 16, where, due to the rollback, we extend the box to be 12.5 cm away from the trench and into the mantle wedge. The asterisks next to the $u(x,t)$ Max value, u^* Max, and $u_{||}^*$ Max values for Exp. 4 represent the fact that faster velocities were observed in this experiment than the value shown, but those higher values were from markers experiencing pure toroidal flow (e.g. Fig. 2d, Zone 4), rather than approaching the trench from the mantle wedge, which is our primary area of interest. β is the estimated volume flux through the gap into the mantle wedge. Dashes in the volume flux column indicate that gap flow measurements were not made.

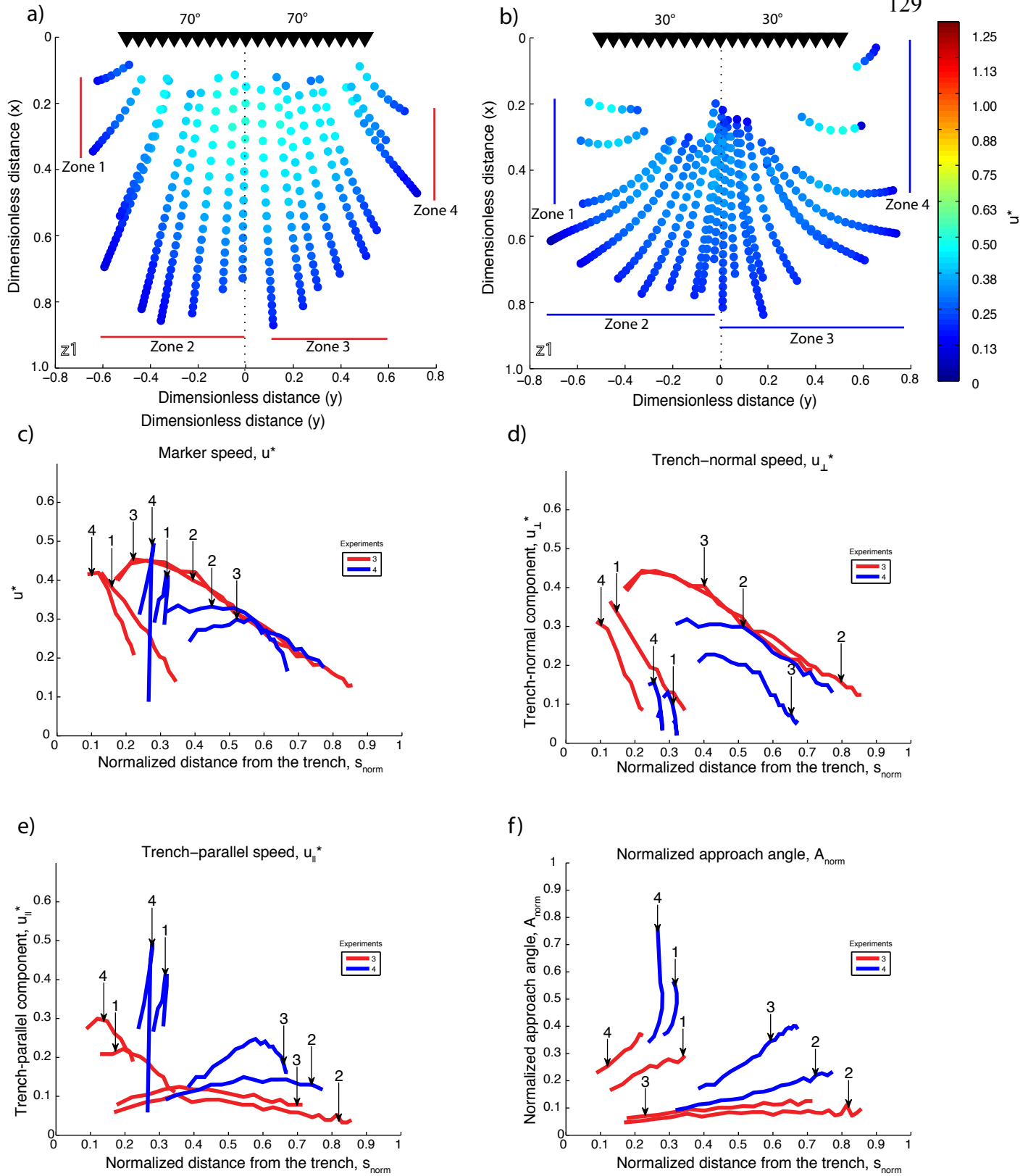


Figure A1. Comparison of Exps 3 ($U_D = 8$ cm/min) and 4 ($U_D = 8$ cm/min) at depth z_1 . These reference cases have a constant dip of a) 70° (Exp. 3) and b) 30° (Exp. 4), respectively. The marker speed, u^* (c), trench-normal speed component, $u_{\perp}^*$ (d), trench-parallel speed component, $u_{\parallel}^*$ (e), and normalized approach angle ($A_{\text{norm}} = A_W$ divided by 180° [= trench-normal]) (f) are shown for selected markers in zones 1-4 (denoted by red and blue lines in a-b). Black lines with triangles show the location of the trench. The dotted line represents the slab centerline. Slab segment dip values are shown above their corresponding halves of the trench.

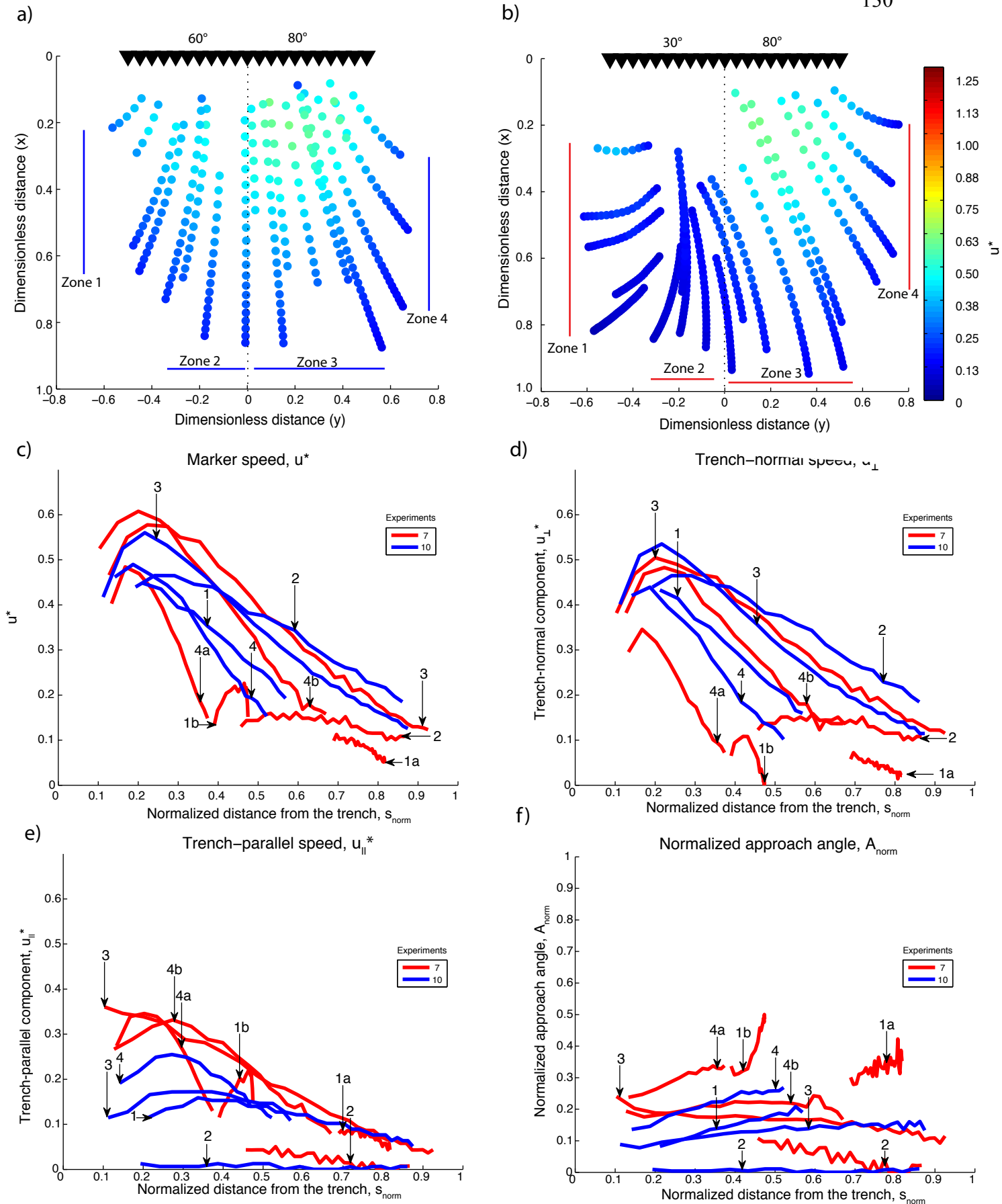


Figure A2. Comparison of Exps 10 ($U_D = 8$ cm/min) and 7 ($U_D = 8$ cm/min) at depth z_1 . Path line plots for (a) smallest gap width at the steepest possible dips (60°/80°, Exp. 10) and (b) the largest gap area (30°/80°, Exp. 7) are shown. The marker speed, u^* (c), trench-normal speed component, $u_{\perp}^*$ (d), trench-parallel speed component, $u_{\parallel}^*$ (e), and normalized approach angle (f) are shown for selected markers in zones 1-4 (denoted by red and blue lines in a-b), where in some zones, multiple markers are plotted to display the full range of variety. Black lines with triangles show the location of the trench. The black dotted line represents the slab centerline. Slab segment dip values are shown above their corresponding halves of the trench.

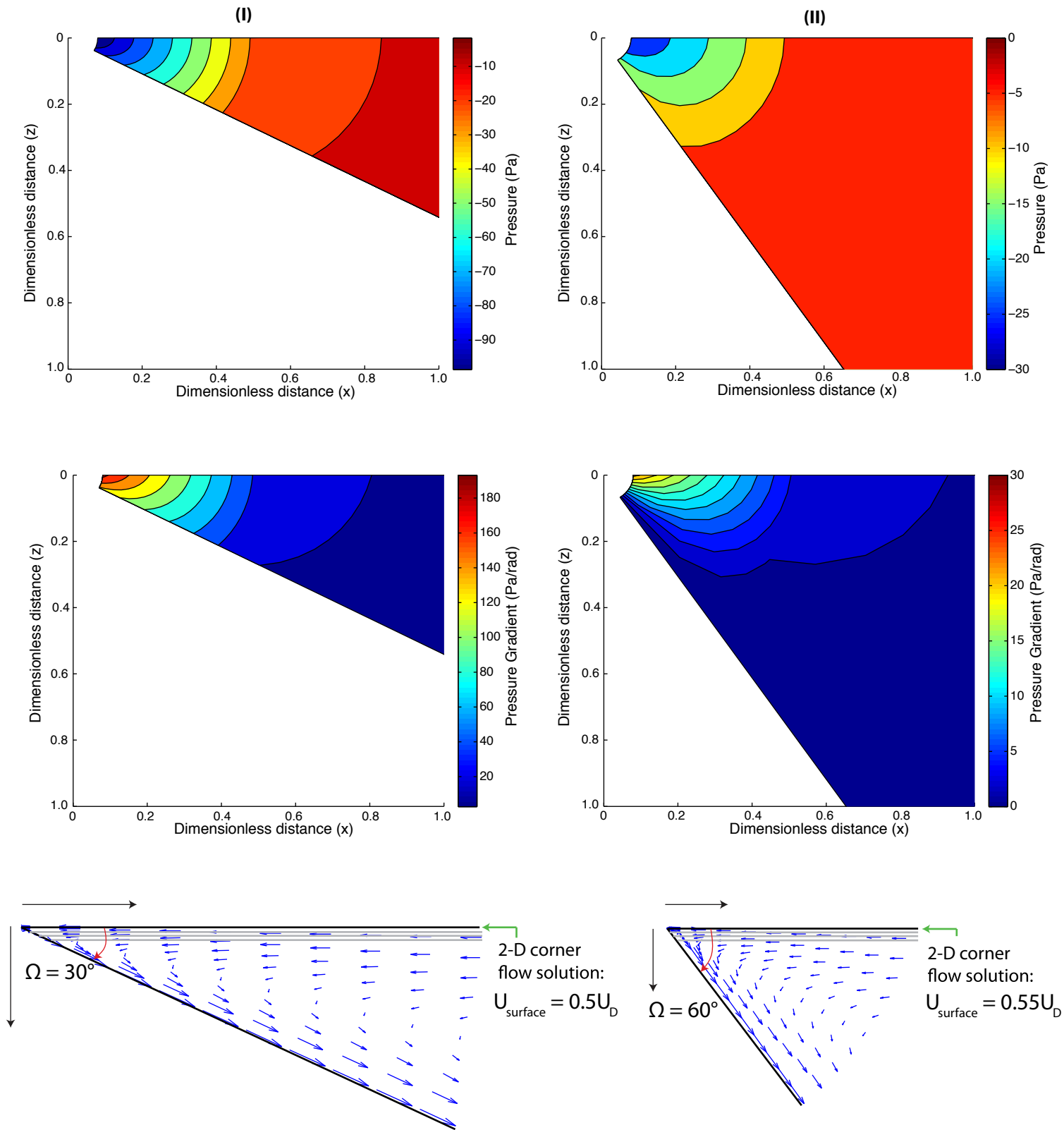


Figure A3. 2-D analytical corner flow solutions for the non-hydrostatic pressure (upper), pressure gradient (middle), and flow velocity (lower) values for (I) $\Omega = 30^\circ$ and (II) $\Omega = 60^\circ$. The grey lines in the flow field diagrams represent the depths at which our horizontal light sheets would fall on these slabs. Black arrows in the velocity plots correspond to the full extent of the vertical and horizontal axes displayed in the pressure and pressure gradient solutions. Distances are normalized by the total slab width ($W = 25$ cm). Note the different color bar ranges.

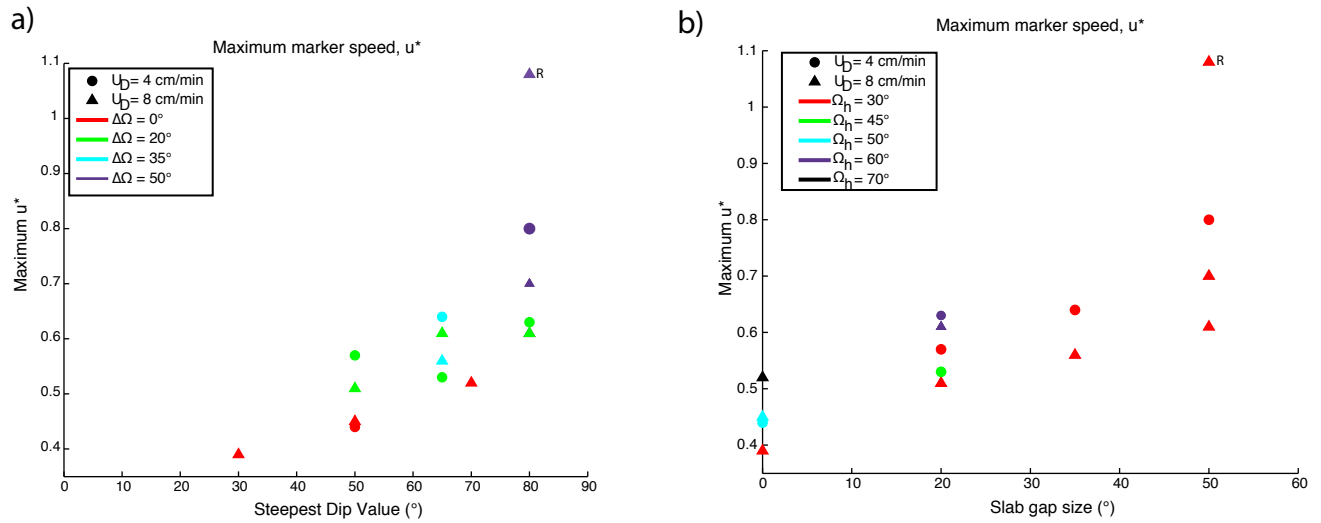


Figure A4. Summary of maximum marker speeds (u^*) plotted according to both slab gap size ($\Delta\Omega$) and steepest dip value. a) Maximum marker speeds calculated over all paths for level z_1 in each experiment, shown according to the steepest dip value in each experiment. (b) represents the same data as in part (a); we have changed the x-axis in this diagram to illustrate that the gap size, rather than the steepest dip value (a), drives the increase in u^* . $\Delta\Omega$ represents the slab gap size, and Ω_h represents the shallowest dipping slab in a given experiment. The 'R' in both (a) and (b) represents the values for Exp. 16, which is the only experiment where we added slab rollback.

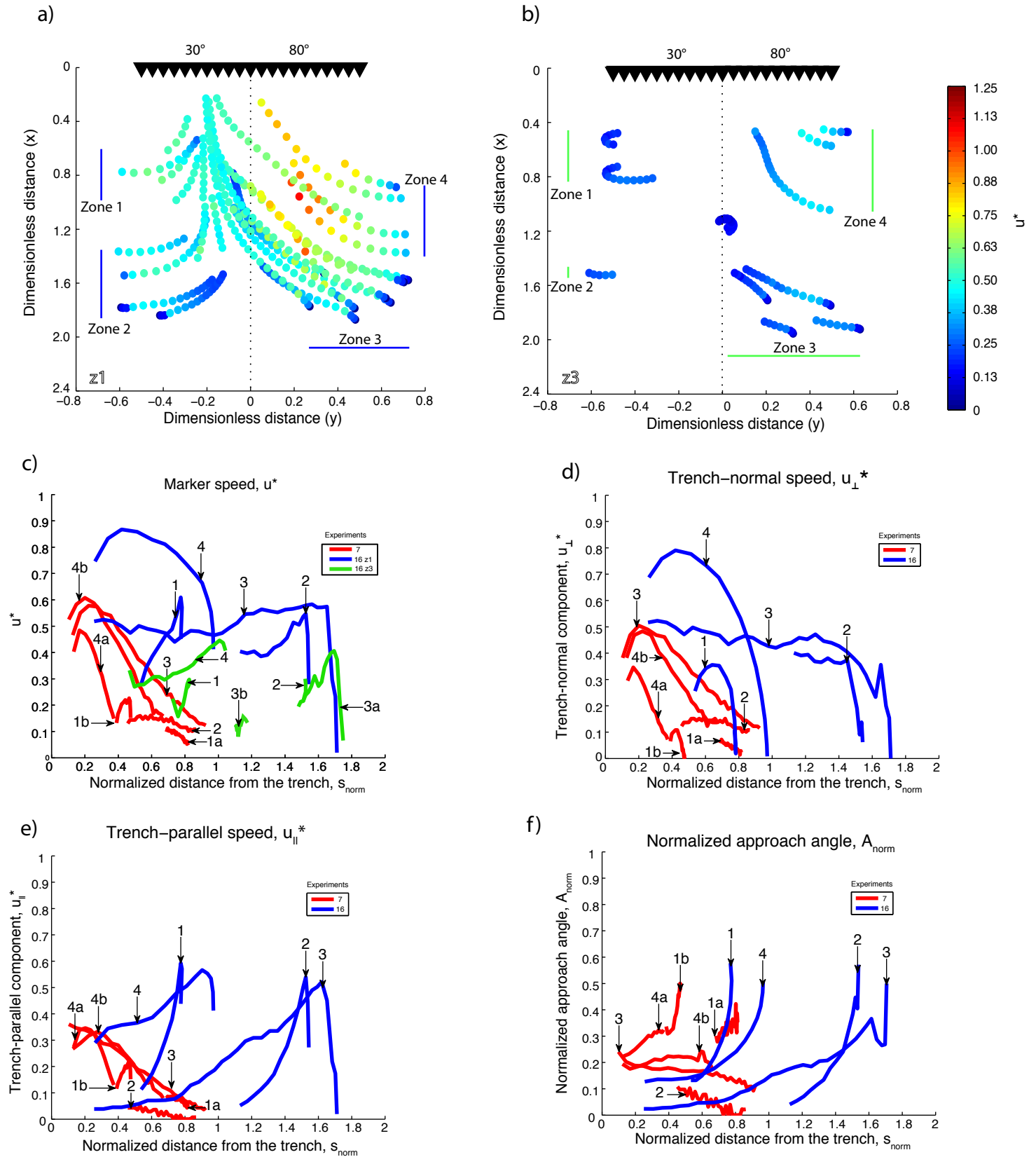


Figure A5. Comparison of Exp. 16 ($U_D = 8$ cm/min, $U_R = 3$ cm/min) at depths z_1 and z_3 , with additional data from Exp. 7 ($U_D = 8$ cm/min). The plots for an experiment with both the maximum slab gap size and rollback (Exp. 16) at depths of a) z_1 ($= 2$ cm ≈ 100 km) and b) z_3 ($= 7$ cm ≈ 350 km) are shown. The marker speed, u^* (c), trench-normal speed component, $u_{\perp}^*$ (d), trench-parallel speed component, $u_{\parallel}^*$ (e), and normalized approach angle (f) are shown for selected markers in zones 1-4 (denoted by blue and green lines in a-b), where in some zones, multiple markers are plotted to display the full range of marker characteristics within that zone. The conditions contrasted in (c-f) are those of the maximum slab gap size without (Exp. 7) and with (Exp. 16) rollback. Red and blue lines are for the z_1 plane comparison of Exp. 16 (Fig. A5a) and Exp. 7 (Fig. A2b), but green lines in (c) correspond to the z_3 plane of Exp. 16 (Fig. A5b). Black lines with triangles show the location of the trench. The dotted line represents the slab centerline. Slab segment dip values are shown above their corresponding halves of the trench.

