

# Shear wave birefringence in wedge-shaped anisotropic regions

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## SUMMARY

A popular method for studies of seismic wave speed anisotropy within the Earth is based on observations of birefringence in subvertically propagating shear waves. A well-recognized shortcoming of the method is that individual birefringence measurements, commonly interpreted as a path-average of anisotropy, do not reveal the potential complexity of anisotropic structure. Consequently, there is a necessary ambiguity in the interpretation of observed birefringence parameters (fast polarization directions and delays between components). In this paper, we explore the influence of observational geometry and the configuration of an anisotropic region on the birefringence parameters. We use full wavefield simulations in 3-D models to access these influences in a subduction zone setting. Our primary findings are that (a) modest wavefield complications associated with the non-planar geometry of a subduction zone can significantly bias the determination of the fast polarization and time delay associated with shear wave birefringence; (b) shear wave splitting associated with flow in the mantle wedge near the trench (e.g. the forearc) is likely to be weak, owing to short path lengths through an anisotropic region and (c) popular splitting estimators are vulnerable to waveform distortions caused by *S*-wave interaction with the dipping slab interface. We show that simultaneous interpretation of numerous birefringence observations from a variety of propagation directions is a more reliable way to constrain parameters of anisotropy at depth.

**Key words:** seismic anisotropy, shear-wave splitting, subduction zone.

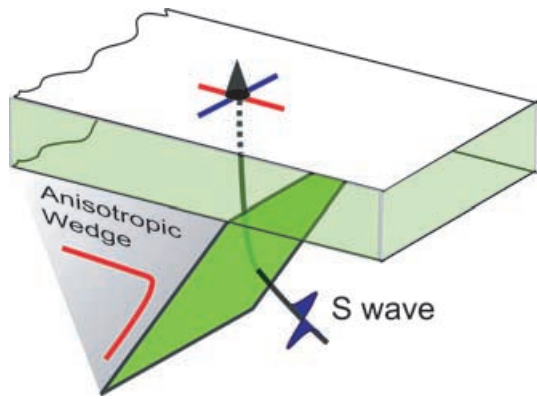
## INTRODUCTION

Seismic wave speed in Earth's mantle is anisotropic, on a range of scales from the hand sample to the whole mantle (Anderson & Dziewonski 1982; Ben-Ismaïl & Mainprice 1998). Anisotropy can arise from a variety of causes, many associated with past or present deformation. In particular, lattice-preferred orientation of olivine crystals (LPO) by deformation is thought to be a dominant cause of anisotropy in the upper mantle (Ribe 1992; Zhang & Karato 1995). Therefore, studies of anisotropic wave speed are of considerable importance for understanding the dynamics of earth's interior (e.g. Park & Levin 2002).

A popular technique to detect anisotropic properties is the birefringence of *S* waves propagating subvertically through the upper mantle (Savage 1999). In an anisotropic medium with simple symmetry (hexagonal, orthorhombic) an *S* wave splits into two components with orthogonal particle motion and different speeds (Crampin 1977). One component propagates with a relatively fast velocity, and its polarization defines the apparent fast-axis of the medium. The other component is polarized in the plane orthogonal to the apparent fast axis, and propagates with a relatively slow velocity. A common method of quantifying the effect of anisotropy on a shear wave is to determine its two 'splitting parameters',  $\phi$  and  $\delta t$ . The parameter  $\phi$

denotes the azimuth of the polarization of the fast shear wave. The parameter  $\delta t$  is the time delay between the 'fast' and the 'slow' components. An observed fast axis is 'apparent' because it represents a complicated averaging of possibly heterogeneous anisotropic structure (Savage 1999; Levin *et al.* 1999; Hartog & Schwartz 2000). A well-recognized shortcoming of the method is the absence of constraints on potential complexity of the anisotropic structure in individual observations of birefringence. Only groups of observations covering a variety of source–receiver geometries offer these constraints. For example, the variation of apparent fast axis with the backazimuth and incidence angle of the incoming shear wave can be used to constrain a layered anisotropic structure, a tilted fast polarization due to an inclined rock texture, or 3-D heterogeneity.

A subduction zone is a tectonic setting where intense deformation of the upper-mantle material is expected (Hall *et al.* 2000; Kneller *et al.* 2005). In a simple case of trench-normal subduction the circulating mantle above the subducting slab should deform in a 'corner flow' regime, with flow lines normal to the trench (Ida 1983). Deformation of dry mantle peridotite within this regime would preferentially align the crystalline lattice with the flow so that the (fast) *a*-axes of the peridotite olivine are oriented coincident with the subduction direction (Fig. 1). As a result, *S* waves oscillating in the trench-normal direction would propagate faster than those



**Figure 1.** A schematic representation of the observational geometry for shear wave birefringence studies of subduction zone mantle wedges. A corner flow regime in the wedge should result in a trench-normal orientation of fast propagation direction (red bar). A common finding in subduction zones is that of a trench-parallel fast direction (blue bar).

oscillating in the trench-parallel direction. Consequently, observations of birefringence in the trench-normal subduction case should yield fast directions  $\phi$  pointing towards the trench (e.g. Fischer *et al.* 2000). For oblique subduction a corresponding obliqueness of  $\phi$  should be expected. Deformation complexity near the edge of a slab would complicate the mantle deformation pattern and associated fast polarization (Buttles & Olson 1998).

Measurements of *S*-wave birefringence in subduction zones, utilizing both teleseismic core-refracted waves like SKS and direct *S* waves from sources in the slab, are plentiful, with an ever-growing data reservoir provided by permanent networks and by temporary arrays. A number of locations where fast polarizations are aligned with the directions of slab descent are known (e.g. Izu-Bonin region, Fouch & Fischer 1996; Long & van der Hilst 2005), including cases where subduction is highly oblique to the trench (e.g. central Aleutians, Bender *et al.* 2004). However, from the global set of subduction zone observations, *S*-wave birefringence measurements actually display a broad range of behaviours, with trench-normal fast directions being notably in the minority. In many instances, trench-parallel fast polarizations are reported in studies of SKS and similar phases' birefringence (e.g. Russo & Silver 1994; Polet *et al.* 2000; Peyton *et al.* 2001; Smith *et al.* 2001), as well as studies of local *S* birefringence (Nakajima *et al.* 2006; Long & van der Hilst 2006). In other settings, however, local *S* birefringence in subduction zones documents significant lateral variability of the fast polarization (e.g. Levin *et al.* 2004; Morley *et al.* 2006).

More complicated flow patterns in the mantle wedge may arise in response to finite slab length (e.g. Buttles & Olson 1998) or slab surface topography (e.g. Hall *et al.* 2000). Alternatively, deviations of seismic anisotropy geometry from that prescribed by the flow pattern may also arise because of alternative causes of seismic anisotropy, such as water's influence on olivine deformation (Jung & Karato 2001; Kneller *et al.* 2005), effects of dynamic recrystallization (Kaminski & Ribe 2001), or the presence of a significant melt fraction (Holtzman *et al.* 2003).

Before we can infer any of the above exciting possibilities as the cause of observed values of shear wave birefringence, a somewhat more mundane problem needs to be addressed. Besides the intrinsic properties of the mantle material, the waveform shape of an *S* wave which has travelled through a mantle corner flow region is affected by many aspects of its propagation geometry. Of obvious importance are both the 3-D configuration of the anisotropic re-

gion and the direction of the seismic ray path through it. As was noted before, it is not possible to separate structural effects from those of the LPO anisotropy on the basis of individual estimates of birefringence. Thus interpretation of shear wave birefringence data from subduction zones always faces a challenge of separating the true effects of intrinsic material properties from those related to observational geometry. Analysis of directional systematics in birefringence measurements may be helpful (e.g. Rumpker & Silver 1998; Saltzer *et al.* 2001; Hartog & Schwartz 2000; Chevrot 2000). The success of these systematics relies on a broad range of propagation directions being sampled—a condition that may be hard to fulfill with a sparse backazimuth distribution of seismic sources.

Another approach relies on prediction of likely observations in complex anisotropic structures through computer simulations of wave propagation. Studies of this type include ray tracing explorations of SKS splitting delays in anisotropic models designed to represent specific subduction zones (Brisbourne *et al.* 1999; Hall *et al.* 2000), finite-difference simulations of shear wave propagation in laterally varying structures (Rumpker & Ryberg 2000), and a study of shear wave birefringence sensitivity to lateral variations in structure performed using spectral-element simulations (Chevrot *et al.* 2004).

In this paper, we use a similar approach to explore the anticipated behaviour of teleseismic *S* waves propagating through an anisotropic wedge. We use full wavefield simulation within a 3-D anisotropic model. Through this calibrated approach we examine, in simulated waveforms, the above-mentioned geometric effects plus the role of wave propagation backazimuth through the wedge. We quantify measurements of birefringence in synthetic waveforms, and in doing so, we also test the performance of commonly employed birefringence analysis techniques. Our primary finding is that wedge geometry of the anisotropic region leads to significant biases in single-phase birefringence estimates, making them a poor guide to the fabric orientation at depth. We conclude that simultaneous analysis of multiple events is necessary in regions with complex anisotropic structure.

## METHODS: A CALIBRATED SERIES OF EXPERIMENTS

We set out to examine the seismological response of wave propagation through a mantle wedge containing corner flow. In order to understand the propagating phase behaviour contained within computed synthetic seismograms, we defined a set of three mantle models:

- (1) a mantle containing pervasive uniform horizontal fabric,
- (2) a wedge-shaped mantle region with horizontal fabric nested in an isotropic mantle and
- (3) a wedge-shaped mantle region with fabric following corner flow trajectories nested in the isotropic mantle.