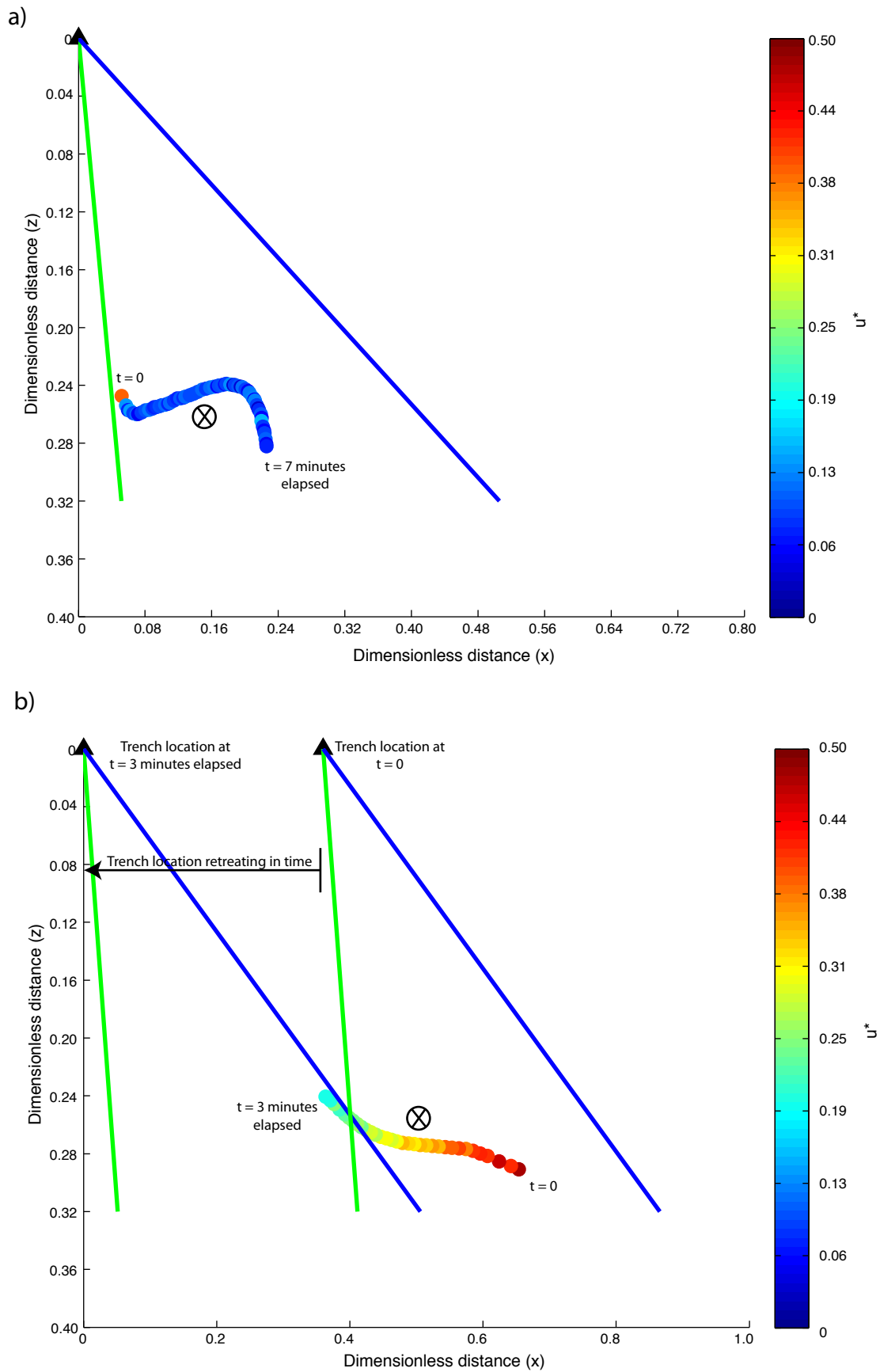


Figure A6. Marker paths over the given time intervals through the slab gap and into the mantle wedge for experiments with the maximum slab gap and subduction rate both a) without (Exp. 7) and b) with rollback (Exp. 16). Trench location is denoted by a black triangle, the steeply-dipping slab is denoted with a green line, and the shallowly-dipping slab is denoted by a blue line. Time values show the evolution of the trench and bead location during rollback. The circular symbol with an 'x' in it denotes that the trench-parallel sense of motion is from beneath the shallowly-dipping slab into the mantle wedge above the steeply-dipping slab (i.e. into the page). Distances are normalized by the total slab width ($W = 25$ cm).

Exp	Phase	Model Block	Backazaz (°)	Anisotropy (%)	FD (ϕ) (°)	ST (dt) (s)	ϕ error (°)	dt error (s)
1620	local S	10	0	7	-17	0.37	1	0.01
1620	local S	10	30	7	-18	0.45	0	0
1620	local S	10	60	7	-16	0.37	0	0
1620	local S	10	90	7	-9	0.39	0	0
1620	local S	10	120	7	0	0.42	0	0.01
1620	local S	10	150	7	-4	0.32	0	0
1620	local S	10	180	7	-17	0.37	1	0.01
1620	local S	14	0	12.5	-17	0.37	1	0.01
1620	local S	14	30	12.5	88	0.16	6	0.06
1620	local S	14	60	12.5	-15	0.37	0	0
1620	local S	14	90	12.5	-7	0.5	1	0
1620	local S	14	120	12.5	2	0.58	1	0.04
1620	local S	14	150	12.5	-34	0.16	1	0
1620	local s	14	180	12.5	-17	0.37	1	0.01
17	local S	10	0	12.5	-8	0.25	0	0
17	local S	10	30	12.5	-22	0.64	1	0.05
17	local S	10	60	12.5	-18	0.5	1	0.01
17	local S	10	90	12.5	-11	0.47	1	0.01
17	local S	10	120	12.5	6	0.18	0	0
17	local S	10	150	12.5	-7	0.41	1	0.01
17	local S	10	180	12.5	-8	0.25	0	0
17	local S	14	0	21	16	0.05	45	0.04
17	local S	14	30	21	-39	0.5	1	0.01
17	local S	14	60	21	-32	0.44	0	0
17	local S	14	90	21	-20	0.49	0	0
17	local S	14	120	21	-25	0.1	4	0.02
17	local S	14	150	21	-31	0.39	0	0
17	local S	14	180	21	16	0.05	45	0.04
1620	SK(K)S	10	0	12	-15	0.72	0	0
1620	SK(K)S	10	30	12	-16	0.66	0	0
1620	SK(K)S	10	60	12	-11	0.35	0	0
1620	SK(K)S	10	90	12	-15	0.84	0	0
1620	SK(K)S	10	120	12	-12	0.72	0	0
1620	SK(K)S	10	150	12	-12	0.63	0	0
1620	SK(K)S	10	180	12	-14	0.69	0	0
1620	SK(K)S	14	0	21.5	-78	0.32	0	0
1620	SK(K)S	14	30	21.5	-13	0.67	0	0
1620	SK(K)S	14	60	21.5	-8	0.95	0	0
1620	SK(K)S	14	90	21.5	1	3.11	0	0.21
1620	SK(K)S	14	120	21.5	-11	0.59	0	0
1620	SK(K)S	14	150	21.5	-9	0.78	0	0
1620	SK(K)S	14	180	21.5	-19	0.36	0	0
17	SK(K)S	10	0	16.5	-15	0.64	0	0
17	SK(K)S	10	30	16.5	-17	0.67	0	0
17	SK(K)S	10	60	16.5	-14	0.47	0	0
17	SK(K)S	10	90	16.5	-16	0.9	0	0
17	SK(K)S	10	120	16.5	-14	0.77	0	0
17	SK(K)S	10	150	16.5	-13	0.63	0	0
17	SK(K)S	10	180	16.5	-15	0.61	0	0
17	SK(K)S	14	0	54	-33	0.77	0	0
17	SK(K)S	14	30	54	-30	0.6	0	0
17	SK(K)S	14	60	54	-32	0.54	0	0
17	SK(K)S	14	90	54	-36	0.45	0	0
17	SK(K)S	14	120	54	-31	0.7	0	0
17	SK(K)S	14	150	54	61	0.02	90	0.02
17	SK(K)S	14	180	54	-34	0.91	0	0

Supplementary Table 1. The results of our synthetic shear-wave splitting model through a range of backazimuths in model blocks 10 and 14. Olivine a -axis azimuths (not shown) were calculated by rotating anisotropic olivine elastic constants to the azimuth of each whisker within the block, and then averaging over all of the whiskers. Strengths of anisotropy (%) for each block were determined based on the value that yielded splitting time (dt) averages closest to the benchmarks discussed in the text of 0.38 s for local S waves, and 0.69 s for SK(K)S waves. Nulls (grey shaded measurements) were not included in the fast direction (ϕ) or splitting time (dt) averages, nor were they included in the determination of the anisotropy percentages shown in Fig. 10. When uncertainties have values of 0, it means that the uncertainties, while not actually zero, are quite small.

Chapter 3

Uppermost Mantle Velocities Beneath the Southern Appalachians from Pn Phases

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Summary

Pn phases from regional earthquakes recorded in northern Georgia by stations of the EarthScope SESAME and Transportable arrays were used to solve for uppermost mantle P-wave velocity. Mantle velocities vary laterally, with values of ~ 7.9 km/s beneath southern edge of the Blue Ridge Mountains that increase to values of 8.4-8.5 km/s further south. The low velocity zone could represent a source of buoyancy that helps to support the mountains, although additional support from the crustal root that overlies the low velocity zone is also likely. Azimuthal anisotropy in upper mantle P-wave velocity is not required by the data, but $\sim 9\%$ anisotropy with a fast direction of 67° provides the best fit to the Pn phases in the south. This fast direction is parallel to the strike of the Appalachian orogenic terranes in this region, and is consistent with compressional deformation in the mantle lithosphere during their development.

Index Terms: Body Waves, Mantle

Keyword: Pn phases, mantle velocity, southern Appalachians

1. Introduction

Present-day topography in the Appalachian orogen is not consistent with simple erosional decay of the original late Paleozoic landscape (e.g. McKeon et al., 2014). In the southern Appalachians, the greatest present-day elevations occur in the Blue Ridge Province (Figure 1), whereas the high topography and thickened crust of the Paleozoic Appalachians presumably extended across the accreted terranes to the southeast, areas that now lie at lower elevations. A substantial root of thickened crust remains beneath the Blue Ridge Mountains, and crustal thickness and elevation are broadly correlated throughout the southernmost Appalachians (Hawman, 1996; French et al., 2009; Hawman et al., 2012; Parker et al., 2013). The buoyancy of the crustal root relative to the mantle has likely been a significant factor in supporting the high Blue Ridge topography relative to its lower elevation surroundings. However, contributions from mantle buoyancy have been more difficult to assess given a lack of spatial resolution in mantle velocity structure.

In this study, we obtain new resolution of P-wave velocity (V_P) structure in the uppermost mantle using Pn phases recorded by stations of the Southeastern Suture of the Appalachian Margin Experiment (SESAME, an EarthScope Flexible Array) and the EarthScope Transportable Array (TA). These results enable us to evaluate possible mantle buoyancy variations beneath the Blue Ridge Mountains in northern Georgia, as well as V_P azimuthal anisotropy and its implications for mantle deformation.

2. Pn Data

Despite the paucity of large earthquakes in the eastern United States, three regional earthquakes produced clear Pn phases from back-azimuths suitable for differential time analysis at SESAME stations. The 23 August 2011 Mineral, Virginia ($M_w = 5.8$) earthquake and the 6 November 2011 Shawnee, Oklahoma ($M_w = 5.6$) earthquake occurred when the SESAME array was only about 50% deployed and were recorded by stations on the W- and D-lines (Figure 1). The 17 May 2012 Timpson, Texas ($M_w = 4.8$) earthquake was recorded by the complete SESAME array and by TA stations in the region. To isolate mantle structure beneath the array, we differenced Pn arrival times between stations of the D-line (and TA stations Y54A and X53A) and those of the W-line (plus TA stations X52A and Y53A, and U.S. ANSS station GOGA). To minimize contributions from structure outside the array to arrival-time differences, the back-azimuthal difference between paths to either station in each pair was limited to less than 3.1° . To enable assessment of azimuthal anisotropy relative to the effects of lateral variations in V_p , the analysis was limited to stations north of 32.9°N where intersecting paths from different earthquakes sample comparable areas (Figure 1). Epicentral distances vary from 5.5° to 11.7° , a range for which Pn phases should be sensitive to mantle structure just below the crust-mantle boundary (Moho) (Smith and Ekström, 1999).

Pn arrival times were picked from seismograms that had been causally filtered with a bandpass of 0.5-2.0 Hz (Figure 2), and the times were differenced between stations

within the 3.1° back-azimuth range. At these frequencies, Pn onsets are sharp at many stations, and waveforms with particularly emergent or complex waveforms were omitted. For the Virginia and Oklahoma earthquakes, we also calculated differential times between stations by cross-correlating windowed Pn phases that had been filtered with a zero-phase bandpass of 0.05-0.15 Hz (Figure S1). The Texas earthquake, which had a smaller magnitude, did not produce sufficient Pn signal-to-noise ratios at these frequencies.

3. Solving for Mantle Velocity

Given the back-azimuth limit for station pairs, we assumed that the Pn phases to each station in a pair sampled similar near-source crustal structure and mantle structure prior to the point where the phase began its upward leg towards the station (Figure 3a). To account for variations in crustal structure beneath each station, we employed the crustal thicknesses obtained by Parker et al. (2013) from H-k stacking of Ps phases, and we assumed their average crustal V_P values of 6.4 km/s for stations in the Inner Piedmont terrane and 6.5 km/s for all others. These V_P values were based on refraction studies (Hawman et al., 2012). For two stations not included in the Parker et al. (2013) analysis, we performed H-k stacking analyses, and found crustal thicknesses of ~ 41 km for W27 and ~ 36 km for W28. Moho dips for each Pn path segment below the array were based on crustal thicknesses, and vary from 1.4° to 3.6° .

Pn differential times were divided into three groups based on the mantle volume sampled by the path to the more distant station (Figure 1). Station pairs in each group are shown in Figure S2. For each of these groups, the differential times were used to solve for V_P on the mantle leg of the phase path. Because the critical angle of the Pn phase varies as a function of mantle velocity relative to the assumed crustal velocity at each station, mantle velocities were determined iteratively. Mantle path lengths and crustal corrections to the differential times were found for an initial mantle velocity, and corrected differential times from mantle path lengths greater than 50 km were used to jointly solve for a new mantle velocity. Assuming the new mantle velocity, mantle path lengths and crustal corrections were re-calculated in the next iteration and used to solve for the next mantle velocity. Iterations were continued until mantle velocity variations between iterations were less than 0.05 km/s.

This approach assumes that Pn phases travel on great circle paths. To test for the presence of significant ray bending or multi-pathing, we conducted a beamforming analysis in which we solved for the back-azimuth and horizontal phase velocity that generate the most coherent wavefront across the array of stations, using the 0.05-0.15 Hz bandpassed waveforms for either the Virginia or Oklahoma earthquake. For both events, the back-azimuth that generated the most constructive interference among the waveforms at all stations was within $\pm 0.5^\circ$ of the mean back-azimuth of the great circle paths. Thus we conclude that using the great circle back-azimuths is a reasonable approximation.

4. Results: Uppermost Mantle Velocities

Uppermost mantle velocities obtained with this approach vary between the geographic groups. The mantle V_P value based on all differential times in the northern group of paths (7.88 ± 0.27 km/s) is significantly lower than the mantle V_P value from all times in the southern group of paths (8.47 ± 0.20 km/s) (Figure 4a; Table 1). The mantle V_P value for the middle group of paths (8.38 ± 0.26 km/s) is similar to the V_P for the southern group, and its one standard deviation uncertainty only slightly overlaps with the uncertainty of the northern group V_P . The fits of the V_P values to the differential times and mantle path lengths for each group are shown in Figure 3b. The V_P value for the northern group of paths agrees with an active source refraction line V_P estimate of ~ 7.9 km/s for the mantle starting at $\sim 35^\circ\text{N}$ on the western edge of the Blue Ridge terrane and extending northwest across the Valley and Ridge and Cumberland Plateau provinces (Prodehl et al., 1984). The V_P estimate for the middle group of paths overlaps with a mantle V_P estimate of ~ 8.2 km/s based on Pn phases from nearby explosions (Dorman, 1972). Zhang et al. (2009) inferred a mantle V_P of 8.1 km/s throughout most of our study region using Pn phases from earthquakes, perhaps reflecting the more widely spaced stations available for their study.

To evaluate evidence for azimuthal anisotropy, we also solved for mantle V_P values based on the differential times for each event in each path group (Figure 4a; Table 1). For the Virginia and Oklahoma earthquakes, which have differential times from both Pn onset picks and waveform cross-correlation, velocities based on only one observation

type are similar to each other and to the velocity based on the combined observations. Velocities for each earthquake in a given path group overlap at one standard deviation, and hence the data do not require azimuthal anisotropy. However, in the northern and middle path groups, the best-fitting velocity for the Oklahoma earthquake is lower than the velocities for the other events, a result that is consistent with azimuthal anisotropy, given the different event back-azimuths.

To assess models of azimuthal anisotropy that are consistent with the Pn data, we fit the back-azimuthally varying event-specific Pn velocities from the middle and southern path groups with $\sin(2\theta)$ functions (Figure 4b). Pn velocities from the northern group were not included because the average velocity from this group is significantly lower. The $\sin(2\theta)$ variation that best fits the Pn velocities as a function of azimuth has a fast direction of 67° and a strength of 9%. A strength of 9% falls in the range of values typically observed for azimuthal anisotropy in Pn mantle velocities (Hearn, 1996; Smith and Ekström, 1999; Baldock and Stern, 2005; Buehler and Shearer, 2010; Buehler and Shearer, 2012). A fast direction of 67° is similar to the average of the Pn velocity fast directions inferred by Zhang et al. (2009) for this region, although the Zhang et al. (2009) anisotropy is much weaker than 9%. 67° is also broadly consistent with the closest Pn velocity fast direction inferred by Smith and Ekström (1999), with the exception that our estimate is more easterly.

The possibility of azimuthal anisotropy raises the issue that differing proportions of paths from the three earthquakes could bias the average mantle V_p values for each path

group. However, the best-fitting mantle V_P values from the Texas earthquake are lower in the northern group relative to more southern paths, indicating that lower uppermost mantle V_P values exist beneath the higher topography even if azimuthal anisotropy is present.

5. Implications for the Southern Appalachian Mantle

The zone of lower mantle V_P values found beneath the array near $\sim 34.5^\circ\text{N}$ (i.e. the northern group) likely resides within Laurentian mantle lithosphere. Interpretations of reflection profiles in northern Georgia typically indicate that the deep crust and mantle lithosphere of Laurentia extend far to the east beneath the Coastal Plain, and that the Blue Ridge, Inner Piedmont, and Carolina terranes are thin crustal slivers that were thrust hundreds of kilometers to the northwest over Laurentian crust and mantle (e.g. Cook and Vasudevan, 2006). This view is supported by initial S_p imaging of crust and mantle discontinuities with data from SESAME and TA stations (Parker et al., 2013; Hopper et al., 2013). In an alternative model, Hibbard et al. (2010) suggest that northwest-directed subduction created a suture between the lower crusts of Laurentia and the Carolina terrane in the vicinity of the transition from high to low P_n mantle velocities. While this scenario implies the possibility of a suture between Laurentian and Carolina mantle lithospheres, the low velocity zone resolved in this study would still lie within Laurentian lithosphere. For either scenario, the low V_P zone could represent mantle with an anomalous composition, for example volatile-rich partially-serpentinized mantle (Hacker et al., 2003). Anomalously high temperatures would also produce low V_P , but the source

of a localized high temperature zone in this tectonic setting is not obvious. Alternatively, the low V_P zone may not in fact be mantle at all. It could be crust with an anomalously high velocity, such as a deep layer in the Appalachian crustal root that experienced garnet growth and partial eclogitization as it cooled. In this case, the seismic interface identified as the Moho by Ps phases and active source imaging (Hawman et al., 2012; Parker et al., 2013) would instead be a metamorphic and/or compositional boundary within the deep crust. Additional information on the lateral extent of the low V_P zone and corresponding shear-wave velocity constraints are needed to better assess its origins.

The implications of the northern zone of anomalous V_P values for buoyancy and support of Blue Ridge topography depend on its origin. If it represents the edge of a body of serpentinized mantle, the serpentinized mantle would be relatively low density and would contribute to supporting the higher elevations. If the anomalous V_P zone is instead partially eclogitized lower crust, it could have a lower or higher density than the surrounding mantle, depending on the silica content of the initial crustal root and the degree of eclogitization. Estimates of crustal root buoyancy (mantle density minus crustal root density) for the southern Appalachians, based on the depth of the seismically-observed Moho, are $\sim 100\text{-}160 \text{ kg/m}^3$ (French et al., 2009; Hawman et al., 2012). These buoyancy values are consistent with a global trend in which crustal root buoyancy tends to decrease with orogenic age (Fischer, 2002). If the anomalous V_P zone is positively buoyant mantle, the southern Appalachian crustal root buoyancy estimates would be even lower, further contributing to the global trend. If the anomalous V_P zone represents a fragment of garnet-rich, partially eclogitized lower crust below the seismically-imaged

Moho, it would increase the actual thickness of the southern Appalachian crustal root, and would also be consistent with the global trend.