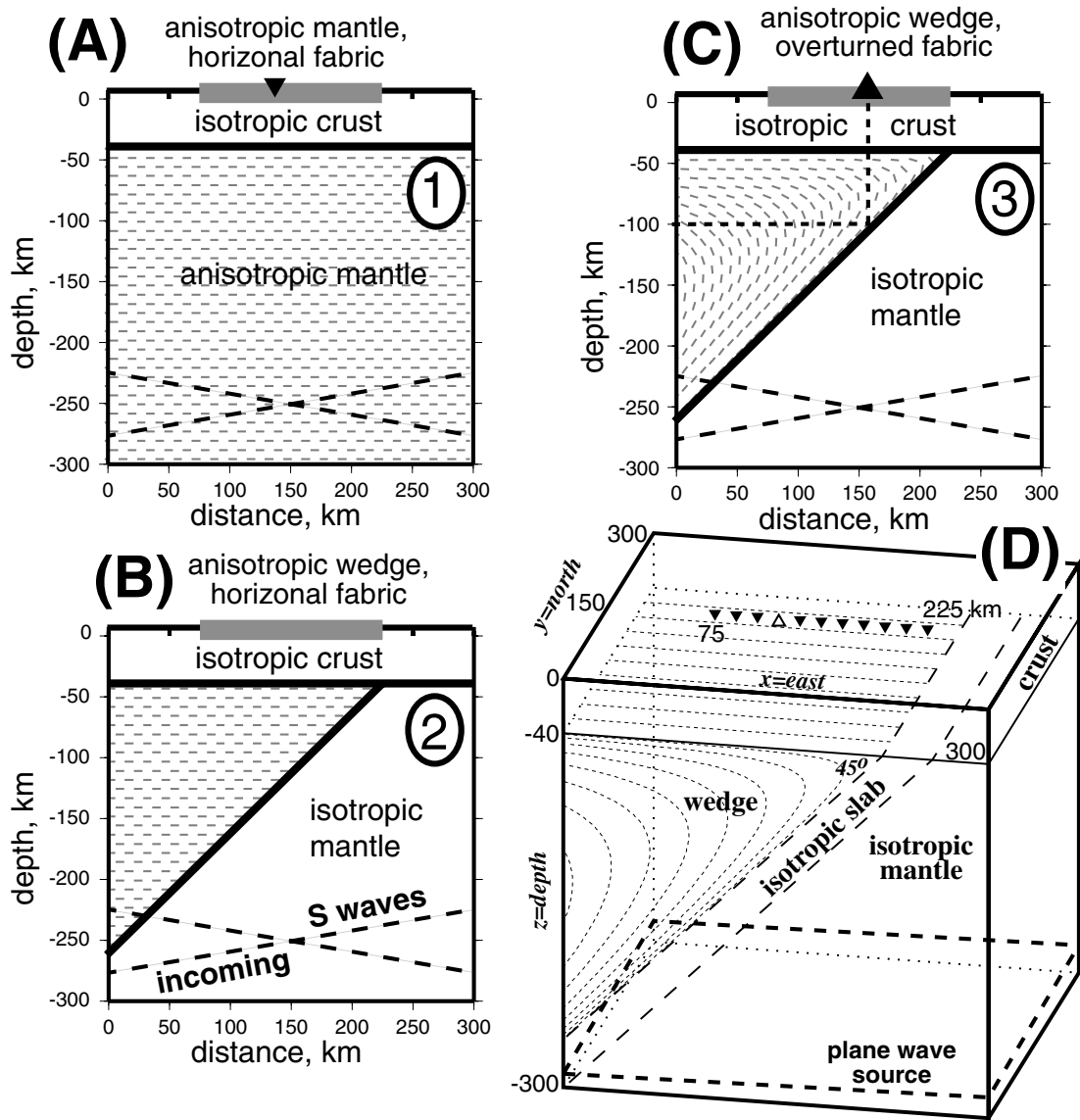
The crust and the mantle surrounding the anisotropic wedge were defined with the goal of reducing the number of extra phases (reflections, reverberations) during the wave propagation simulation. To achieve this, both the crust and the mantle outside the wedge had no anisotropy, and also no distinction was made between the subducting lithosphere and the underlying asthenosphere. Through these models we propagated an upcoming plane wave using a suite of backazimuths to characterize the response to the aligned materials within the models. The corresponding suites of seismograms were analysed by two commonly used techniques for estimating

birefringence parameters (discussed in detail in a subsequent section), and the recently developed group inversion technique of Menke & Levin (2003).

#### Mantle models

To calibrate our procedures, we defined a sequence of three mantle models of increasing complexity. These mantle models were emplaced into an overall 2-D model, which was projected into the third dimension for use in numerical wave propagation. The background model has a 40-km isotropic crust overlying a half-space mantle (Fig. 2a). Model 1 uses this simple crust-mantle configura-

tion wherein the mantle is uniformly anisotropic with a horizontal fast direction. Model 2 has a mantle wedge defined by an interface dipping at  $45^\circ$ . Material within the wedge has anisotropy with a horizontal fast axis while the mantle beneath the interface is isotropic (Fig. 2b). This configuration implies a subduction zone where the slab has the same isotropy as the underlying mantle. This essentially removes the slab as a distinct layer in order to suppress slab-related phases. In Model 3 the mantle wedge defined by an interface dipping  $45^\circ$  has a 2-D flow pattern, defined using analytical curves bounded by the overlying crust and subducting slab (Fig. 2c). Similar to Model 2, the mantle underlying the anisotropic wedge is isotropic in order to suppress other phases.



**Figure 2.** Anisotropic mantle flow models of increasing corner flow geometry. All models have a combination of anisotropic and isotropic mantle overlain by a 40-km thick isotropic crust. Mantle properties are defined to produce 4 per cent *S*-wave anisotropy. The subducting slab and subslab mantle are defined to have same physical properties in order to suppress slab-related seismic phases. Seismic waves are propagated upwards with varying backazimuths and recorded in the 150-km long observing array located above the wedge nose (grey bar). (a) Model 1 has a pervasive and uniform horizontal fabric throughout its mantle. An inverted triangle marks the location for which examples in Fig. 3 are computed. (b) Model 2 has a wedge above an isotropic mantle. Here the corner flowfield is defined as uniformly horizontal. (c) Model 3 has a wedge whose corner flowfield is subparallel to its corner boundaries. Dashed lines and a triangle mark the surface location of 100-km depth contour of the wedge lower boundary, which corresponds to the position of the volcanic arc in real subduction zones. (d) Model 3 depicted in 3-D rendering. Location of the receiver array illustrated by a line of triangles. An open triangle marks the position for which seismograms are shown in Appendix A. Other models are similarly projected in 3-D.

Fig. 2(d) illustrates a 3-D projection of mantle Model 3. Models 1 and 2 are similarly projected. In all three cases, the numerical model is a 300-km earth cube. The crust defines the top 40 km. Models 2 and 3 have a 25-km thick slab dipping  $45^\circ$  whose physical properties are identical to the underlying mantle, making it seismically transparent. For all three model cases, properties of the isotropic crust are:  $V_p = 6000 \text{ m s}^{-1}$ ,  $V_s = 3316 \text{ m s}^{-1}$ , and density  $= 2700 \text{ kg m}^{-3}$ . Mantle regions when isotropic have  $V_p = 8260 \text{ m s}^{-1}$ ,  $V_s = 4691 \text{ m s}^{-1}$ , and density  $= 3300 \text{ kg m}^{-3}$ . The supra-slab wedge flow-field in Model 3 is geometrically defined using families of nested hyperbolas bounded by the overlying crust and descending slab. The fast symmetry axis of the velocity anisotropy was oriented tangential to the hyperbolas on a per-node basis creating a  $315^\circ$  range of (overturning) symmetry axes (Fig. 2d). The ‘receiver array’ (Fig. 2d) and the wave front snapshots (Appendix A; Fig. A2) are situated in the vertical *east-depth* plane at  $y = 150 \text{ km}$ . In Models 2 and 3 the ‘eastern’ (right-most) edge of the receiver array is located just above the supra-slab wedge tip.