If azimuthal anisotropy of $\sim 9\%$ exists, the 67° fast direction implied by the anisotropy is roughly parallel to the strike of Appalachian orogenic terranes at these latitudes; they are consistent with the orientation of olivine a -axes that would be predicted for compression in the mantle lithosphere normal to orogen strike, assuming A, C, or E olivine fabric types (e.g. Ribe, 1989; Karato et al., 2008). However, we reiterate that azimuthal anisotropy is not required by the data at a one standard deviation level of confidence.

6. Conclusions

Pn data from three central and eastern U.S. earthquakes recorded by northern stations of the SESAME array indicate that upper mantle Pn velocities are lowest beneath the southern edge of the Blue Ridge Mountains of northern Georgia and increase to the south. The area of anomalously low V_p beneath the Blue Ridge Mountains could represent a region of buoyant upper mantle material that, together with thicker crust, helps to support their high elevations. Alternatively, the low velocity is also consistent with a localized zone of garnet-rich, partially eclogitized lower crustal material that exists below the velocity increase interpreted as the Moho. While azimuthal anisotropy is not required to explain variations in mantle V_p estimates between events, the data are best fit

by 9% anisotropy with a fast direction (67°) that is normal to the direction of compression during terrane accretion in this region of the Appalachian orogen.

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We thank the many people who made the SESAME array possible: the staff of the IRIS PASSCAL Instrument Center, the student/post-doc field crew, and the landowners who hosted stations. Thanks also to Emily Hopper who contributed H-k stacking at stations W28 and W27, and to Don Forsyth for his insights about Pn phases. This work was funded by the NSF EarthScope Program under the American Recovery and Reinvestment Act of 2009 (Public Law 111-5) through awards EAR-0844276 (K.M.F.), EAR-0844186 (L.S.W.), and EAR-0844154 (R.B.H.).

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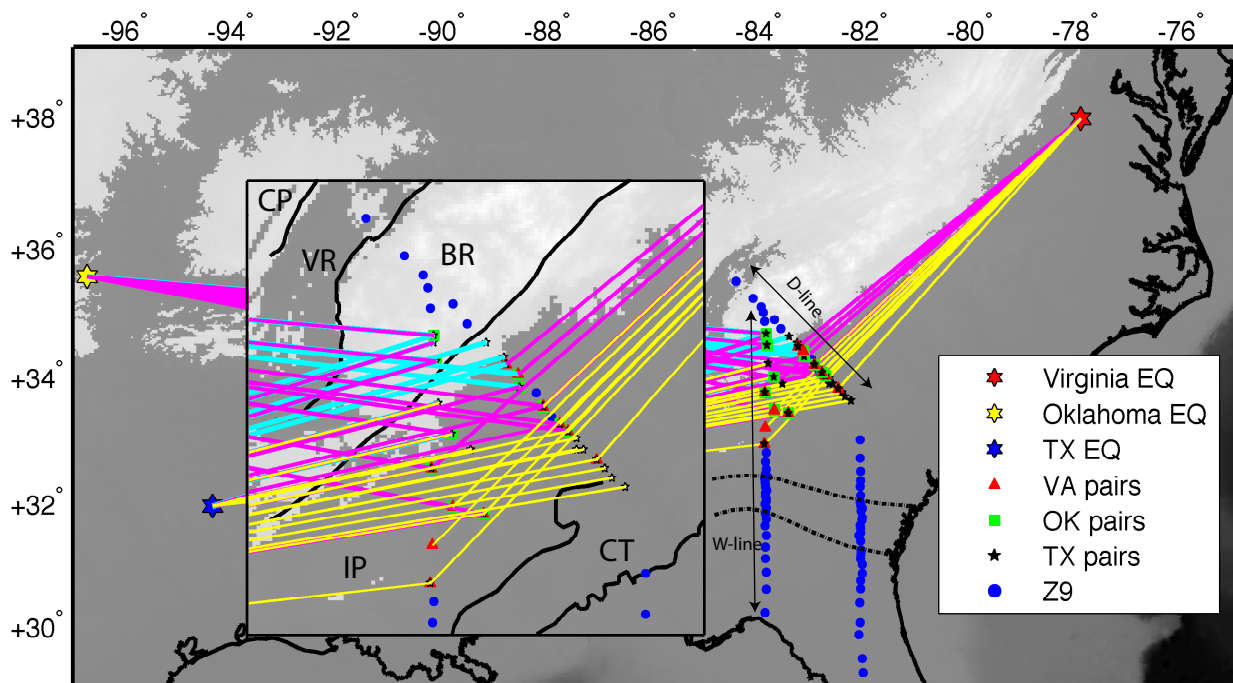


Figure 1. Pn paths superimposed on topography. Large stars show event locations. SESAME and TA stations used in the Pn velocity analysis are shown by different symbols based on whether they contributed data for the Virginia earthquake (red triangles), Oklahoma earthquake (green squares), and Texas earthquake (black stars). Blue circles show SESAME stations not used in the analysis. Pn paths are shown for the northern group (cyan), the middle group (magenta), and the southern group (yellow). Black lines in the inset are terrane boundaries. CP: Cumberland Plateau. VR: Valley and Range terrane. BR: Blue Ridge terrane. IP: Inner Piedmont terrane. CT: Carolina terrane. RB: South Georgia Rift Basin. SW: Suwannee terrane. Dashed lines in the background map show the Suwannee Suture.

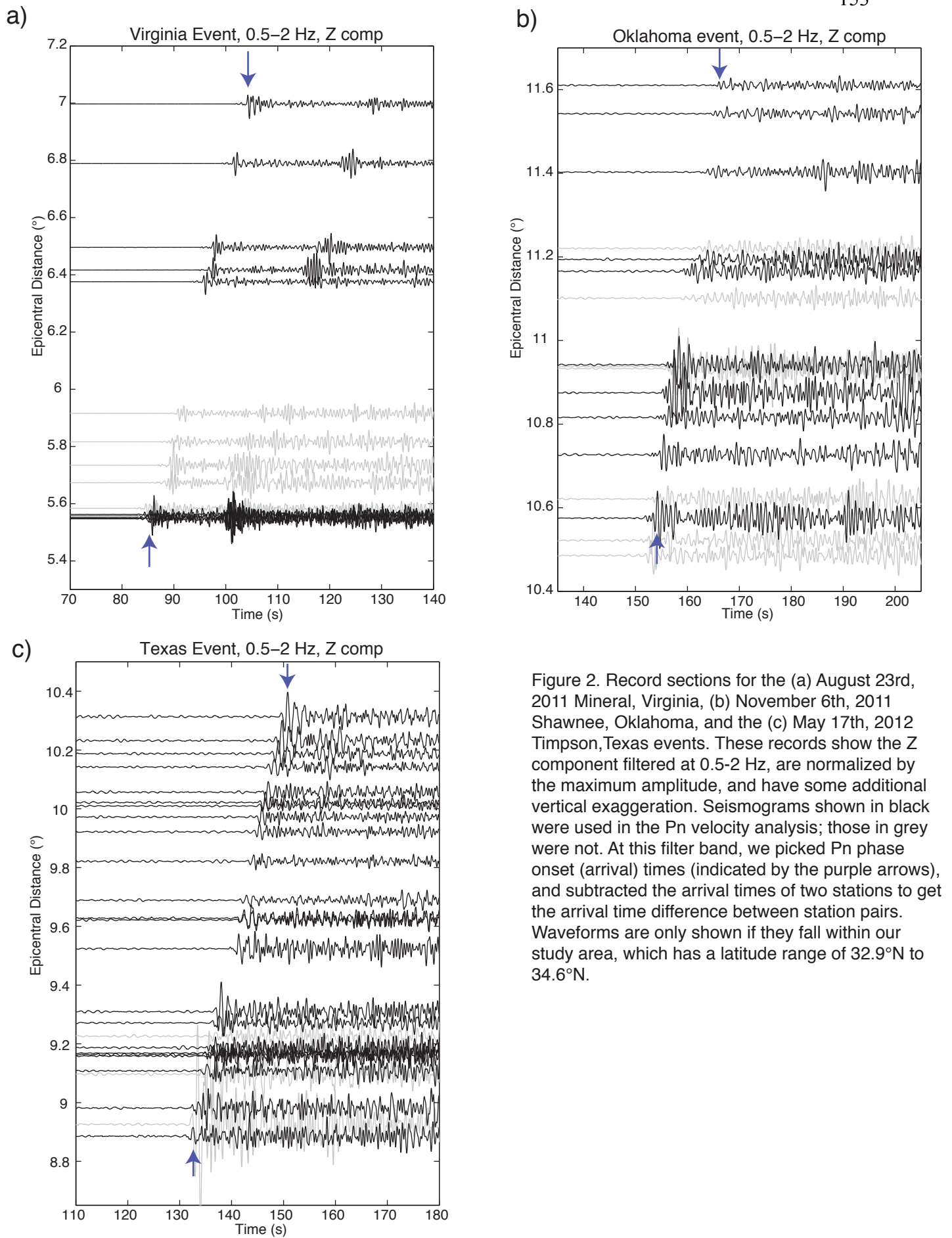


Figure 2. Record sections for the (a) August 23rd, 2011 Mineral, Virginia, (b) November 6th, 2011 Shawnee, Oklahoma, and the (c) May 17th, 2012 Timpson, Texas events. These records show the Z component filtered at 0.5-2 Hz, are normalized by the maximum amplitude, and have some additional vertical exaggeration. Seismograms shown in black were used in the Pn velocity analysis; those in grey were not. At this filter band, we picked Pn phase onset (arrival) times (indicated by the purple arrows), and subtracted the arrival times of two stations to get the arrival time difference between station pairs. Waveforms are only shown if they fall within our study area, which has a latitude range of 32.9°N to 34.6°N.

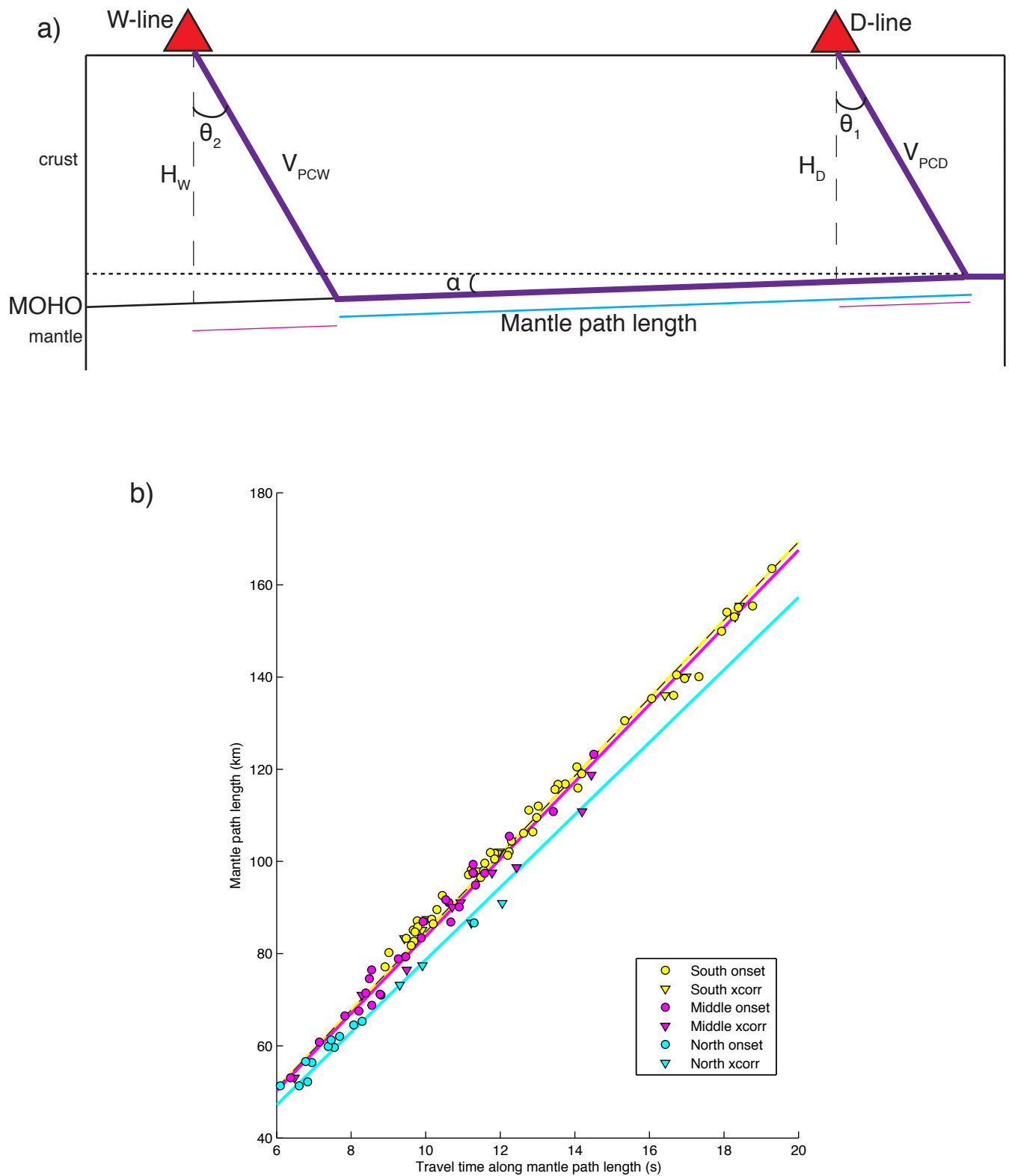


Figure 3. (a) Parameters and (b) results from our Pn velocity analysis. H_D and H_W , V_{PCD} and V_{PCW} , and θ_1 and θ_2 are the crustal thicknesses, the crustal velocities, and the critical angles beneath the W and D line stations, respectively. α represents the dip of the Moho, measured from the horizontal. The blue line represents the mantle path length, and the pink lines represent the extent of the crustal corrections for the W- and D-line stations. In general, the Oklahoma and Texas events have paths that point up-dip (from the W-line to the D-line), while the Virginia event has paths that point down-dip (from the D-line to the W-line). The results of our Pn velocity analysis, both with cross-correlation (upside-down triangles) and arrival time difference (circles) calculations, are plotted above. The colored lines (the yellow line has a black dashed line through it to make it more visible) correspond to the best-fitting Pn velocity for each of the groups ('All paths' values in Table 1).

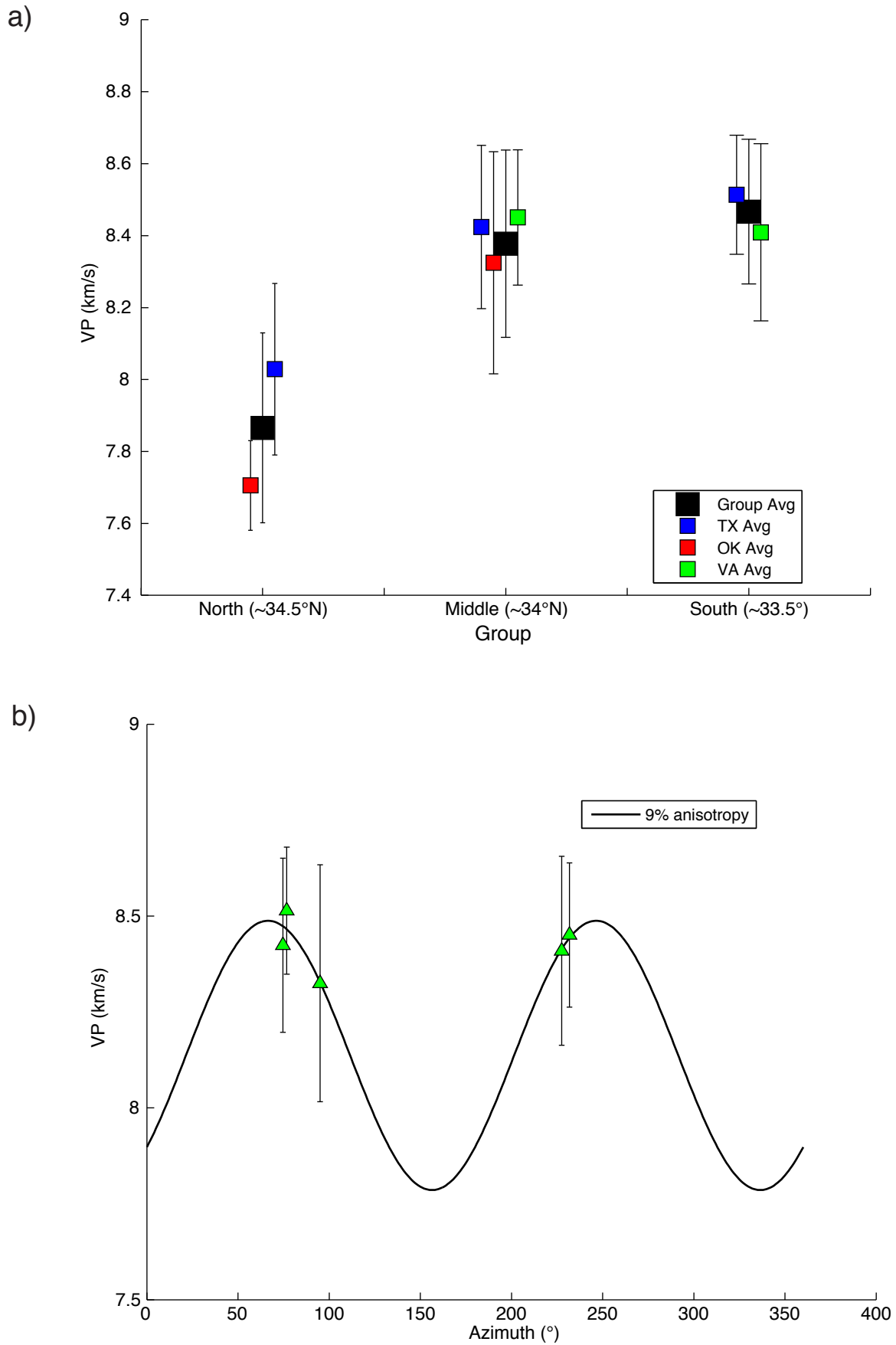


Figure 4. (a) Best-fitting uppermost mantle Pn velocity values shown by group (black squares) and by earthquake (colored squares). (b) Pn velocity results displayed as a $\sin(2\theta)$ azimuthal anisotropy plot. We observe lateral variation in the upper mantle V_p values, with slower values towards the north, and faster values to the south. Green triangles are upper mantle V_p estimates and their uncertainties from Pn phases for individual earthquakes in the Middle and Southern Groups. This $\sin(2\theta)$ curve represents 9% azimuthal anisotropy.

Group	Mantle V_p (km/s)	Number of differential times
<u>Northern Group</u>		
All paths	7.88 ± 0.27	14
OK paths	7.71 ± 0.12	5
TX paths	8.09 ± 0.19	9
<u>Middle Group</u>		
All paths	8.38 ± 0.26	36
OK paths	8.32 ± 0.31	17
TX paths	8.42 ± 0.23	14
VA paths	8.45 ± 0.19	5
<u>Southern Group</u>		
All paths	8.47 ± 0.20	58
TX paths	8.51 ± 0.17	34
VA paths	8.41 ± 0.25	24

Table 1. Calculated upper mantle Pn velocities. Uncertainties are calculated using the sum of the square of the residuals between the individual station pair Pn velocities and the best-fitting Pn velocities for that group.

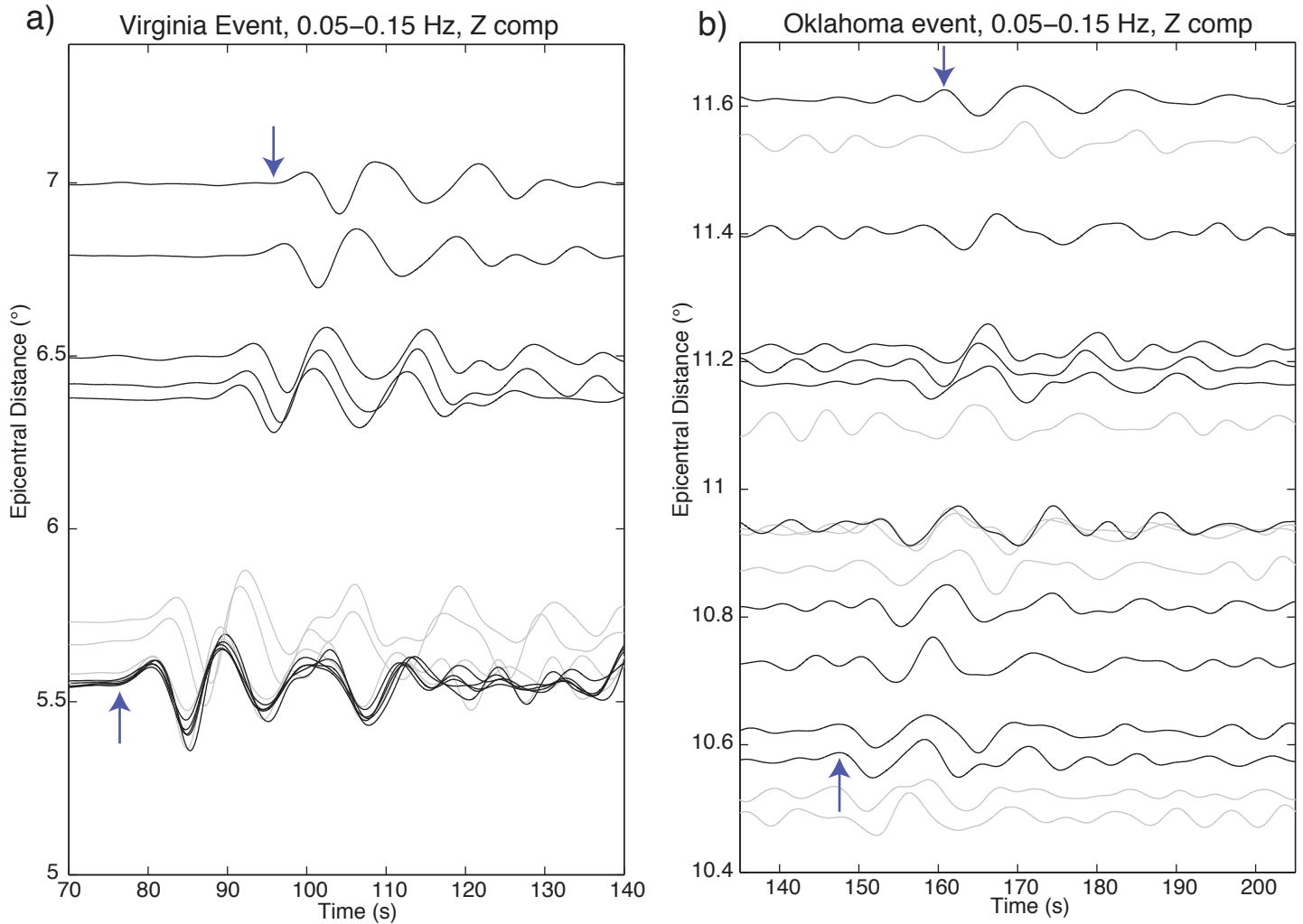


Figure S1. Record sections for the (a) August 23rd, 2011 Mineral, Virginia and (b) November 6th, 2011 Shawnee, Oklahoma events. Both records are displaying the Z component filtered at 0.05–0.15 Hz, are normalized by the maximum amplitude, and have some additional vertical exaggeration. Seismograms shown in black were used in the Pn velocity analysis; those in grey were not. At this filter band, we windowed the Pn phases (with the windows starting at the point in time indicated by the purple arrows) and used cross-correlation to determine phase arrival time differences between station pairs. Waveforms are only shown if they fall within our study area, which has a latitude range of 32.9°N to 34.6°N.

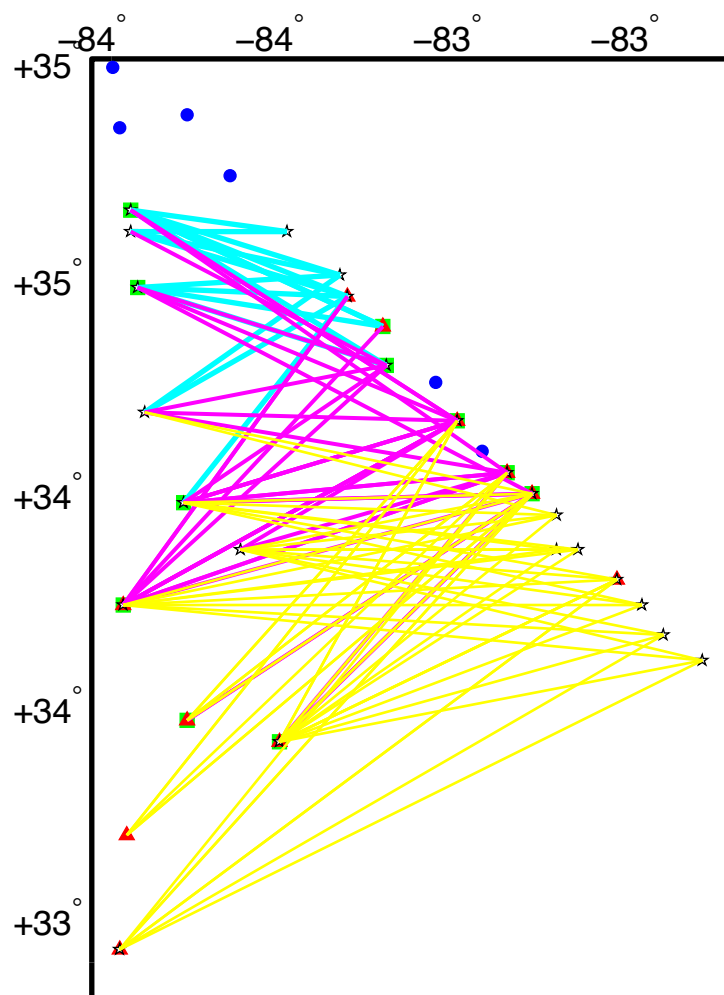


Figure S2. Station pair map. Station pairs used in the Pn velocity analysis are connected by a line with a color that corresponds to the group divisions shown in Figure 1.

Chapter 4

Teleseismic shear-wave splitting in the southeastern US and implications for lithospheric and asthenospheric anisotropy

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Summary

We have explored lithospheric and asthenospheric seismic anisotropy in the southeastern United States through teleseismic shear-wave splitting analyses. The dataset for these analyses comprises teleseismic SK(K)S phases from the SESAME array (7/2010-5/2014) and teleseismic SKS phases from the stations in the Transportable Array that overlap the SESAME array in both time and space. Within the SESAME array, fast direction variation is attributed to different anisotropic fabrics in the lithosphere, including both fossilized fabrics (NE-SW oriented fast directions in the Blue Ridge terrane and E-W oriented fast directions in the Inner Piedmont terrane) and a potential rifting signature (NW-SE fast directions) in eastern Georgia due to the Mesozoic rifting that created the South Georgia Rift Basin. We find that the backazimuthal dependence of fast direction measurements in some of these places, such as within and south of the Suwannee suture zone, as well as in eastern Georgia, indicates that these fast directions can be modeling using two or three layers of anisotropy, where the deepest layer is representative of asthenospheric mantle flow, and the upper layer(s) is representative of lithospheric fabrics. Similar variations are observed regionally with the Transportable Array data; here, the variation suggests either distinct fast direction differences between cratonic Laurentia and the foreign Suwannee terrane, or more active flow in the asthenosphere driven by interference of the cratonic keel or the remnant Farallon slab.

1. Introduction

Seismic anisotropy, or when a material has different seismic velocities depending on the path through that material, can be indicative of the material's composition or stress/strain state. On a small scale, time-varying anisotropy has been used to measure the evolving stress state of the crust surrounding active volcanoes (e.g. Savage et al., 2010a; Savage et al., 2010b; Johnson et al., 2011; Roman et al., 2011). On a larger scale, seismic anisotropy measured by teleseismic body waves has been used as a proxy for the direction and strength of asthenospheric mantle flow (e.g. Russo and Silver, 1994; Audoine et al., 2000; Anderson et al., 2004; Civello and Margheriti, 2004; Long and Silver, 2008; Abt et al., 2010; Long, 2013).

One of the most common ways to measure seismic anisotropy is shear-wave splitting. Because the medium is anisotropic, the seismic velocity through the medium is path-dependent. When body waves encounter an anisotropic medium, the incoming shear wave splits and propagates along two orthogonal axes within the medium: the 'slow' axis, and the 'fast' axis. With a higher seismic velocity, the shear wave propagating along the fast axis will arrive at the surface before the slow wave; the orientation of the fast shear wave at the surface, with respect to North (the fast direction, ϕ), and the difference in arrival time between the fast and slow shear waves (the splitting time, dt) are the two shear-wave splitting parameters that quantify the seismic anisotropy. The fast direction constrains the alignment of anisotropy within the medium generated by any number of processes (e. g. mineral lattice preferred orientation (LPO) fabrics, crack alignment/the presence of melt in the crust, etc.). The splitting time is a measure of the strength of anisotropy or the path length through the anisotropic medium.

Our study area, the southeastern US, has been previously characterized by teleseismic shear-wave splitting studies using broadband data at permanent and temporary stations. Most authors (Barruol et al., 1997; Liu, 2009; Wüstefeld et al., 2009; Long et al., 2010; Wagner et al., 2012) agree that when anisotropy is present, teleseismic shear-wave splitting yields fast polarization directions that are typically parallel to sub-parallel to the direction of absolute plate motion (APM) of the North American Plate or that are consistent with flow around the edges of thicker cratonic lithosphere (Fouch et al., 2000). However, some areas have splitting results that do not follow this trend. For instance, results from permanent station GOGA (IRIS/US) show splitting results that are either normal to APM (Vauchez and Barruol, 1996; Liu, 2009; Wüstefeld et al., 2009) or null, that is, no splitting was observed (Long et al., 2010; Wagner et al., 2012). These types of discrepancies between the predicted fast directions due to APM and the observed fast directions at the surface are potentially indicative of ‘fossilized’, or frozen-in anisotropy in the lithosphere.

Anisotropy in the lithosphere is typically attributed to mineral fabrics that retain information about the tectonic environment (i.e. compressional, tensional, strike-slip, etc.) at a point in time. In an accretionary environment like the southeastern US, however, different terranes may be preserving information about the tectonic environment from when they were not yet attached to cratonic Laurentia. Therefore, any teleseismic shear wave that samples the lithosphere in this part of the US may be sampling fabrics with pre-accretionary and/or post-accretionary alignments. Wagner et al., (2012) find many teleseismic fast directions that are parallel to the New York-Alabama (NY-AL) lineament and the Brunswick Magnetic Anomaly (BMA); the former comprises a series of faults

and/or terrane boundaries that essentially outline the trace of the Appalachian Mountains, and the latter may indicate the extent of the boundary generated by the collision of the exotic Suwannee terrane and cratonic Laurentia that began at ~320 Ma (Secor et al., 1986).

Intrigued by the seeming sensitivity of teleseismic shear-waves to the hypothesized Suwannee suture zone, we conducted a teleseismic shear-wave splitting analysis on data collected on the temporary SESAME deployment (7/2010 – 5/2014). We also included nearby TA (Transportable Array) stations. We find small-scale fast direction variations that can be explained by a combination of fossilized lithospheric anisotropy fabrics and asthenospheric flow.

2. Tectonic Setting

Our study area is dominated by structure due to a series of major tectonic events: the Taconic Orogeny, the Acadian Orogeny, and the Alleghanian Orogeny. The accreted terranes of the Taconic and Acadian Orogenies, in order of increasing distance from cratonic Laurentia, include the Blue Ridge terrane, the Inner Piedmont terrane, and the Carolina terranes (Hatcher, 1972; Keller and Hatcher, 1999; Peavy et al., 2004; Mueller et al., 2013) (pink lines in Figure 1). These terranes were accreted onto Laurentia and subsequently experienced extensive ‘thin-skinned’ deformation (i.e. only affecting the upper crust and lithosphere) during the closing of the Iapetus Ocean and other oceanic basins from the Late Cambrian to the Early Silurian (Horton et al., 1989; Hibbard et al., 2002).

The Alleghanian Orogeny, which was the final stage of accretion onto the southeastern edge of cratonic Laurentia, began in the Carboniferous (Secor et al., 1986). This occurred during the later stages of the creation of the Pangaeon supercontinent (Heatherington and Mueller, 1999). The Suwannee terrane, which comprises modern-day southern Georgia, southern Alabama, and Florida, remained attached to North America when NW-SE oriented rifting and the subsequent opening of the Atlantic Ocean basins began separating landmasses during the Mesozoic (Mueller et al., 2013). This rifting also occurred onshore, creating the South Georgia Rift Basin, which consists of a series of normal fault-bounded basins that are up to 6-7 km thick (McBride et al., 1989). The Suwannee terrane has an affinity with Gondwana, specifically western Africa (e.g. Dallmeyer, 1987) and/or northern South America (e.g. Heatherington and Mueller, 1999). While the exact nature of the Suwannee Suture, whether it is collisional (e.g. Secor et al., 1986) or transpressional (e.g. Mueller et al., 2013), is not known for certain, the boundary itself is well-marked by both dipping reflectors (McBride and Nelson, 1988) and the BMA (McBride and Nelson, 1988; Heatherington and Mueller, 1999), both of which are attributed to an abrupt compositional difference representing the transition between cratonic Laurentia and the exotic accreted terranes.

3. Data and Methods

Teleseismic SK(K)S phases from two networks deployed in the southeastern US comprise the data set for our shear-wave splitting analysis (Figure 1). One of these networks is one that we were responsible for installing and maintaining, the SESAME

array (IRIS/Z9). This array was deployed from 7/2010 to 5/2014, and is mainly within the state of Georgia, but also extends into Florida, North Carolina, and Tennessee. The ~85 broadband stations of the SESAME array are arranged along three lines: the Western (W) line, the Diagonal (D) line, and the Eastern (E) line (Figure 1). The stations were organized in such a way so that the W and E lines were oriented normal to the hypothesized Suwannee Suture Zone, and the D line is oriented normal to the strike of most of the accreted terranes in the area. Station spacing within the suture zone is ~ 7 km; outside of the suture zone, the station spacing is ~25 km. We also used data recorded on selected stations from Transportable Array (IRIS/TA) that overlapped with the SESAME array in both time and space. The TA stations are deployed in a grid, with station spacing of ~70 km. We confined our data set to those seismic events that had epicentral distances greater than 90° and magnitudes larger than 5.8.

We imposed a number of quality control measures on the shear-wave splitting results of SK(K)S phases recorded on the SESAME array. To pass through the first round of culling, the measurements from our MATLAB shear-wave splitting GUI (e.g. Abt et al., 2009) had to have fast direction uncertainties less than or equal to 30° , and splitting time uncertainties less than or equal to 50% of the splitting time value. The second round of culling consisted of running the selected measurements from the first round through SplitLab (Wüstefeld et al., 2008). While the first GUI uses the eigenvalue minimization method of Silver and Chan (1991), SplitLab incorporates three methods of measuring shear-wave splitting: eigenvalue minimization, energy minimization (e.g. Silver and Chan, 1991), and rotation correlation (e.g. Bowman and Ando, 1987). Using the same shear-wave windows and culling criteria as those in the first round, we only select those

measurements in SplitLab that additionally have overlapping (within uncertainty range) shear-wave splitting results between the three shear-wave splitting methods, as well as highly linear corrected particle motions. Measurements that make it successfully through both rounds of culling are considered to be of the highest quality, and are the only SESAME shear-wave splitting results used in subsequent analyses (Figure 2).

To expedite the shear-wave splitting analysis on the TA stations, we used a modified version of the semi-automated shear-wave splitting code MFAST (Savage et al., 2010), which proceeds through the shear-wave splitting process with the S-pick (or, in our case, the SKS pick) as the only input. The MFAST codes are based on the Teanby et al. (2004) shear-wave splitting codes, which use the eigenvalue minimization methods of Silver and Chan (1991), but also incorporate cluster analysis, which searches over a number of shear-wave windows to identify the best-fitting shear-wave splitting measurement. The MFAST codes calculated 1077 SKS shear-wave splitting measurements and 833 null measurements for ~160 TA array stations from 1/2011-6/2013. Ideal shear-wave splitting measurements create a linear corrected particle motion and a flat transverse component (i.e. the shear-wave splitting measurement perfectly accounts for all of the anisotropy). To separate out the best measurements from the rest, we employed two criteria: the best measurements had corrected particle motion correlation coefficients (i.e. the fit of the data to a linear slope) of $r > 0.8$ and maximum amplitude ratio between the uncorrected and the corrected transverse components greater than 2. The rest of the measurements did not meet these criteria. Of our 1077 splits, 114 are the highest quality measurements. The highest quality nulls had uncorrected particle motion correlation coefficient values of $r > 0.8$ (i.e. their particle motions are already

linear because there is no splitting). Of the 833 nulls, 207 are the highest quality measurements.

4. Results

4.1 *SESAME Array*

The results of our manual teleseismic SK(K)S shear-wave splitting results for the SESAME array from 7/2010 to 10/2013 are displayed in Figures 2 and 3. With a total number of 236 measurements, most of the fast directions fall between NE-SW and NW-SW orientations, and have an average value of 64° East of North. Splitting times range from 0.39 to 3.23 s, with an average of 1.13 s.

A number of shear-wave splitting patterns emerge in this analysis. The first is that NE-SW fast directions dominate the trend throughout the entire SESAME array. These observations typically have splitting times of 1.0-1.5 s. Another, more subtle anisotropy pattern is most clearly evinced on the northern E line. Starting at the southernmost extent of the suture on the E line and moving northwards, an increasing number of fast directions are oriented at roughly $45\text{-}50^\circ$ West of North (blue measurements in Figure 2) with splitting times of about 1.5 s. Yet another trend is evident in the stations in the middle of the D line, as well as the W line stations near 34°N : these stations have a high number of null measurements that are accompanied by strongly oriented E-W measurements with smaller splitting times. Additionally, shear-wave splitting observations in the suture zone (the area outlined by the yellow dashed line) show high

complexity (Figures 2 and 3). Both the NW-SE and NE-SW fast direction patterns can be found here.

4.2 *Transportable Array*

The shear-wave splitting results for the TA array are plotted in Figures 4 and 5. If a station had any of the highest quality splits or nulls (blue lines or black crosses in Figures 4 and 5), then only those measurements were plotted. Otherwise, all of the remaining splits or nulls (red lines or grey crosses in Figures 4 and 5) are plotted at a given station. The highest quality measurements range from N-S to E-W, and have an average fast direction of 65° East of North. The splitting times for these measurements range from 0.5 to 4.0 s, and an average splitting time of 1.4 s. The remaining fast directions also vary from N-S to E-W, but they have an average value of 82° East of North. These splitting times range from 0.5 to 4.0 s and have an average splitting time of 2.14 s.

Some broad splitting patterns that emerge are the NE-SW fast direction orientations in the northeastern group of stations and the southern group of E-W fast directions. The complexity we see in the SESAME array exists in the same geographical area between longitudes of 82°W - 84°W and latitudes of 31°N - 36°N . These TA stations have a similar fast direction complexity to those in the SESAME array. The southernmost TA stations west of 84°W have fast directions that are primarily oriented E-W, but moving northwards, their complexity also increases, and some rotation towards NE-SW is observed.

4.3 *Backazimuthal Variation*

In an effort to explain some of these patterns, we look for backazimuthal variation within the fast directions of our measurements. These results, including the SESAME array measurements and the high quality TA measurements, are grouped by geographic area and are shown both as individual measurements and as binned weighted averages (Figures 6-11). Coherent backazimuthal variation with a periodicity of 90° may be an indication of the presence of two or more layers of anisotropy (e.g. Silver and Savage, 1994). Groups 1 and 2 (Figures 6 and 7) do not show obvious signs of a backazimuthal variation, while Group 3 and Groups 4, 5, and 6 (Figures 8-11) hint that such variations may exist.

The backazimuthal variation for Group 3 (Figure 8) shows a very rapid change in fast direction from backazimuths of 240° to 310° . In an attempt to model this kind of variation with multiple layers of anisotropy, we explored a range of possible anisotropy constraints with SynthSplit, which uses the particle motion perturbation method of Fischer et al., (2000). The pattern of red squares in Figure 8b represents the synthetic backazimuthal variation in fast direction with a three layer model: the uppermost layer has an olivine a-axis azimuth of 50° , a strength of anisotropy of 10%, and a layer thickness of 25 km. The middle layer has an olivine a-axis azimuth of 170° , a strength of anisotropy of 50%, and a layer thickness of 50 km. The deepest layer has an olivine a-axis azimuth of 60° , a strength of anisotropy of 50%, and a layer thickness of 150 km. This model does not quite replicate the observed behavior of the Group 3 values, but it is a good starting model for future iterations. Once the last of the SESAME data is retrieved

from the field, and its shear-wave splitting results are compiled, we plan to use the observed backazimuthal variation in splitting to solve for the best-fitting model parameters in each anisotropic layer; both two-layer and three-layer models will be considered.

The backazimuthal variation for Groups 4, 5, and 6 (Figure 11) shows a more gradual change in fast direction over the same backazimuthal range. The red squares in Figure 11b again represent modeling done in SynthSplit, only in this case, it is done with a two-layer model. This two-layer model has an upper layer with an olivine a-axis azimuth of 20° , a strength of anisotropy of 40%, and a layer thickness of 100 km, while the deeper layer has an olivine a-axis azimuth of 60° , a strength of anisotropy of 40%, and a layer thickness of 100 km. This model again does not exactly match the data, but we hope to hone in on better-fitting anisotropy model with the last six months of SESAME data from 2014.

5. Implications

We believe that there are three different anisotropic signals at work in this area of the southeastern US: (1) the alignment of material due to flow in the asthenosphere, (2) fossilized anisotropy in the lithosphere within the different accreted terranes, and (3) more recent alignment of fabrics within the lithosphere due to Mesozoic rifting in the South Georgia Rift Basin. The presence of fast direction variation both on a local level (Figures 2 and 3) as well as on a regional level (Figures 4 and 5) indicates the possibility

of variations in anisotropy both within the lithosphere and the asthenosphere, respectively.