Mantle anisotropy is defined in the same manner for all three models, using an elastic tensor with hexagonal symmetry and a fast axis. Specific values of wave speed in anisotropic regions of the model are: along the axis  $V_p = 8260 \text{ m s}^{-1}$ ,  $V_s = 4691 \text{ m s}^{-1}$ ; normal to the axis  $V_p = 8098 \text{ m s}^{-1}$ ,  $V_s = 4507 \text{ m s}^{-1}$ ; density  $= 3300 \text{ kg m}^{-3}$ . Thus anisotropy strength is 6 per cent for  $V_p$  and 4 per cent for  $V_s$ . While hexagonal symmetry does not capture properties of a deformed peridotite fully, it is commonly used in first-order interpretations of shear wave splitting observations (Vinnik *et al.* 1984; Silver 1996; Savage 1999).

In the case of dry olivine alignment by large shear deformation the fast axis should align with the flow (Zhang & Karato 1995). Thus in models 1 and 2 we simulate horizontal flow, while in model 3 we simulate corner flow.

A 100 km-thick mantle layer of uniform fabric with a horizontal symmetry axis and anisotropic parameters as described above is expected to yield a splitting delay of  $\sim 1 \text{ s}$  within subvertically propagating shear waves (Silver 1996). We built our models so that our upcoming plane waves propagate through the mantle for approximately 200 km in order to accentuate the splitting effect.

### Wavefield simulation

The wave propagation simulations were performed using a 3-D finite difference implementation of the anisotropic elastic wave equation (Okaya & McEvilly 2003). To handle anisotropic media the tensorial form of the stress-strain relationship was employed instead of the isotropic form using Lamé’s constants. Full arbitrary rotation of the elasticity tensors was implemented in order to allow for anisotropy orientations aligned with 2-D and 3-D patterns of mantle flow. The finite difference code uses the stress-velocity form of elasticity and a discretization that is staggered fourth-order in space and second-order in time (Graves 1996).

A plane wave source was used to initiate propagation. The plane wave had a reference point at a depth of 250 km measured at the lateral centre of the model ( $x, y = 150, 150 \text{ km}$ ). The plane wave’s inclination was  $10^\circ$  and a sequence of runs used backazimuths at  $90^\circ$ ,  $120^\circ$ ,  $135^\circ$ ,  $150^\circ$ ,  $180^\circ$ ,  $210^\circ$ ,  $225^\circ$ ,  $240^\circ$  and  $270^\circ$ . Thus the plane wave source was inclined but in different orientations at  $30^\circ$  and  $45^\circ$  increments between east to west through south. The plane wave source waveform was a single cycle 0.2 Hz sine wave and was excited using the body components of a stress pulse at each node point along the inclined plane; the primary component of stress was thus upward.

A stress source within the elastic wave equation spawns propagating  $P$  and  $S$  waves which in the case of our source definition produced larger amplitude  $P$  waves. The  $P$  waves do not represent a realistic observational situation, because  $P$  waves from a distant earthquake would arrive much earlier than in our simulations. In selecting time windows for shear wave birefringence analysis we made sure that  $P$  wave energy did not interfere with the measurement.

While the plane wave source was computed throughout the anisotropic model, we defined measurement points in an east-west three-component linear receiver array located between  $x = 75$ – $225 \text{ km}$  at  $y = 150 \text{ km}$  (Fig. 2); for this study we use a station spacing of  $dx = 15 \text{ km}$ . Waves were propagated for 100 s using an internal time step of 0.050 s for stability. Computation was performed using a multi-CPU shared memory platform; each backazimuth simulation required 2.4 Gbytes memory and 20 cumulative CPU hours of computation. Resulting time-series and wavefields are illustrated in Appendix A.

### Birefringence measurements and methods of analysis

Evaluation of the birefringence parameters  $\phi$  and  $\delta t$  may be performed in a number of ways. Naturally, the results of different measurement methods are very similar when the original signal is simple and clear, that is, birefringence arising from a uniformly anisotropic medium. However, as explored at some length in Menke & Levin (2003), noise and/or complicated birefringence (e.g. from two or more anisotropic regions) may lead to algorithm-dependent parameter values. If values of  $\phi$  and  $\delta t$  are a function of the estimator, the geological interpretation becomes problematic.

To evaluate the importance of technique-specific biases, we use two commonly employed methods. The first, hereafter referred to as *Xcor*, seeks a rotation of observed seismograms that would yield waveforms on two components that are most similar and then evaluates the delay between them via cross-correlation (Ando *et al.* 1983; Bowman & Ando 1987; Levin *et al.* 1999). The second, hereafter referred to as *minT*, seeks a combination of a rotation and a time delay that minimizes one of the components of motion (Vinnik *et al.* 1984; Silver & Chan 1991), effectively removing the effect of anisotropy.