Local lithospheric anisotropy fabric variations due to different accreted terranes are visible in the fast direction variation between the Blue Ridge (NE-SW measurements) and Inner Piedmont (E-W) terranes in Figure 2. The fact that the corresponding geographic groups (Groups 1 and 2, respectively) do not show signs of backazimuthal dependence (Figures 6 and 7) supports the idea that fossilized anisotropy fabrics in the lithospheric mantle of each terrane, while slightly different from each other, still have fast directions that are close to the fast direction in the asthenosphere.

The backazimuthal variation models in Figures 8 and 11 both have deeper layers that are representative of the asthenospheric mantle flow and upper layers that are representative of one or more layers of anisotropy within the lithosphere. For the backazimuthal dependence of Group 3 (blue measurements in Figure 2; Figure 8), our starting model has a NNW-SSE oriented fabric in the deeper lithospheric mantle that is sub-parallel to the direction of Mesozoic rifting, and an uppermost layer that is both reminiscent of the dominant fast directions observed over the area and that is consistent with compression in the mantle lithosphere during collision. A 50° fast direction in the uppermost lithosphere matches the orientation of azimuthal anisotropy beneath this region as constrained by Pn velocity anisotropy (MacDougall, Chapter 3). The starting model for the backazimuthal variation of Groups 4, 5, and 6 (Figure 11) also has a bottom layer that represents the asthenospheric mantle, but it instead contains a single lithospheric mantle layer that is oriented NNE-SSW (yellow measurements in Figure 2; Figure 11), also possibly parallel to the orientation of rifting.

The fast direction patterns observed within the SESAME array are duplicated to some extent in the regional analysis with the TA stations, a fact that is supported by the observation that TA data points add to the observed pattern of backazimuthal dependence in Figures 8 and 11. In Figures 4 and 5, Florida contains fast directions that are strongly aligned in the E-W direction, while South Carolina has a strong alignment of NE-SW fast directions. Otherwise, the fast directions exhibit a complexity that we also observe in the SESAME array. This scale of fast direction variation likely represents a combination of lithospheric anisotropy, for example between continental Laurentian and Suwannee lithospheric material, and/or localized drivers of asthenospheric flow, such as the presence of a cratonic keel (e.g. Fouch et al., 2000) or the remnant Farallon slab (e.g. Forte et al., 2007; Long et al., 2010) cannot be ruled out.

6. Conclusions

We have compiled the results of shear-wave splitting analyses from teleseismic SK(K)S phases recorded on the SESAME array from 7/2010-10/2013 and teleseismic SKS phases recorded on Transportable Array stations from 1/2011-6/2013. On the SESAME array, we observe distinct fast direction variations between the Blue Ridge and Inner Piedmont terranes, indicating that the lithosphere is likely contributing to the overall anisotropy measured at each station. However, the lack of backazimuthal variation in this northern sub-region suggests that lithospheric and asthenospheric fast directions are similar. Measurements within the Suwannee suture and the eastern part of the South Georgia Rift Basin show fast direction complexity that can be modeled using

multiple layers of anisotropy. On a regional scale, the fast direction orientations between Florida and South Carolina are oblique to one another, and may represent the difference between the accreted Suwannee terrane and cratonic Laurentian fossilized anisotropy fabrics, respectively. However, the presence of a cratonic keel or a remnant slab penetrating the asthenospheric mantle may also explain more variable asthenospheric flow that occurs on shorter length scales than that expected by purely plate-driven shear flow.

7. Acknowledgments

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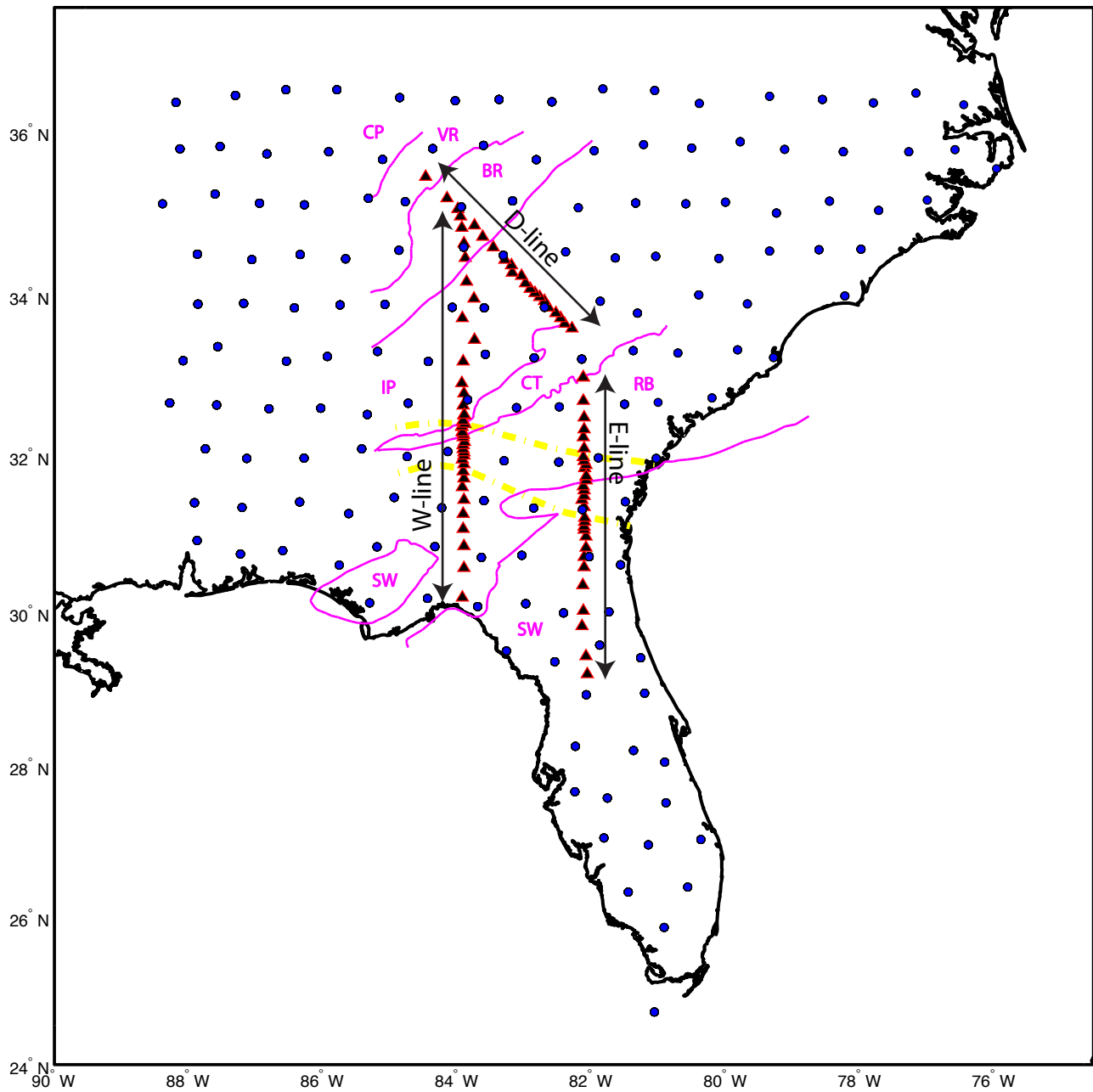


Figure 1. Station map. SESAME broadband stations are represented with red and black triangles, and are distributed along three lines: the Western line (W-line), the Diagonal line (D-line), and the Eastern line (E-line). TA stations used in this analysis are shown with blue circles. The Suwannee suture zone is denoted by the yellow dashed lines, as described in McBride and Nelson (1988). Magenta lines are terrane boundaries from Parker et al. (2013) and Mueller et al. (2013). CP: Cumberland Plateau. VR: Valley and Range terrane. BR: Blue Ridge terrane. IP: Inner Piedmont terrane. CT: Carolina terrane. RB: South Georgia Rift Basin. SW: Suwannee terrane.

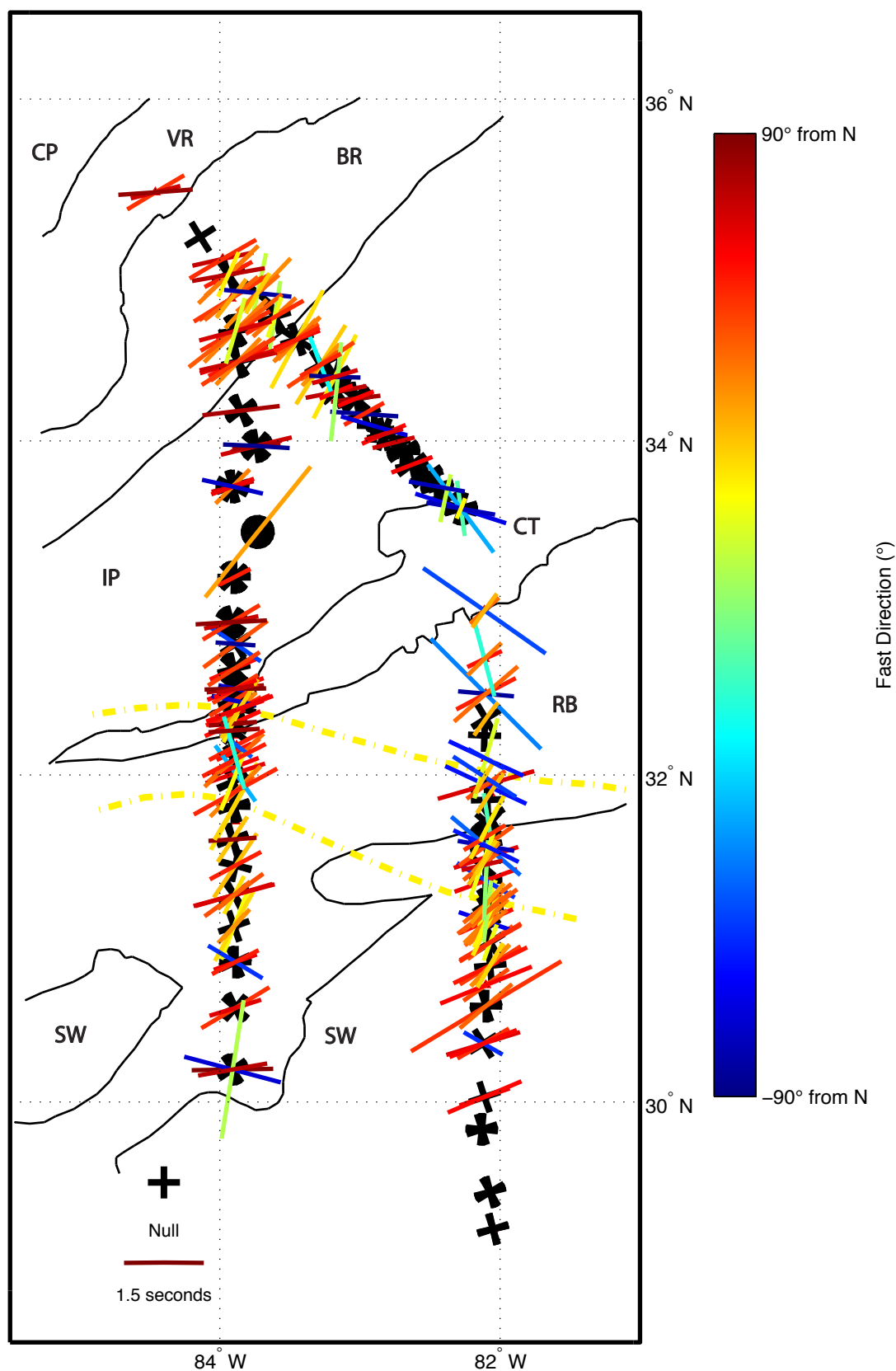


Figure 2. Teleseismic SK(K)S shear-wave splitting results. Colored lines represent the measurements; the orientation of the lines represent the fast direction (with respect to North), and the length of the lines scale with the splitting time. Black crosses represent null measurements, with the azimuth of the crosses plotted parallel and normal to the event backazimuth.

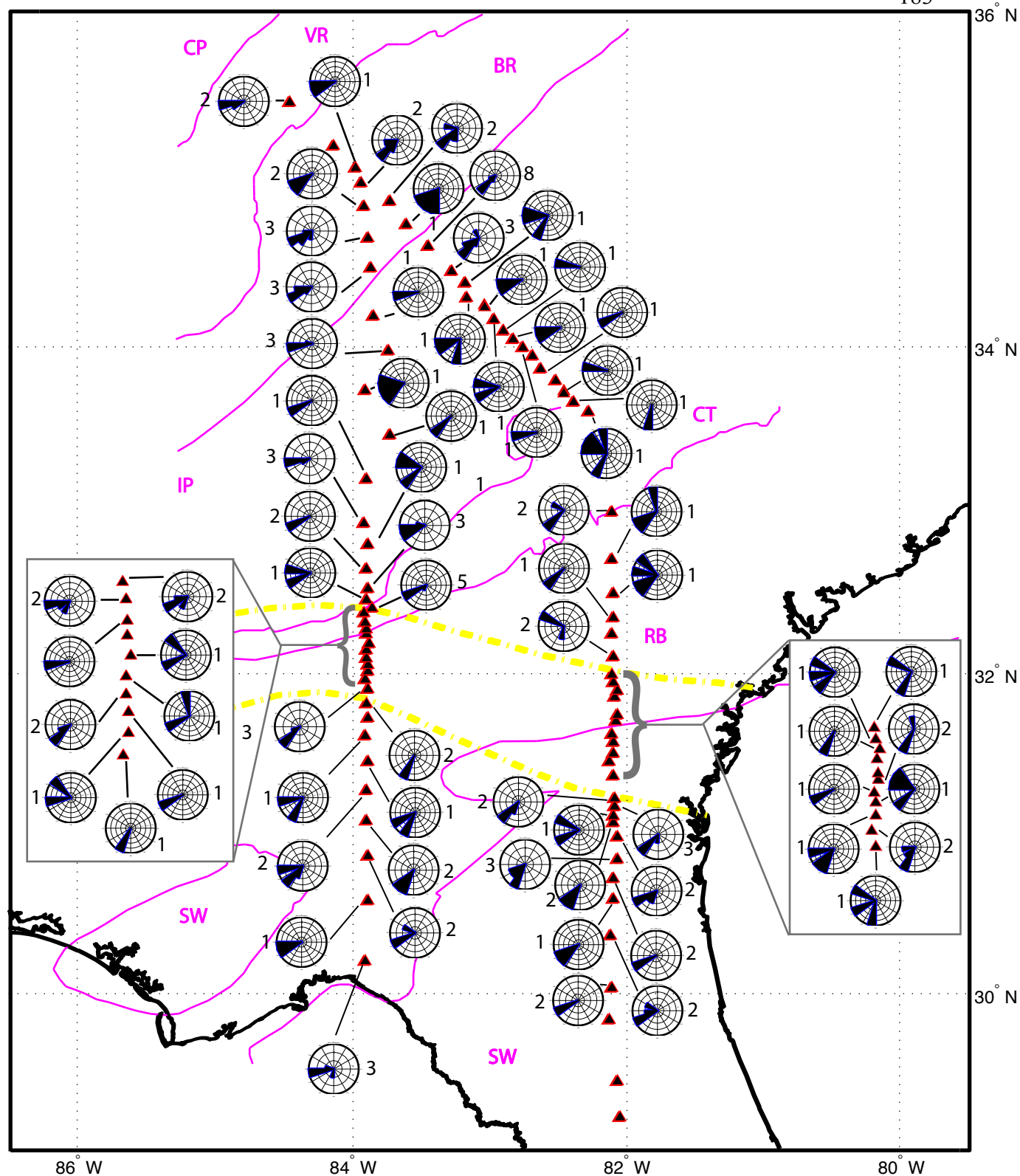


Figure 3. Teleseismic SK(K)S fast direction measurements plotted as rose diagrams. The number outside each rose diagram represents the maximum number of measurements in a fast direction bin.

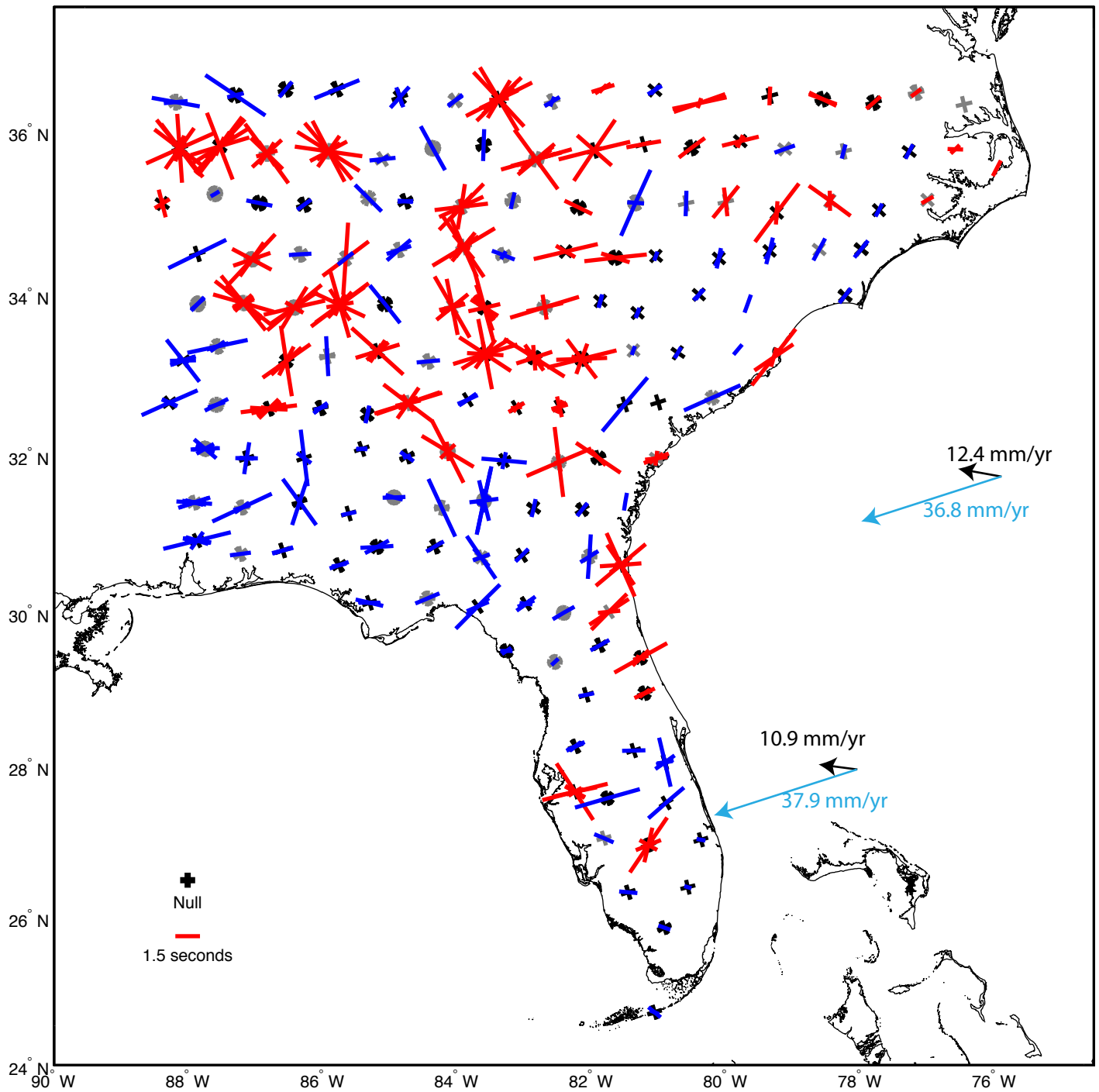


Figure 4. Teleseismic SKS fast direction measurements from the TA array (1/2011 - 6/2013). Either all of a station's best splits (blue lines) and nulls (black crosses) or all of a stations remaining (i.e. do not fit the 'best split' criteria) splits (red lines) or nulls (grey crosses) are shown. Arrows indicate the absolute plate motion vectors calculated in the "no net rotation" (black arrows) and "Pacific Hotspot" (blue arrows) reference frames.