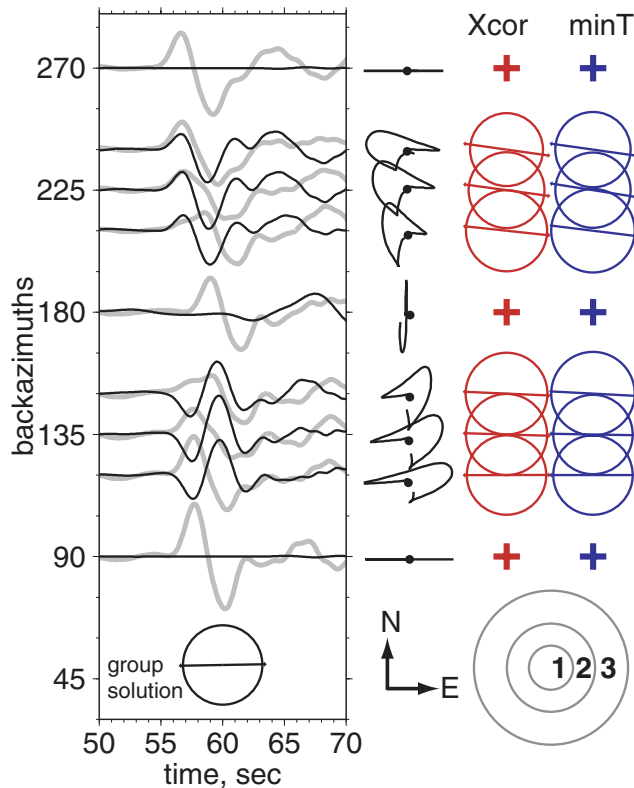
An expected feature of shear wave propagation in an anisotropic medium is the lack of birefringence when the propagation is either along or normal to the direction of the anisotropic symmetry axis. For such a waveform an algorithm estimating birefringence should return negligible  $\delta t$  value, and  $\phi$  identical to the backazimuth. Such observations are often called ‘null’ measurements (Silver & Chan 1991). In practice noise or waveform complexity is likely to yield a spurious value of  $\phi$ . Also, a feature of our implementation of the *minT* algorithm is to return  $\delta t$  value close to the maximum allowed in the search in cases where birefringence is very small or else the noise level is high (Menke & Levin 2003). This unwanted ‘feature’ of the *minT* algorithm is commonly known among seismologists who use the technique. Recognizing this, we introduce a broader definition of the ‘null’: for the *minT* algorithm: a ‘null’ is defined when the estimated delay time  $\delta t$  exceeds 2.8 s (our  $\delta t$  search range is 0–3 s). The *Xcor* algorithm is less vulnerable to this behaviour, so a ‘null’ is defined when  $\delta t < 0.05 \text{ s}$ . We note that in real data analysis ‘null’ measurements present a challenge. A ‘null’ observation can signify either the absence of birefringence, or else the difficulty of performing a measurement, due to noise or structural complexity.

In addition to birefringence estimators based on individual waveforms we apply the method proposed by Menke & Levin (2003) that is based on explicit wavefield comparison of ‘observed’ and

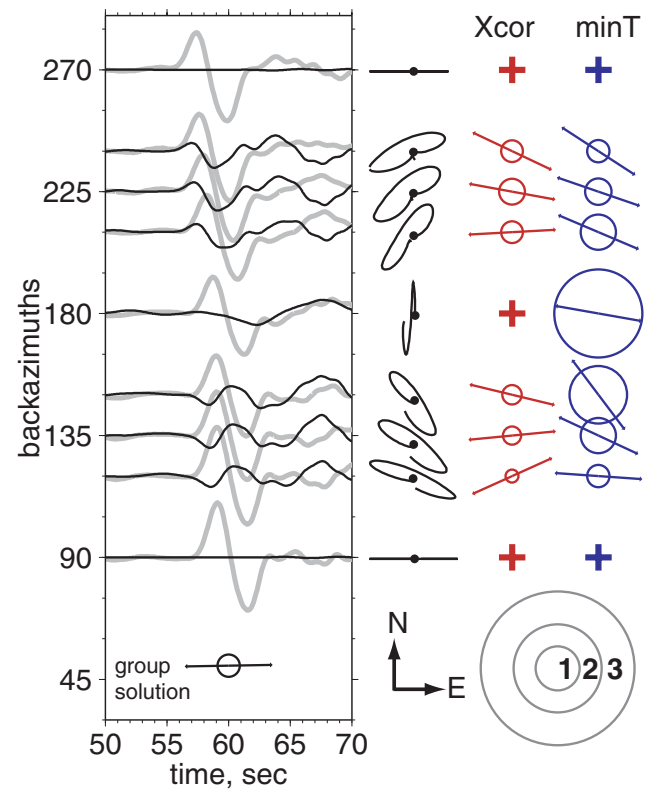
‘predicted’ wavefields and allows a search for group solutions using multiple observations. This group estimate, discussed in further detail below, requires solutions to fit a set of observations simultaneously and thus tends to produce more stable results.

#### Waveforms and resulting measurements of shear wave birefringence

Fig. 3 illustrates waveforms resulting from our computations, and results of applying different measurement methods. We show results for a single node of the synthetic observing array, and for all nine simulated directions of propagation. In Model 1 (Fig. 3a) we obtain delays between fast and slow components on the order of 2 s, as expected given model parameters (200 km thick layer of horizontal fabric with 4 per cent anisotropy). Fast directions close to E–W are recovered for all propagation directions except when the wave path is in the plane of the symmetry axis or else normal to it. In these special cases we get ‘null’ measurements, defined as described in the previous section. Particle motion plots show that the first motion of the simulated *S* wave is indeed E–W. In spite of that, simulations for



**Figure 3.** An illustration of the measurement strategy. Parts (a) and (b) of the figure represent Models 1 and 3, respectively. Left panels show radial (shaded) and transverse (solid) components of the *S* wave simulated for 9 different backazimuths and recorded by the ‘receiver’ at the model coordinate 135 km (see Fig. 2). The middle columns show horizontal particle motion diagrams (N and E directions are marked on the plots). Time windows used to plot particle motions are 53–60 s in (a) and 56–63 s in (b). Two rightmost columns depict birefringence estimates obtained for each backazimuth by two different methods (as marked). ‘Null’ measurements are marked by fat crosses. See text for the definition of ‘null’ for each measurement technique. The circle size is proportional to the delay value, and the vector is aligned with the fast direction. A group solution using all nine traces shown is illustrated by a black diagram beneath the waveforms. Shaded concentric circles show  $\delta t$  scale in seconds.



**Figure 3.** (Continued.)

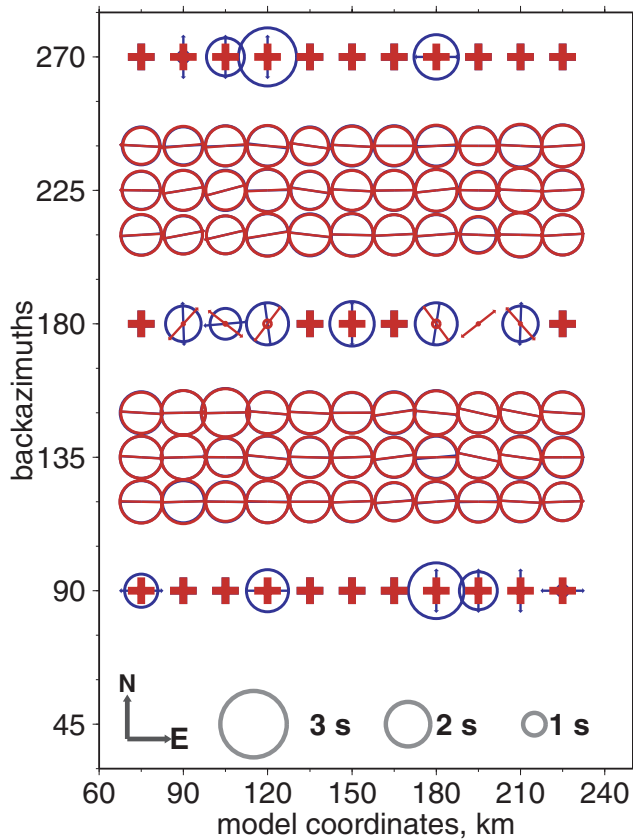
backazimuths 210°–240° yield  $\phi$  values that differ somewhat from the expected E–W alignment. As the measurement involves analysis of a time window  $\sim 20$  s long, we can conclude that complications along the wave path of a particular waveform (most likely related to the presence of 40 km thick isotropic crust) contribute enough energy to the resulting broad-band waveform to alter the anisotropy-induced birefringence.

In Model 3 (Fig. 3b) we see a more complicated picture. Particle motions of shear waves that have traversed a wedge-shaped region with overturning fabric show clear ellipticity, but polarization of the fast wave is hard to evaluate from them. Measurements of birefringence using both methods yield ‘nulls’ for propagation along the direction of the symmetry axis, but only *Xcor* method yields a ‘null’ for propagation normal to it. The *minT* technique returns a  $\Delta t$  value that is large but lower than the ‘null’ threshold, and a  $\phi$  value that significantly differs from the expected E–W direction. For propagation directions oblique to the symmetry axis *Xcor* and *minT* methods yield divergent results, except for propagation with backazimuths 225° and 240°. In these latter cases measurements by two methods yield similar values, with  $\phi$  being significantly different from the expected value. In both exercises results of the group inversion based on all 9 waveforms are very close to the expected parameters of birefringence.

## RESULTS

### Model 1: uniform horizontal fabric

This model provides a reference for the more complicated Models 2 and 3. A wavefield propagation snapshot through this model is shown in Fig. A2(a) of Appendix A.



**Figure 4.** Estimates of shear wave splitting in Model 1 are presented as functions of the ray backazimuth (vertical axis) and the array node position (horizontal axis). See Fig. 2 for the location of the array. See Fig. 3 for an explanation of the symbols. We plot values every 15 km along the array. Blue symbols are for the *minT* method, red symbols are for the *Xcor* method. Where only red symbols appear, blue symbols are directly underneath and the results of two methods are identical.

Fig. 4 shows splitting measurements obtained by the *Xcor* and *minT* methods for all positions sampled along the receiver array, and for all nine values of wave backazimuth. For waves propagating at an angle to the planes containing the anisotropic symmetry axes the splitting signal is stable, and is recovered by both measurement techniques, with little discrepancy. The majority of measurements yield  $\phi$  close to the ‘true’ E–W orientation, and  $\delta t \sim 2$  s. For some combinations of the wave propagation direction and array position estimated  $\phi$  values deviate from the E–W orientation. As this model has no lateral changes, these minor deviations are most likely caused by the finite lateral extent of the model, suggesting the accuracy limit of our synthetic seismograms.

Wave propagation within planes of the fast and the slow axes of the velocity ellipsoid (i.e. backazimuths of  $90^\circ$ ,  $180^\circ$ , and  $270^\circ$ ) should yield no splitting. We note that measurement results from two techniques often diverge for simulations with these backazimuths. For along-axis propagation, *Xcor* method returns ‘nulls’ for all receiver array positions. Results of the *minT* technique are varied, with some ‘null’ results, but also some relatively large  $\delta t$  values and  $\phi$  values orthogonal to the E–W true direction. For simulations along backazimuth  $180^\circ$  (normal to the symmetry axis in the model) we get some ‘null’ values for both techniques, but mostly results are erroneous, in that  $\delta t$  values do not correspond to the null definition (see above) while  $\phi$  values diverge strongly from the expected E–W orienta-

tion. Clearly, we observe a manifestation of the method-dependent outcome of birefringence measurement discussed before.