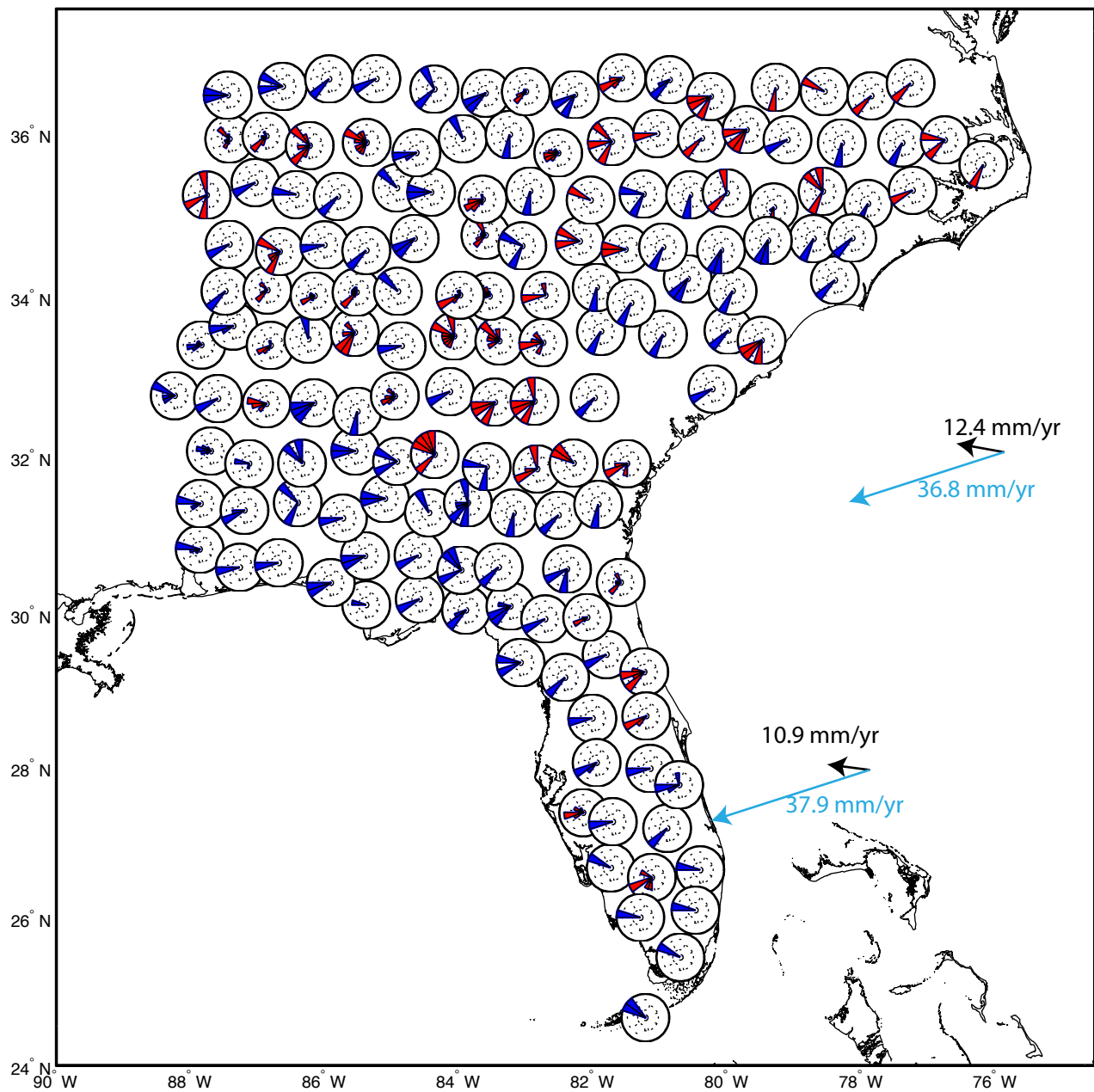


Figure 5. Teleseismic SKS fast direction measurements from the TA array (1/2011 - 6/2013) plotted as rose diagrams. Within each rose diagram, either all of a station's best (blue) measurements or all of a stations remaining (red) measurements are shown. Arrows indicate the absolute plate motion vectors calculated in the "no net rotation" (black arrows) and "Pacific Hotspot" (blue arrows) reference frames.

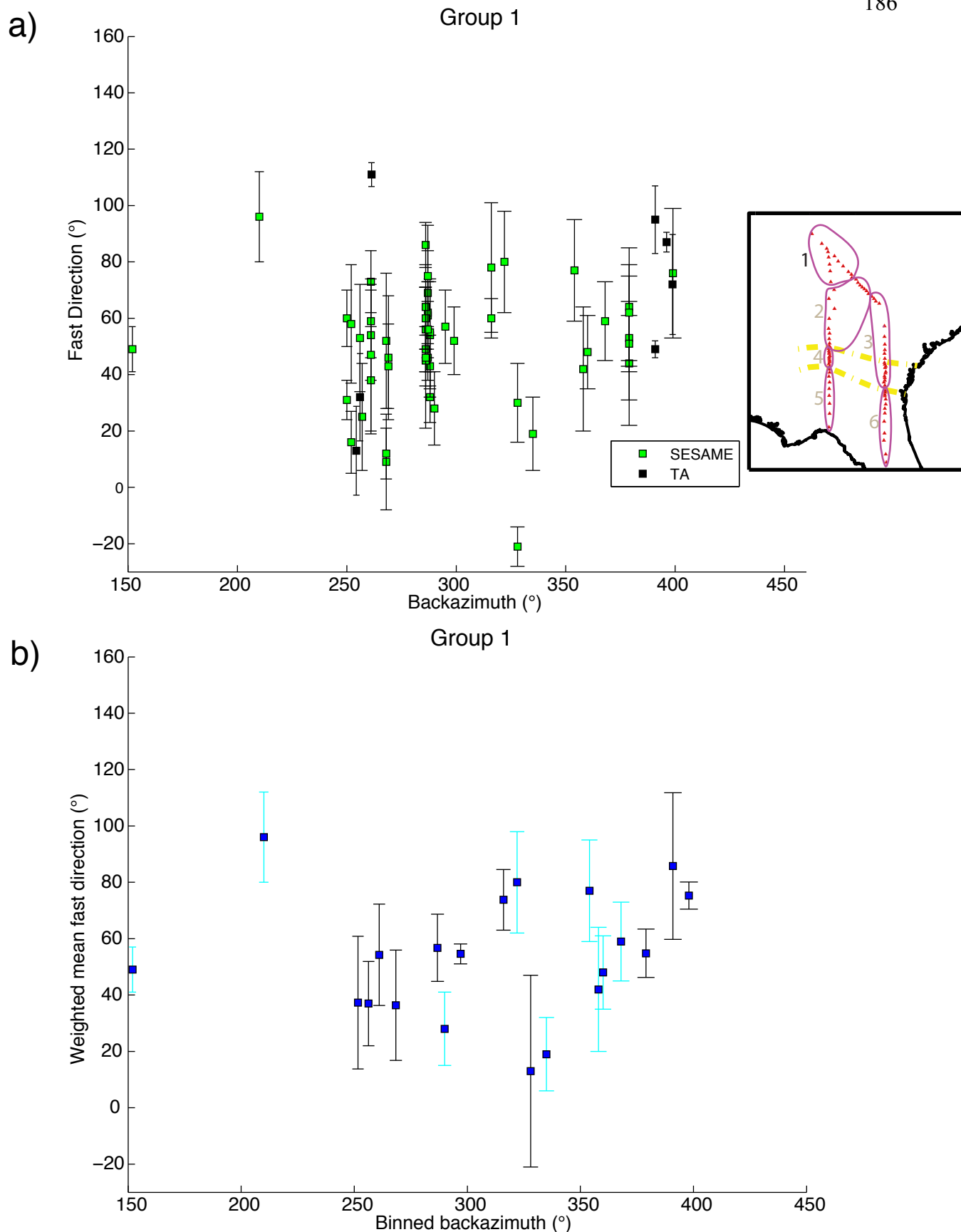


Figure 6. Backazimuth vs. fast direction for (a) raw and (b) binned data for Group 1. The geographic area represented by Group 1 is shown in the inset in (a). All of the SESAME measurements (green squares) and only the highest quality TA measurements in the same geographic area (black squares) are both shown in (a). (b) Both SESAME and TA measurements are binned in 5° backazimuth bins, and then the weighted mean and weighted standard deviation (errorbars) are calculated. If only one measurement is in a bin, then that measurement's fast direction uncertainty is plotted (cyan errorbars).

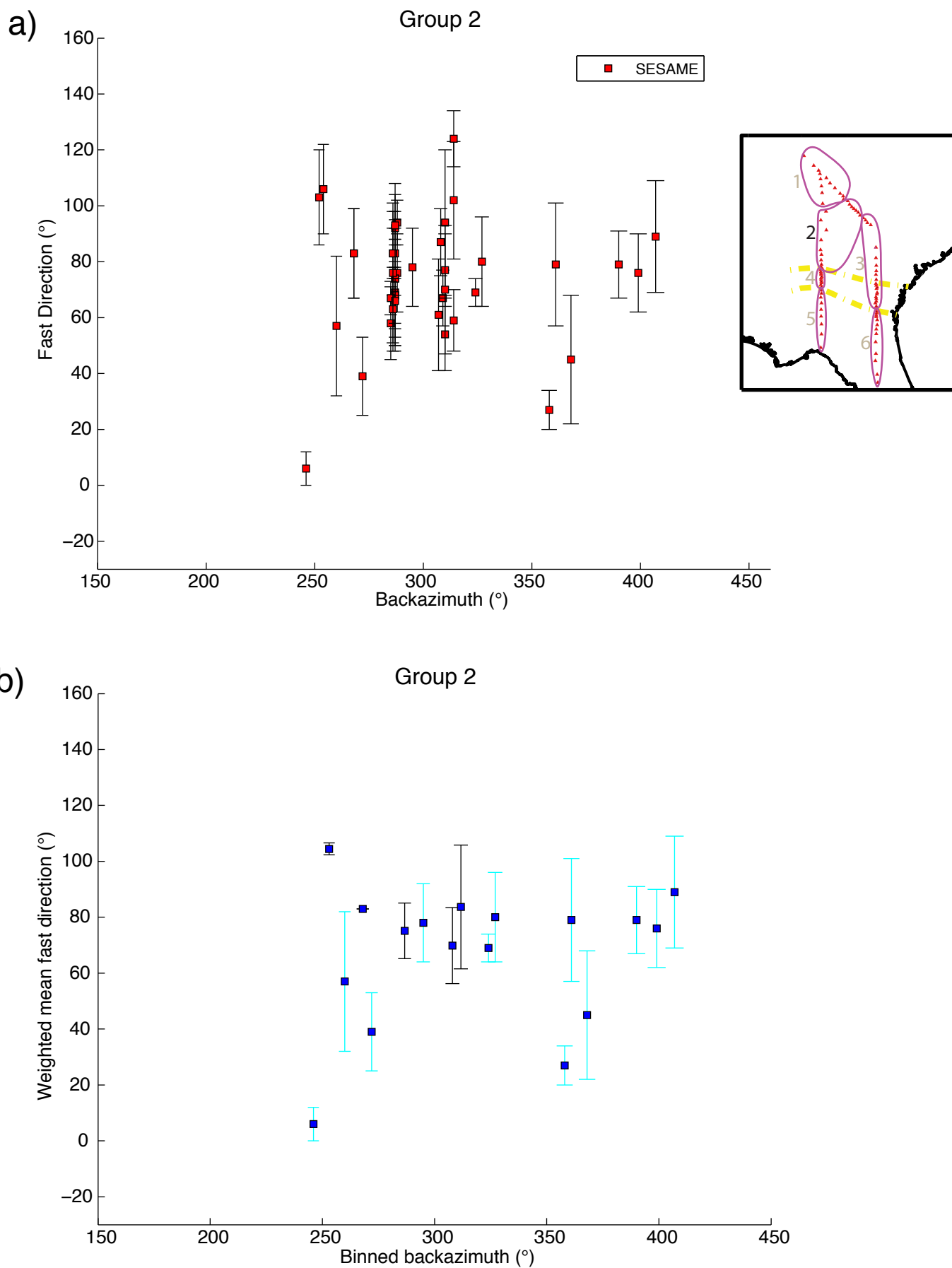


Figure 7. Backazimuth vs. fast direction for (a) raw and (b) binned data for Group 2.

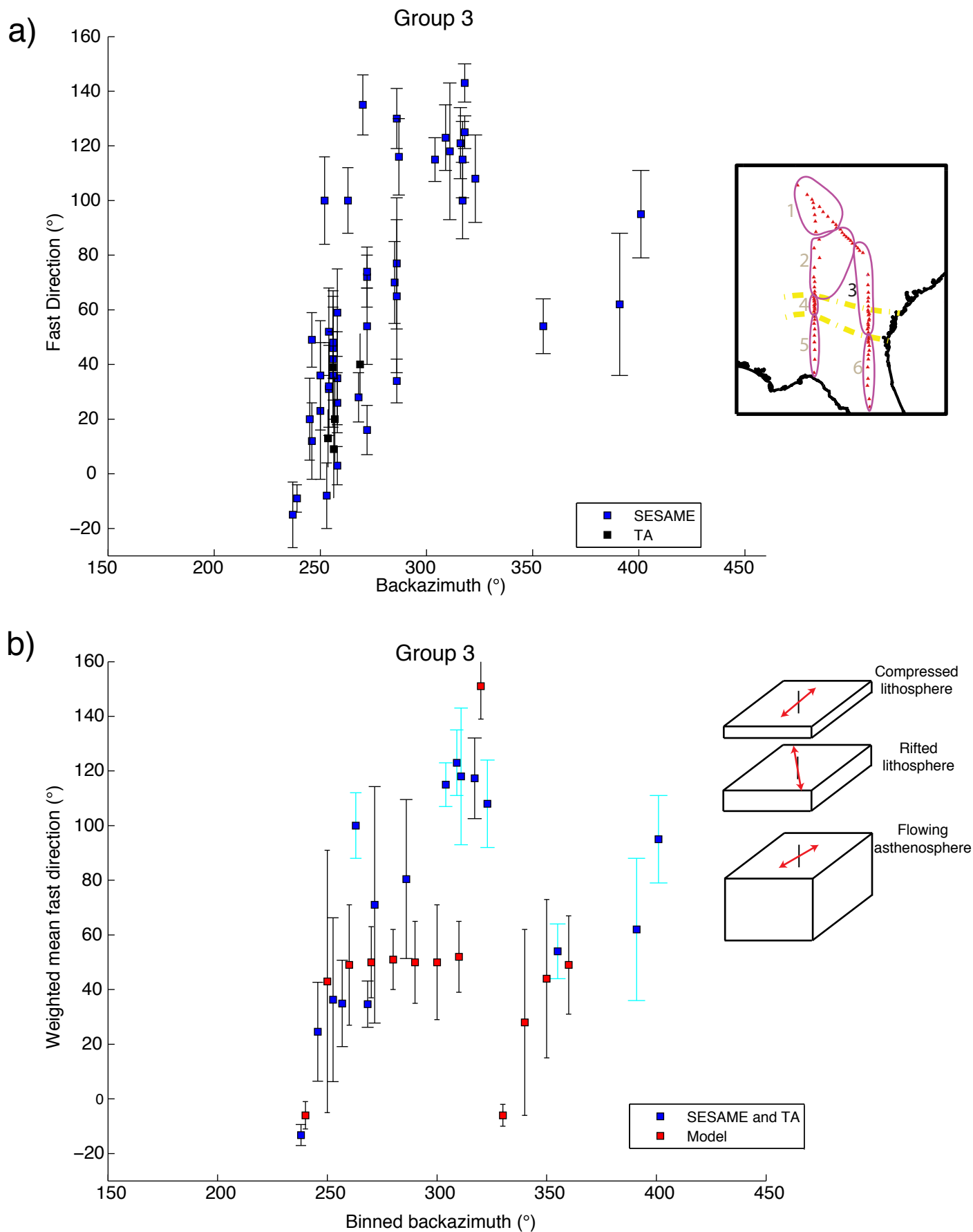


Figure 8. Backazimuth vs. fast direction for (a) raw and (b) binned data for Group 3. Red squares in (b) represent the model discussed in the text. A diagram of the model is shown in (b).

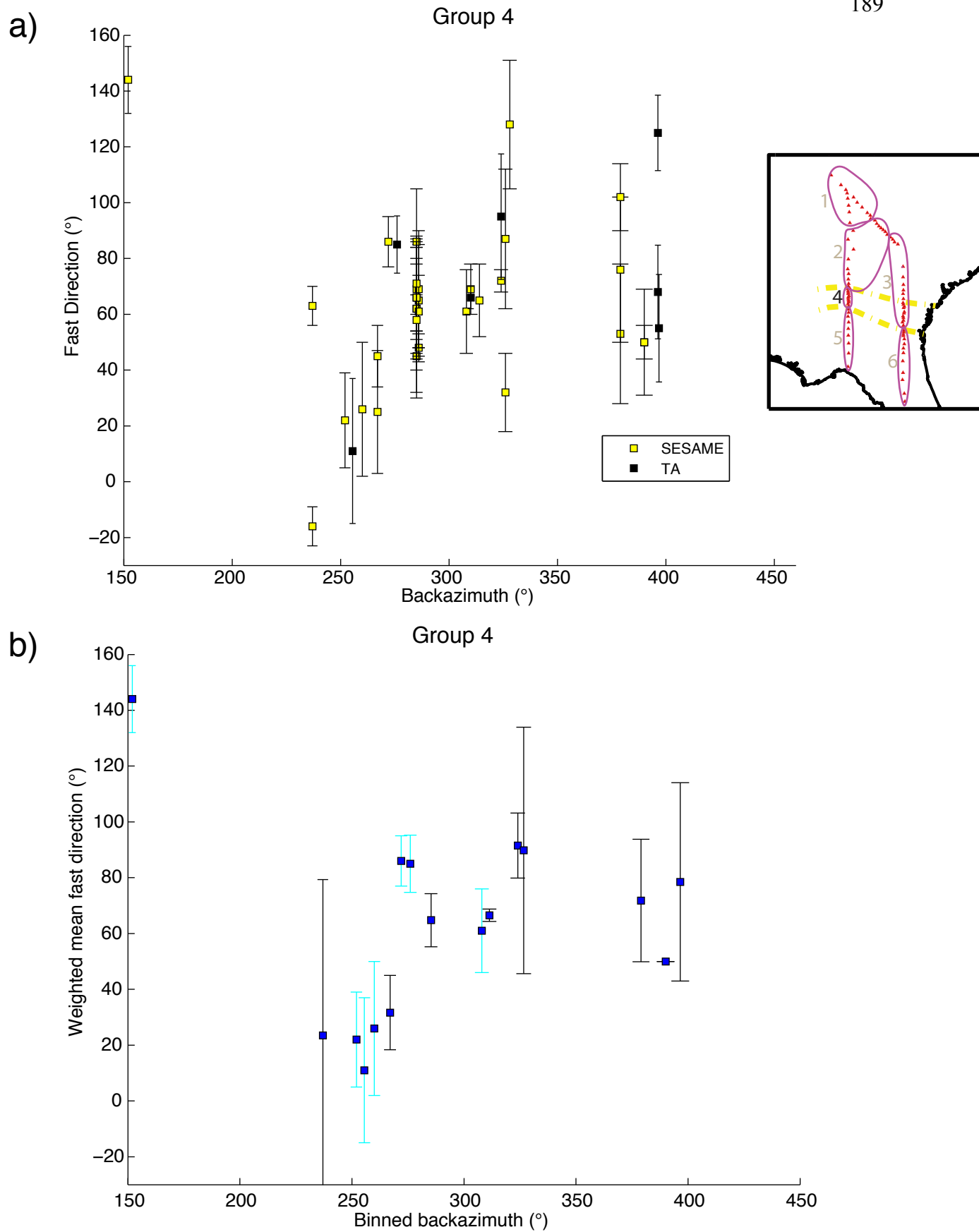


Figure 9. Backazimuth vs. fast direction for (a) raw and (b) binned data for Group 4.

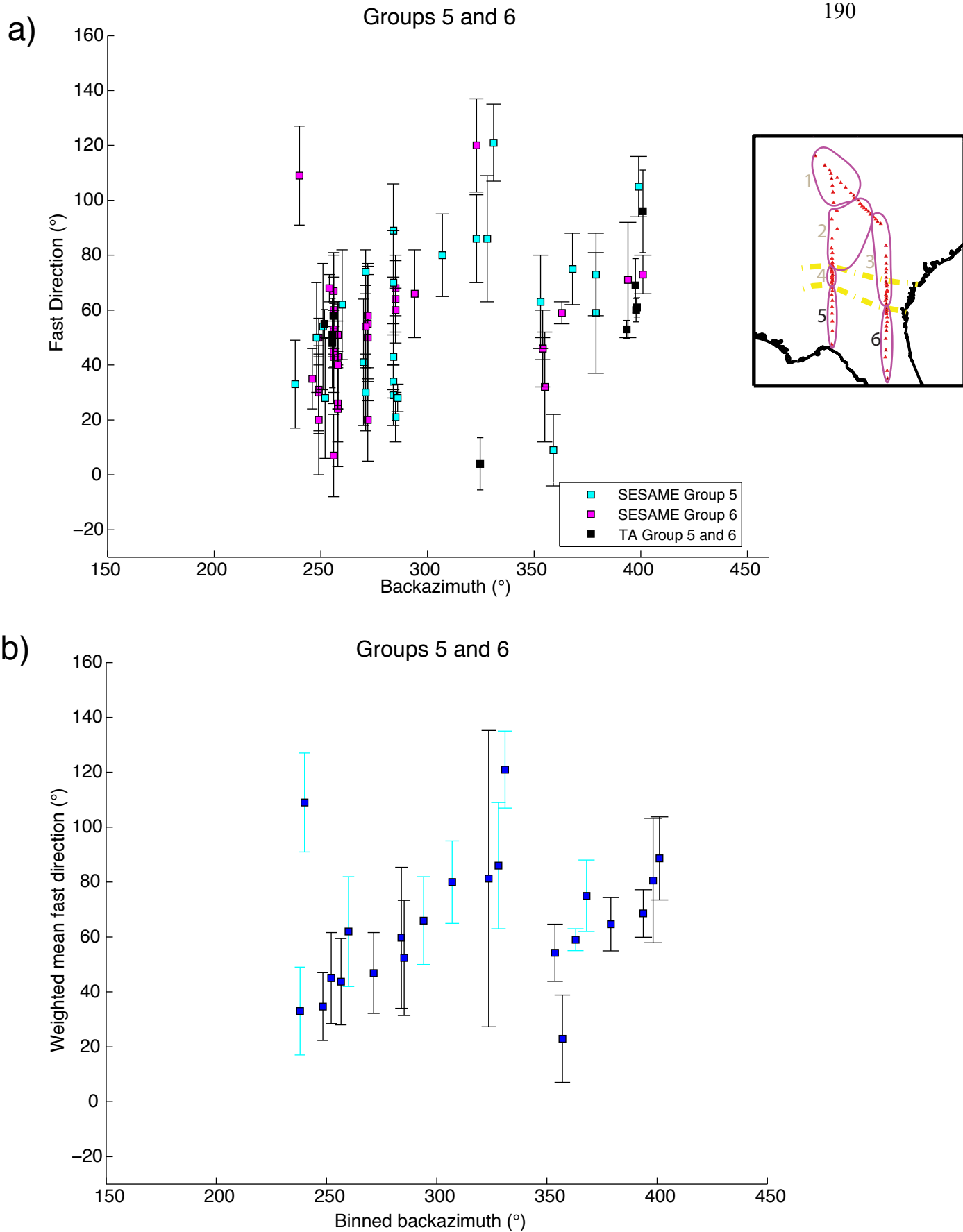


Figure 10. Backazimuth vs. fast direction for (a) raw and (b) binned data for Groups 5 and 6.

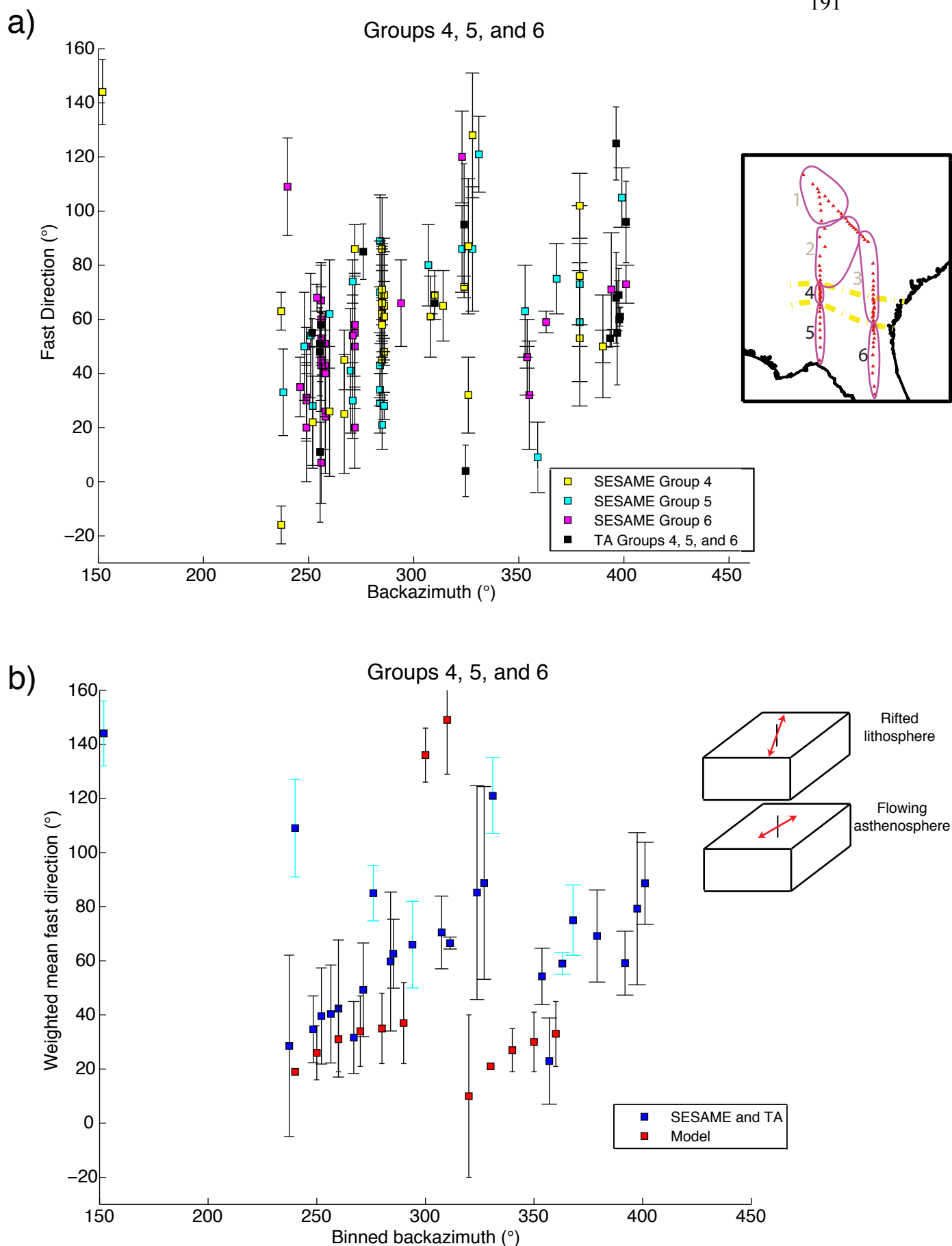


Figure 11. Backazimuth vs. fast direction for (a) raw and (b) binned data for Groups 4, 5, and 6. Red squares in (b) represent the model discussed in the text. A diagram of the model is shown in (b).

Chapter 5

Seismic anisotropy beneath the Hindu Kush-Pamir region

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Summary

With the goal of measuring seismic anisotropy and understanding mantle deformation at the western edge of the India-Eurasia collision zone, we measured shear-wave splitting in local S and teleseismic SKS/SKKS phases recorded by six stations of the Tajikistan National Seismic Network distributed in the Pamir Mountains and the Tadjik Basin which lies to their west.

S phases from local earthquakes produced splitting times of less than 0.5 s, whereas shear-wave splitting in SK(K)S phases ranges from 0.6 s to 2.7 s. Local S and SK(K)S fast polarizations recorded in the Pamirs are comparable and typically range from 45°- 86° E of N. In contrast, local S phases that sample the mantle beneath the Tadjik Basin produced fast directions of 85°-145° E of N, whereas SK(K)S fast polarizations at these stations are predominantly 50°-80° E of N, except for a group in the west with fast directions ranging from 0°-30° E of N. These shear-wave splitting observations were used to constrain vertically-layered models of mantle anisotropy for the Hindu Kush/Pamir sub-region and for the Tadjik Basin. A fundamental conclusion of this study is that significant anisotropy exists in the asthenosphere. The orientation of olivine *a*-axes in the asthenosphere beneath the Hindu Kush and Pamirs is consistent with shearing in the asthenosphere due to the absolute motion of the Eurasian plate, but beneath the Tadjik basin olivine *a*-axes in the asthenosphere are 20°-30° counter-clockwise from absolute plate motion, and may represent deflection of asthenospheric flow around the margins of the dipping lithospheric slab remnant beneath the Hindu Kush. Anisotropy in the shallow mantle beneath the Hindu Kush and Pamirs is consistent

with compression of the Eurasian lithosphere. Local S splitting beneath the Tadjik Basin suggests that either mantle deformation differs significantly from observed crustal compression, or the local S splitting in this sub-region is due to crustal anisotropy.

1. Introduction

In the broad region deformed by the India-Eurasia convergence, previous studies have debated whether seismic anisotropy, as measured by shear-wave splitting in teleseismic phases such as SKS, is located in the mantle lithosphere or asthenosphere (e.g. Flesch et al., 2005; Huang et al., 2011). The Hindu Kush and Pamir Mountains lie at the western syntaxis of the India-Eurasia collision zone (Figure 1). Beneath these mountains, dipping zones of seismicity, with hypocenters as deep as 280 km (e.g. Sippl et al., 2013), offer an unusual opportunity to resolve the distribution of mantle anisotropy with depth using shear-wave splitting in local S phases in combination with teleseismic phases.

The Hindu Kush and Pamir Mountains are themselves a manifestation of the dramatic lithospheric shortening that has occurred in this region and in the adjacent intermontane Tadjik Basin due to the continued collision of India northwards into the Eurasian continent. Deformational structures and earthquake source mechanisms are also consistent with significant shortening (Lukk et al., 1995; Bourgeois et al., 1997).

The deep seismicity that lies beneath the Hindu Kush and Pamir Mountains is surrounded by dipping zones of anomalously high velocity mantle, and these features have been interpreted as partially subducted lithospheric slabs that remained behind after the break-off of their deeper portions (Koulakov and Sobolev, 2006; Negredo et al., 2007). With the exception of these dipping slab features, the overall thickness of the Eurasian lithosphere in this region is ~ 150 km (Villaseñor et al., 2001; Hanna and Weeraratne, 2013).

2. Data and Shear-wave Splitting Method

Waveforms used in the shear-wave splitting analysis were recorded by six permanent stations of the Tajikistan National Seismic Network (CHGR, DUSH, GARM, IGRN, MANEM, and SHAA) (Figure 1). Well-resolved splitting measurements were obtained in S phases from 14 local events and SKS and SKKS phases (SK(K)S phases) from 73 teleseismic events. For local S phases, paths were limited to vertical incidence angles of 40° or less to minimize phase shifts at the surface or in the crust, and waveforms were filtered with a bandpass of 0.05-2 Hz. Teleseismic SK(K)S phases were from epicentral distances of 90° - 120° and were filtered with a bandpass of 0.05-0.35 Hz.

Shear-wave splitting fast polarization directions and splitting times were measured from the horizontal components of the shear-wave using the eigenvalue minimization method of Silver and Chan (1991). 95% confidence limits for the splitting parameters were calculated using an F-distribution, as described in Abt and Fischer

(2008). Phases that yielded fast polarization direction uncertainties of greater than 30° or splitting time uncertainties of more than 50% of the splitting time were discarded.

3. Shear-wave Splitting Results

For local S paths traveling from Hindu Kush events to station MANEM, located in the Pamir Mountains, most fast polarizations (ϕ) are 45° - 86° E of N (Figure 2a). For these paths, splitting times (dt) range from 0.17 s to 0.43 s (Figure 3a). All but one of the largest splitting times (> 0.33 s) come from events deeper than 180 km with longer paths to the station, consistent with the idea that these paths sample mantle anisotropy with a relatively uniform fast direction that persists to depths of more than ~ 150 km. Two events from deeper than 190 km yield anomalous splitting with fast polarizations of 163° - 170° and small splitting times (0.11 s and 0.19 s), suggesting that localized variations in anisotropy also occur. Most of the local S paths to MANEM travel roughly parallel to the strike of the Hindu Kush/Pamir seismic zone, although the paths from the three most northern events are more up-dip (Figures 2 and 4); all of these paths sample the dipping high velocity mantle that surrounds the seismic zone (e.g. Negredo et al., 2007) and the Eurasian lithosphere.

In contrast, local S phases to stations IGRN and CHGR sample the mantle beneath the Tadjik Basin (Figure 4). These phases produce fast polarizations at 85° - 145° E of N and splitting times of 0.21 s to 0.23 s. The corresponding events lie at a wide range of depths (45 km to 265 km). The minimal variation in splitting time indicates that if the fast direction and strength of anisotropy are uniform, the anisotropy resides at

relatively shallow depths. Alternatively, the deeper phases may encounter anisotropy with varying orientations that cause splitting from different depth ranges to cancel out.

Teleseismic SK(K)S splitting times range from 0.6 s to 2.7 s, with an average splitting time of 1.32 seconds; they are significantly larger than local S splitting times. Teleseismic fast polarizations are 50° - 80° E of N, except at the westernmost station in the Tadjik Basin (SHAA) where many fast polarizations are 0° - 30° E of N.

Clear differences exist between the splitting in the SK(K)S phases and the splitting in the local S phases. In all cases, the SK(K)S splitting times are significantly larger than the local S times, and at stations IGRN and CHGR, the local S fast polarizations have a distinctly different orientation than the SK(K)S fast polarizations. Not all of the local S phases lie close to an SK(K)S ray path. However, the volumes sampled by the local S and teleseismic SK(K)S phases overlap enough (Figure 4) that it is reasonable to conclude that the SK(K)S splitting indicates the presence of anisotropy at depths greater than those traversed by the local S phases.

4. Modeling the Vertical Distribution of Anisotropy

To quantify anisotropy in mantle structure beneath the region, we: 1) divided our splitting observations into two groups, one for the Hindu Kush/Pamir mantle (station MANEM) and one for the Tadjik Basin (stations CHGR, DUSH, GARM, IGRN, and SHAA) mantle, 2) used local S phase splitting to solve for average olivine-orthopyroxene anisotropy in the 50 km to 270 km depth range, 3) corrected SK(K)S splitting for the

effects of anisotropy in this upper layer, and 4) used the corrected SK(K)S splitting to solve for average anisotropy in the layer between 270 km and 410 km depth. We also experimented with solving for anisotropic parameters in two layers between 50 km and 270 km depth, but found that anisotropy parameters in the deeper layer were not well-resolved.

In each anisotropic model, the elastic constants are uniform and are dictated by a mix of 70% olivine (the averaged forsterite constants of Anderson and Isaak (1995) and Abramson et al., (1997)) and 30% orthopyroxene (the bronzite constants of Frisillo and Barsch (1972)). The fast axes of the olivine and orthopyroxene crystals, however, tend to align normal to one another; this slightly reduces the overall anisotropy observed in bulk rocks with this composition (Babuska and Cara, 1991). The anisotropy is described by two parameters: the olivine a -axis azimuth (θ), assuming that the a -axis is horizontal, and the strength of anisotropy (α), measured as the percent of the total anisotropy possible with the single crystal elastic coefficients.

The total depth range was divided into vertical sub-layers, each with a thickness of 10 km. The ray path for a given phase was discretized to provide path direction cosines in each 10 km sub-layer. We then solved the Christoffel equation in each layer for the given path to obtain the incremental splitting in each sub-layer, and integrated the splitting along the path in the particle motion of a synthetic shear wave. In essence, the synthetic shear wave is rotated and split in each sub-layer until it reaches the surface,

where we calculated synthetic shear-wave splitting measurements using the same splitting analysis method that was applied to the observations.

The best-fit anisotropy models, with parameters θ and α , are those that minimize the summed misfits between synthetic and observed shear-wave splitting measurements for the same path. Model parameters were tested with a 10° increment in θ and a 10% increment in α . To determine if any other models fell within the 95% confidence interval of a best-fit model, we multiplied the misfit of the best-fit model by an F-test confidence limit; any models with misfits that were less than or equal to that value are considered to be acceptable.

For the Hindu Kush-Pamir mantle, the best-fitting model for the 50-270 km depth range, based on the local S phases at MANEM, has an olivine a -axis orientation (θ) of 40° and a strength of anisotropy (α) that is 10% of the single crystal values (Table 1, Figure 1). Since no other model fell within the 95% confidence limit of these parameters, their uncertainties can be considered to be $\leq 10^\circ$ and $\leq 10\%$, respectively. The anisotropy in the deeper mantle layer for this sub-region has a more easterly θ value of $80^\circ \pm 10^\circ$; the strength of anisotropy in deeper mantle layer is also greater, $80\% \pm 10\%$ of the single crystal values, although if anisotropy was allowed at depths greater than 410 km, this value would be lower. For the mantle beneath the Tadjik Basin, the best-fitting anisotropy in the 50-270 km layer has a θ value of $130^\circ \pm 10^\circ$ and an α value that is 10% of the single crystal values; θ in the deeper layer is $50^\circ \pm 10^\circ$ and α is $50\% \pm 10\%$.

We did not include the upper 50 km in our anisotropy models because our data are too sparse to robustly constrain anisotropy in the crust relative to anisotropy the mantle. In any case, crustal anisotropy would relate to aligned cracks, faults, melt/fluid pockets, or to mineral fabrics in crustal rocks, as opposed to the olivine-orthopyroxene anisotropy that we assumed for the mantle. If crustal anisotropy did exist with a fast axis that is roughly parallel to the olivine a -axes in the shallow mantle layer, as suggested by one local S splitting measurement at MANEM from a depth of 54 km and one at IGRN from a depth of 45 km, the best-fitting strength of anisotropy for the shallow mantle layer would be even weaker than 10%. In fact, given that the local S splitting times at IGRN and CHGR do not increase with path length below the crust, all of the local S shear-wave splitting at these stations could be explained by crustal anisotropy. However, attributing all of the local S splitting at IGRN and CHGR to the crust would not significantly alter the estimate of the anisotropy required at these stations in the mantle layer deeper than 270 km, assuming that the paths from the deepest local events to IGRN sample mantle in the 50-270 km layer that is comparable to that sampled by nearby SK(K)S paths (e.g. Figure 4a).

Vertical variations in azimuthal anisotropy can create a dependence of shear-wave splitting on phase back-azimuth. To investigate whether backazimuthal patterns shed additional light on the vertical distribution of anisotropy in our study region, we analyzed both the local S and SK(K)S fast directions for any backazimuthal dependence. However, we did not observe a clear backazimuthal dependence in either the SK(K)S phases or the local S phases.

5. Implications for Mantle Deformation

A significant implication of these results – both the differences in shear-wave splitting times between local S and SK(K)S and the modeling of average anisotropy beneath the stations – is that much of the SK(K)S splitting is due to anisotropy at sub-lithospheric depths (> 150 km; e.g. Hanna and Weeraratne, 2013; Villaseñor et al., 2001). This finding contrasts many previous interpretations of teleseismic shear-wave splitting measurements in the broad region of India-Eurasia convergence that attribute much of the observed splitting to deformation in the lithosphere (e.g. Flesch et al., 2005). Rather, our results are consistent with interpretations that explain observed teleseismic splitting by flow within the asthenosphere (e.g. Huang et al., 2011).

Overall, the SK(K)S shear-wave splitting and the deeper layer of anisotropy in this study are broadly consistent with the orientation of asthenospheric shear that would be predicted by the absolute motion of the Eurasian plate. In a no-net-rotation reference frame (Kreemer et al. 2003), the Eurasian plate is moving at 2.3 cm/yr at an azimuth of $\sim 83^\circ$ with respect to the deeper mantle, and in a Pacific hotspot reference frame it is moving in the opposite direction, $\sim 267^\circ$, at 3.0 cm/yr (Gripp and Gordon, 2002). Beneath the Hindu-Kush/Pamir sub-region, the best-fitting olivine *a*-axis orientation in the deeper layer of anisotropy ($80^\circ \pm 10^\circ$) is parallel to these plate motions. Beneath the Tadjik Basin, the deeper layer *a*-axis ($50^\circ \pm 10^\circ$) is reasonably close to the direction of absolute plate motion, but rotated 20° - 30° counter-clockwise, in part due to the zone of SK(K)S fast polarizations at station SHAA that have orientations closer to 30° . This mis-match with apparent absolute plate motion could reflect local flow deflection around the western

margin of the dipping lithosphere associated with the Hindu Kush seismicity, or other flow deviations due to variations in lithospheric thickness (e.g. Fouch et al., 2000; Miller and Becker, 2012) and far-field buoyancy (e.g. Conrad et al., 2007).

The olivine *a*-axis orientation in the shallower mantle layer beneath the Hindu Kush and Pamir Mountains ($40^\circ \pm 10^\circ$) is approximately parallel to the strike of the mountains. This anisotropy likely reflects a combination of olivine alignment in the dipping lithosphere and compression in the Eurasian plate due to its collision with India.

Shallow anisotropy beneath the Tadjik Basin is more enigmatic. Here the best-fitting olivine *a*-axis in the shallow mantle layer ($130^\circ \pm 10^\circ$) lie approximately parallel to the direction of compression in the Tadjik Basin inferred from structural geology and paleomagnetic data (Bourgeois et al., 1997) and earthquake source mechanisms (Lukk et al., 1995). If the shallow anisotropy actually represents mantle deformation, the direction of compression that it would imply would be normal to the compression inferred at crustal levels. Alternatively, because the local S splitting times at Tadjik Basin stations do not increase with phase path length below the crust, all the anisotropy that they reflect could be located within the crust. In this case, their fast polarizations would be consistent with crustal fractures that align parallel to the direction of inferred crustal compression.

6. Conclusions

Shear-wave splitting in local S phases from earthquakes in the Hindu-Kush/Pamir seismic zone to Tajikistan National Seismic Network stations in the Pamir Mountains and

Tadjik Basin is characterized by splitting times of less than 0.5 s. In contrast, shear-wave splitting in SK(K)S phases recorded by these stations is much larger (0.6 s to 2.7 s). Given that many of the local events lie at depths of more than 150 km, this result alone indicates the presence of significant anisotropy in the asthenosphere beneath this region.

Local S and SK(K)S fast polarizations at station MANEM in the Pamirs are comparable and mostly lie at 45°- 86° E of N. In contrast, local S phases that sample the mantle beneath the Tadjik Basin produced fast directions of 85°-145° E of N, and SK(K)S fast polarizations at these stations are predominantly 50°-80° E of N, except for a group in the west at 0°-30° E of N.

Both the shear-wave splitting measurements and the models of average layers of anisotropy that they constrain for the mantle beneath Hindu Kush/Pamir region and for the Tadjik Basin demonstrate that much of the SK(K)S splitting is produced at asthenospheric depths. The orientation of olivine *a*-axes in the asthenosphere beneath the Hindu Kush and Pamirs is ~80° E of N. This alignment is parallel to the direction of absolute Eurasian plate motion in either the Pacific hotspot or no-net-rotation reference frames, and is consistent with simple models of shearing in the deeper asthenosphere due to plate motion. Olivine *a*-axes in the asthenosphere beneath the Tadjik Basin are ~50° E of N, and suggest deflection of asthenospheric flow around the margins of the dipping lithospheric slab remnant beneath the Hindu Kush.