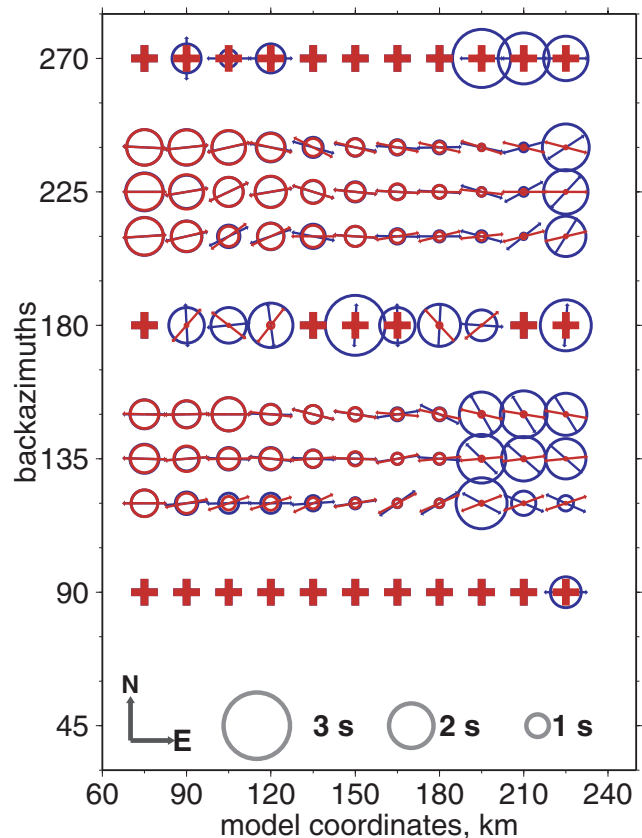
#### *Model 2: An anisotropic wedge with horizontal flow*

In this model the anisotropic velocity is restricted to a wedge-shaped region (Fig. 3b). The fabric orientation and the intensity of anisotropy are the same as in Model 1 described above. The position of the receiver array with respect to the wedge-shaped anisotropic region is illustrated in Fig. 2(b).

Fig. A2(b) shows snapshot wave fronts through this model, with the dashed pattern illustrating the region where horizontal-axis anisotropic fabric is present. The most notable difference in the wavefield from Model 1 (bulk anisotropic mantle; Fig. A2a) is the restricted lateral extent of the transversely polarized *S* wave. Naturally, it only arises where the anisotropic material is present.

Fig. 5 shows measurements of shear wave splitting along the receiver array. The upper surface of the ‘slab’ is at 100 km depth beneath an array element located 100 km from its western (left) end ( $X = 175$  km in the coordinates of the model; see Fig. 2c). Noting the position of this depth, significant because it is the locus of volcanic arcs above subduction zones, is useful for relating our findings to results of studies. The notable aspects of the simulation are as follows.

The overall intensity of birefringence is low. The shear wave splitting delay exceeds 1 s only over the leftmost third of the array, where the wedge-shaped anisotropic region is the thickest. In addition, the splitting is systemically larger for southwestward backazimuths



**Figure 5.** Measurements of shear wave splitting in a wedge with horizontal fabric (Model 2), notation as in Fig. 4.

compared to those from the southeast due to increased travel path through the anisotropic wedge. In the area between the ‘volcanic arc’ and the tip of the wedge the birefringence is especially weak, and  $\phi$  values returned by both techniques do not match the expected E–W orientation. The deviation of  $\phi$  are especially prominent for backazimuths  $120^\circ$  through  $150^\circ$ . The *minT* method yields  $\delta t$  on the order of 2 s, and fast directions that are nearly orthogonal to the expected value. The *Xcor* method returns small  $\delta t$ , and  $\phi$  values that are closer to the expected values. At distances of 45 km and more from the wedge tip (array nodes at model coordinates 75–190 km) the two techniques largely agree, and in places (e.g. backazimuth  $210^\circ$  through  $240^\circ$ , array elements at model coordinates 95–135 km) manifest significant deviations of the ‘effective’ birefringence from the expected value. As in the case of Model 1, propagation along the symmetry axis yields ‘null’ measurements for most array nodes. In a few cases *minT* method returns measurements that do not qualify as ‘nulls’, with fast directions either along the symmetry axis or normal to it. Propagation normal to the symmetry axis results in problematic measurements. For some array nodes both birefringence estimators return ‘nulls,’ as expected from the fabric orientation. However, in most cases the *minT* technique produces a measurement that satisfies the criteria for reliability, sometimes with a very large  $\delta t$ . The *Xcor* method also returns several false measurements, typically with small  $\delta t$  and  $\phi \sim 45^\circ$  from the E–W direction.

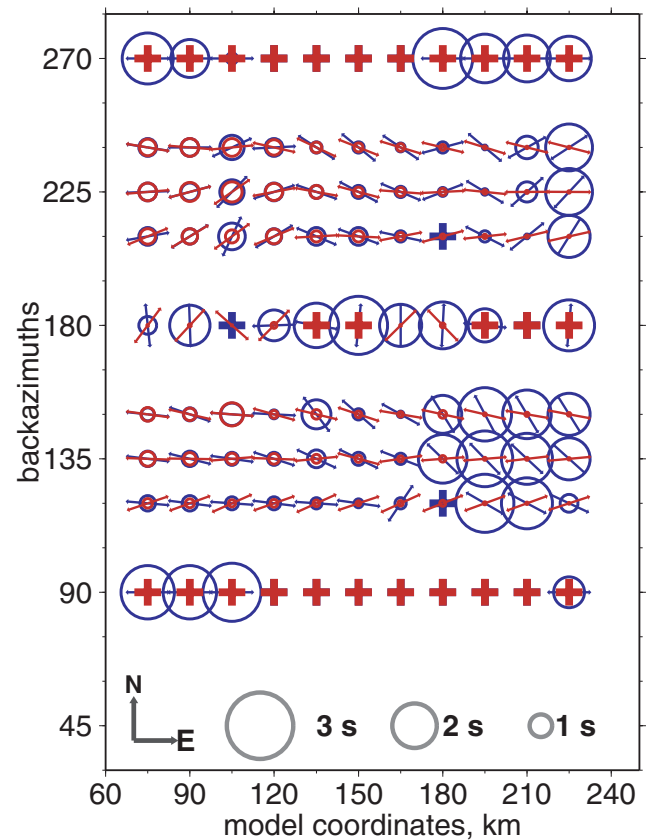
Overall, we conclude that the wedge geometry of the anisotropic region leads to significant biases in single-phase birefringence estimates.