Anisotropy in the shallow mantle beneath the Hindu Kush and Pamirs is consistent with compression of the Eurasian lithosphere. Local S splitting beneath the Tadjik Basin is not consistent with mantle deformation parallel to the direction of crustal compression; they indicate that either mantle deformation differs significantly from that in the crust, or the local S splitting in this sub-region is due to crustal anisotropy.

7. Acknowledgements

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9. List of Figures and Tables

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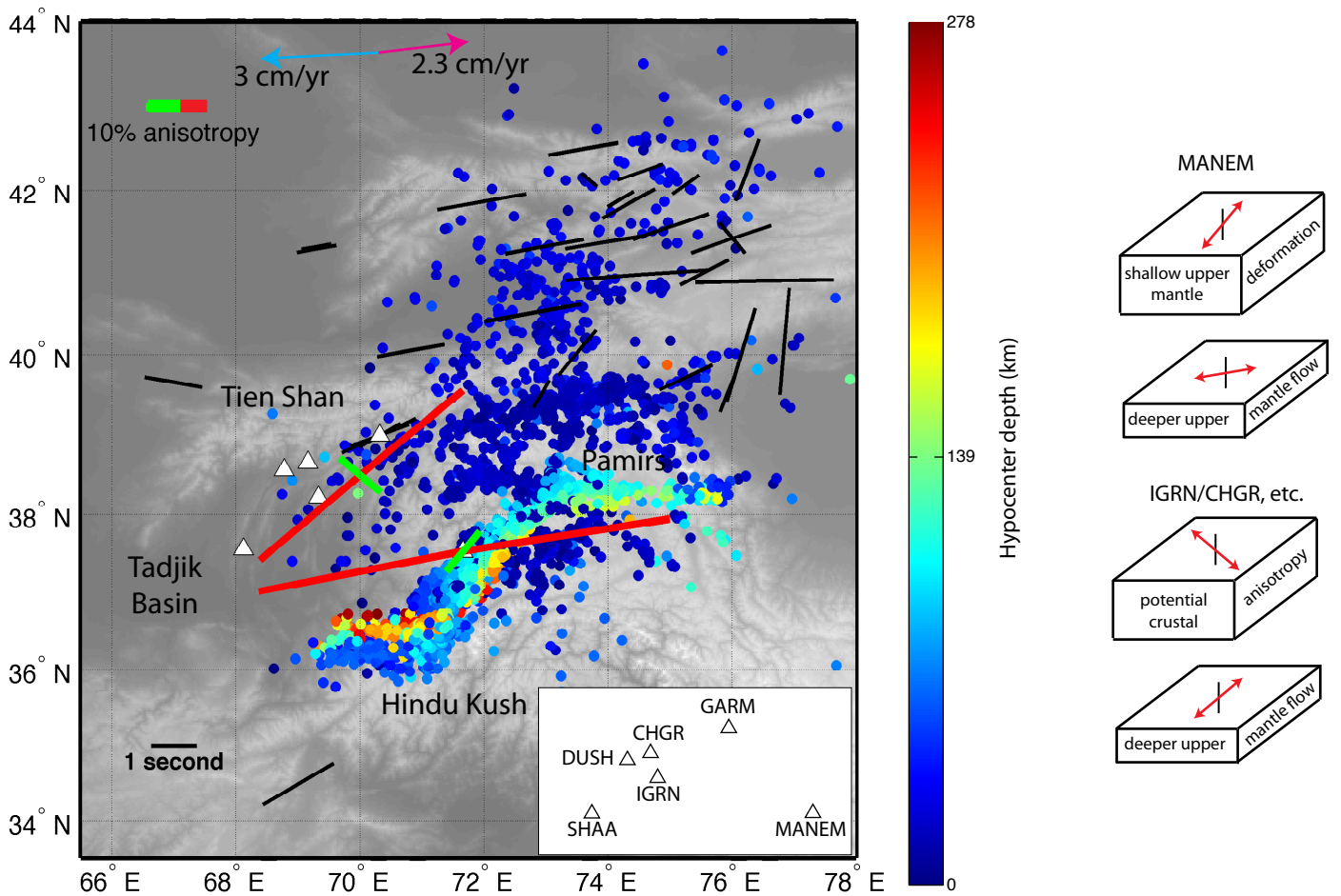


Figure 1. Map of study region. Seismic stations are represented by white triangles (also shown in inset). Seismicity (from Sippl et al., 2013) is displayed in colored dots, with hypocenter depths scaled by color. Black lines represent previously published teleseismic $^{*}KS$ shear-wave splitting measurements from Makeyeva et al. (1992), Vinnik et al. (1992), Helffrich et al. (1994), Wolfe and Vernon (1998), Li and Chen (2006), and Huang et al. (2011); the splitting times are scaled by the scale bar in the bottom left hand corner of the map. Red and green lines represent our anisotropy models for MANEM and the IGRN/CHGR/DUSH/GARM/SHAA station groups. Green lines represent shallower anisotropy constrained by local S phases, and red lines represent deeper anisotropy constrained by SK(K)S phases corrected for shallow anisotropy. The strengths of these models are scaled by the scale bar in the upper left hand corner of the map. Diagrams of our anisotropy models are shown on the right. The model shown for IGRN/CHGR etc. assumes that all of the anisotropy sampled by the local S phases resides in the crust. However, more complex lithospheric anisotropy is also possible. Pink and blue arrows at the top of the map represent the absolute plate motion of the Eurasian plate in the “no-net-rotation” and “Pacific hotspot” reference frames, respectively (Gripp and Gordon, 2002).

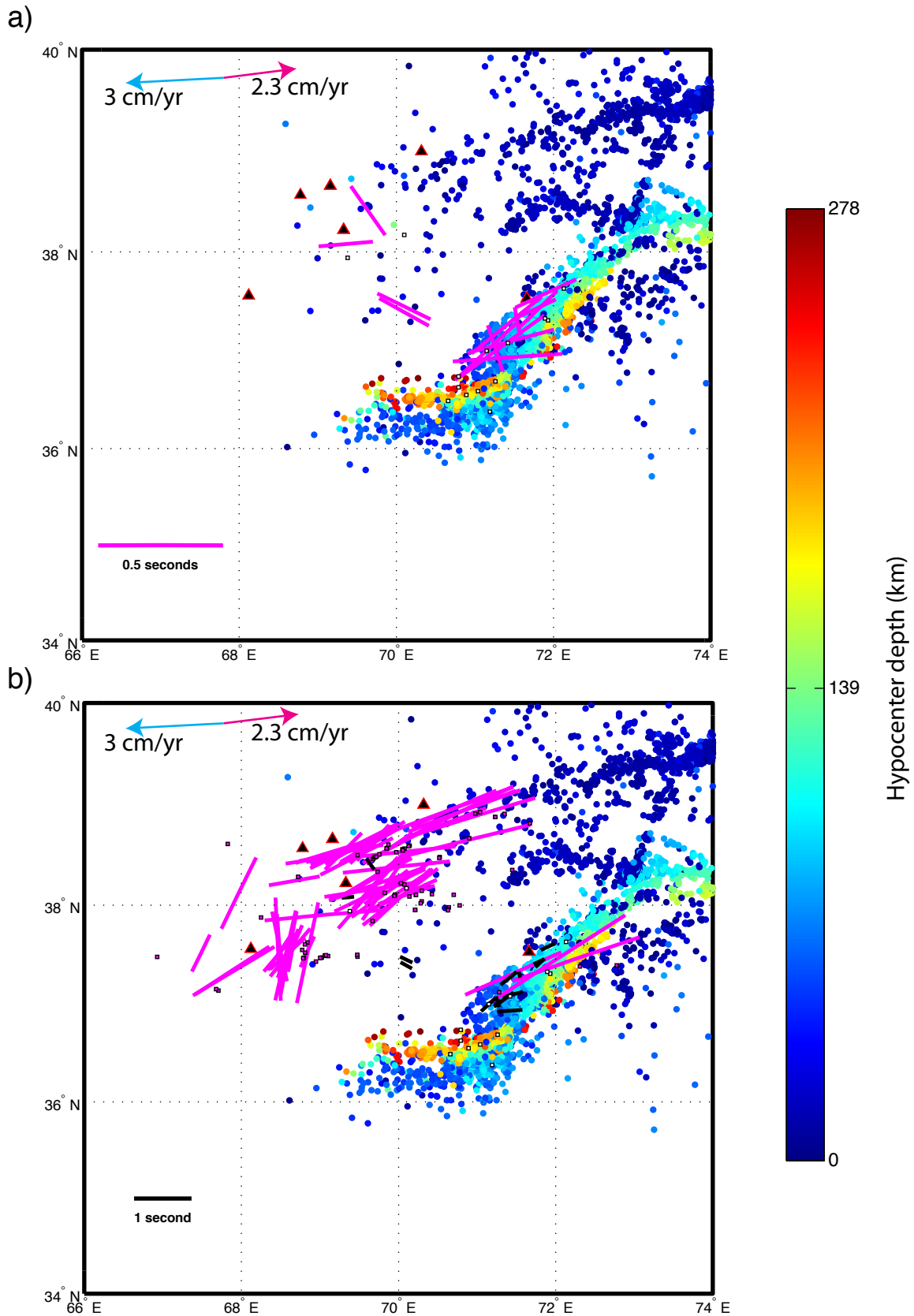


Figure 2. Shear-wave splitting results for the local S and teleseismic SK(K)S events. The orientation of each pink line, with respect to North, is the fast polarization direction (ϕ), and the length of the line, as indicated by the scale bar, represents the splitting time (dt). (a) Local S splitting results (pink lines) plotted at their path midpoints. White squares represent the hypocenters of the corresponding earthquakes. (b) Teleseismic SK(K)S results (pink lines); local S splitting results from (a) are shown to scale as black lines. Pink squares represent the pierce point of the teleseismic ray path at 410 km, and the corresponding splitting measurement is plotted halfway between this point and the station. Local seismicity from Sippl et al. (2013) is shown by the color-coded circles, with hypocenter depths indicated by the colorbar on the right. Pink and blue arrows at the top of the map represent the absolute plate motion of the Eurasian plate in the “no-net-rotation” and “Pacific hotspot” reference frames, respectively (Gripp and Gordon, 2002).

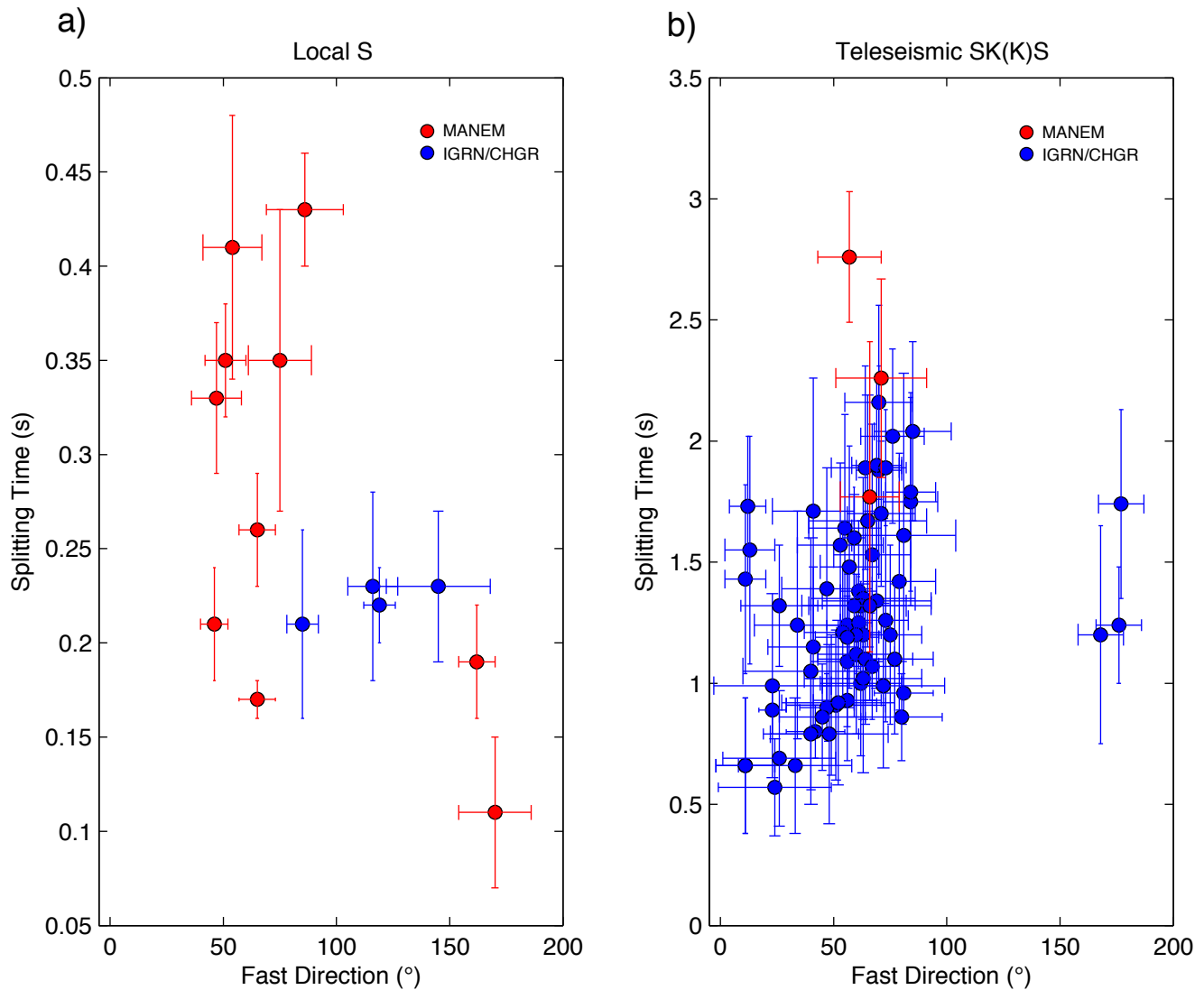
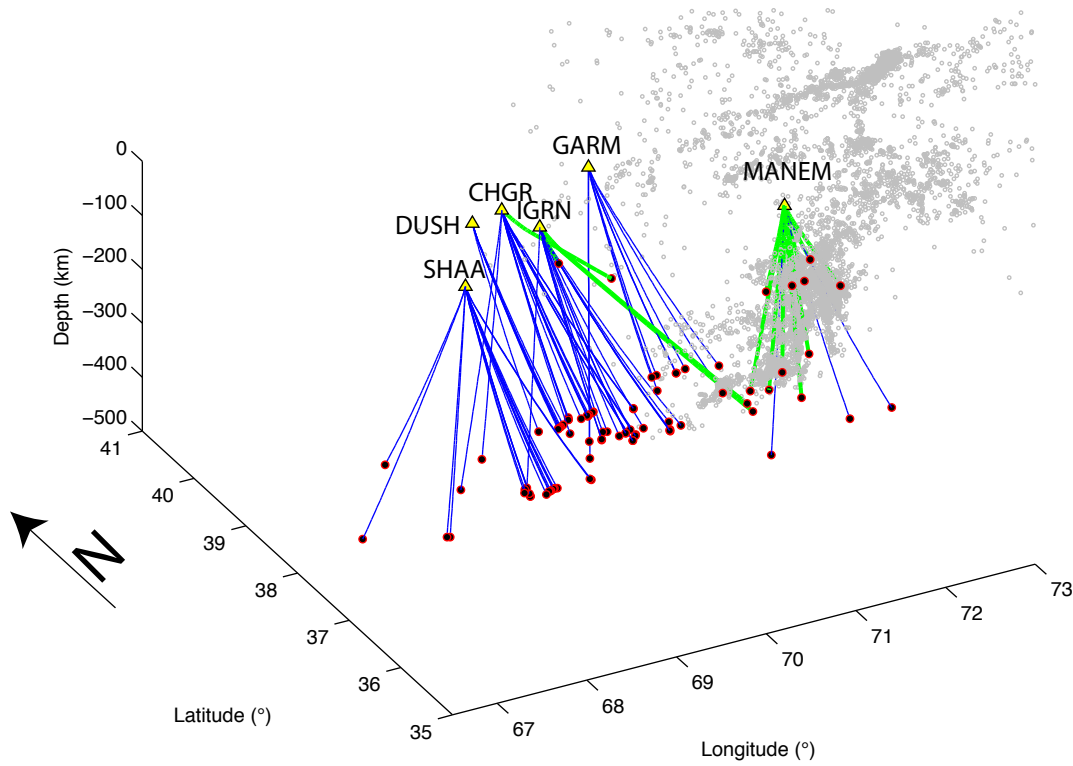


Figure 3. Shear-wave splitting fast polarization direction and splitting time results for (a) local S and (b) teleseismic SK(K)S events. Here, ‘MANEM’ represents all of the local and teleseismic results measured at station MANEM. ‘IGRN/CHGR’ represents the results from all events measured at stations that are not MANEM, which for the local events, include IGRN and CHGR, and for the teleseismic events, includes CHGR, DUSH, GARM, IGRN, and SHAA.

a)



b)

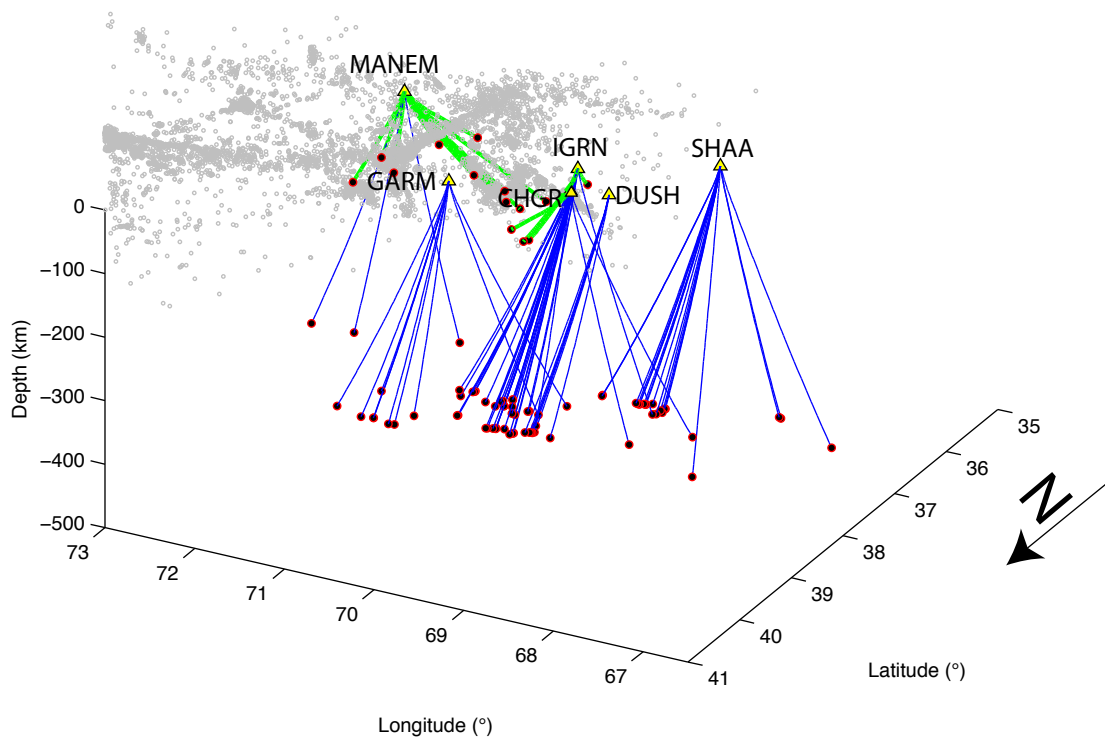


Figure 4. Ray path for phases with well-constrained shear-wave splitting measurements. The paths of the local S (green lines) and teleseismic SK(K)S (blue lines) have red dots at the hypocenter depths and 410 km pierce points, respectively. Grey dots represent the seismicity of Sippl et al. (2013).

Station	Model Group #	Phase	Type	Best-fit θ ($^{\circ}$)	Best-fit α (%)
MANEM	1	local S	Shallow mantle	40 $\pm$ 10	10 $\pm$ 10
MANEM	1	corrected SK(K)S	Deeper mantle	80 $\pm$ 10	80 $\pm$ 10
IGRN/CHGR	2	local S	Shallow mantle	130 $\pm$ 10	10 $\pm$ 10
IGRN/CHGR	2	corrected SK(K)S	Deeper mantle	50 $\pm$ 10	50 $\pm$ 10

Table 1. Modeling results. Station ‘MANEM’ represents all shear-wave splitting measurements recorded at MANEM. Station ‘IGRN/CHGR’ represents all shear-wave splitting measurements recorded at stations other than MANEM, which, for local S phases, is IGRN and CHGR, and for teleseismic SK(K)S events, are CHGR, DUSH, GARM, IGRN, and SHAA. “Corrected SK(K)S” phases indicate that the teleseismic SK(K)S phases recorded at these stations were back-calculated along a ray path through the lithosphere, while accounting for the best-fit anisotropy model for depths shallower than 270 km. These corrected SK(K)S waveforms were then reanalyzed for shear-wave splitting that accrued at depths greater than 270 km. “Shallow mantle” represents anisotropy accrued from depths of 270 km to 50 km. “Deeper mantle” represents anisotropy accrued from depths of 410 km to 270 km. θ represents the olivine a-axis orientation, with respect to North. α represents the strength of anisotropy. Best-fit models are those that have the total minimum misfit between the synthetic and recorded shear-wave splitting results. Diagrams of these model results are shown on the right in Figure 1.