### Model 3: An anisotropic wedge with corner flow

The final and most realistic simulation includes the corner-flow geometry of the fabric within the wedge (Figs 2c and d). The wavefield (Fig. A2c) is generally similar to that computed in the wedge with horizontal-axis fabric, but the transverse component is noticeably weaker. Fig. 6 shows birefringence estimates. While the overall nature of the result is similar to that from Model 2, influences of the complicated fabric and the wedge-shaped region combine, with stronger overall effect on the wavefield. Specifically, the birefringence is very weak and the discord in individual measurements occurs over a broader range of backazimuth-position combinations. In particular, the *minT* technique yields very large delay values and fast directions at high angles to the expected orientation for backazimuth values between  $90^\circ$  and  $180^\circ$ , at all array positions between the ‘volcanic front’ and the wedge tip. The *Xcor* technique yields fast directions closer to those expected (i.e. E–W), but delay values are very small. Further from the wedge tip both *Xcor* and *minT* techniques are commonly in agreement on a ‘wrong’ solution, in the sense that the fast polarization deviates from the expected E–W orientation.

## GROUP ANALYSIS

The birefringence estimator of Menke & Levin (2003) is based on an explicit comparison of ‘observed’ and ‘predicted’ wavefields, and searches for group solutions using multiple observations. The wavefield comparison is achieved by convolving observed seismograms  $R(\tau)$  and  $T(\tau)$  for the radial and transverse components, respectively, with simulated radial and transverse impulse responses  $r(m, \tau)$  and  $t(m, \tau)$  of a test model  $m$ , in the following way:  $x(\tau) = r(m, \tau) * T(\tau)$  and  $y(\tau) = t(m, \tau) * R(\tau)$ , where  $\tau$  is time and  $*$  denotes convolu-

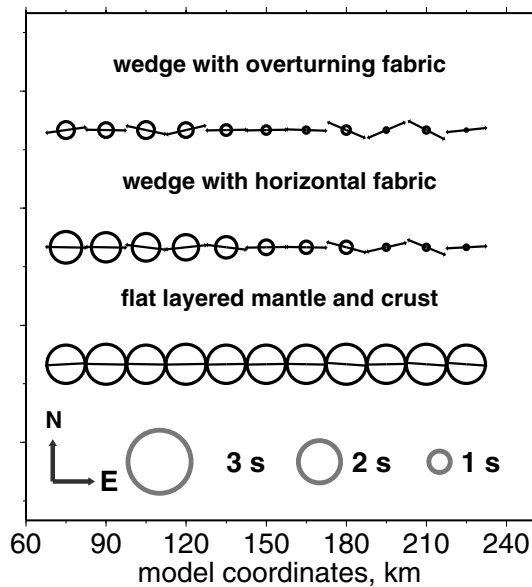


**Figure 6.** Measurements of shear wave splitting in a wedge with corner flow (Model 3), notation as in Fig. 4.

tion. This method exploits the commutative property of convolutions, namely that an incoming shear wave  $P(\tau)$  will be convolved by the radial and transverse impulse responses  $r'(\tau)$  and  $t'(\tau)$  of the underlying medium to produce the observed  $R(\tau) = r'(\tau) * P(\tau)$  and  $T(\tau) = t'(\tau) * P(\tau)$ . If  $r(m, \tau) = r'(\tau)$  and  $t(m, \tau) = t'(\tau)$ , then the birefringence is simple and the time-series  $x(\tau)$  and  $y(\tau)$  should be identical. In practice, the birefringence has some complexity and the best-fitting earth model is the one that minimizes the misfit  $e(\tau) = x(\tau) - y(\tau)$ . The technique works best (i.e. recovers the expected values most closely) when a group of birefringence observations is used with a wide range of backazimuth (Menke & Levin 2003).

In the context of this paper we used the group technique to determine the best-fitting model that consists of one anisotropic layer with a horizontal symmetry axis. We also tested whether the wedge geometry biases the splitting to resemble the effects of two-layer anisotropy, or anisotropy with a tilted axis of symmetry (see Discussion section). For comparison with other methods of measuring shear wave splitting, anisotropic properties of the models are described by a combination of  $\delta t$  and  $\phi$  in a vertically propagating SV wave that goes through it. For each node of the simulated seismometer array we used waveforms computed for all backazimuths to find a best-fitting pair of  $\delta t$  and  $\phi$ . These are illustrated in Fig. 7, for the three models described above.

In the case of the 200-km layer of horizontal fabric (Model 1), group solutions are nearly identical to the theoretically predicted result: in the best-fit model a vertically propagating SV wave accumulates 2 s of delay with a fast polarization being E–W. Best-fitting



**Figure 7.** Parameters of anisotropy obtained by inverting groups of nine shear wave splitting observations at each node of the synthetic receiver array. Values of anisotropy are represented by a fast direction (bar, east is horizontal) and a delay that will be imposed on a vertically propagating shear wave (circle).

models for the case of a wedge with horizontal fabric (Model 2) display a progressive decrease in the amount of delay time  $\delta t$  from left to right (from backarc to the trench), consistent with the decreasing thickness of anisotropic material. Very small delays are predicted in waves sampling the ‘nose’ of the wedge. Also, deviations from the expected fast orientation are seen for the right-most third of the synthetic array, where the anisotropic material is most thin.

In the case of a wedge with an overturning fabric pattern (Model 3) the birefringence is very weak, with group-solution time delays for all array positions  $\delta t < 1$  s. The deviation of group-solution fast directions  $\phi$  from the expected E–W orientation is stronger, and takes place over a broader range of distances from the wedge tip. We conclude that the complexity of wedge anisotropy in this corner flow context leads to bias in birefringence estimates.

## DISCUSSION

We used a common value of anisotropic intensity, 4 per cent, in our simulations for which 1 s of shear wave birefringence is expected per 100 km of propagation distance in our bulk mantle model. For the models containing the spatially limited mantle wedge, we obtained rather modest amounts of shear wave birefringence (under 1 s). When compared with commonly reported values (see a review by Savage 1999; also Polet *et al.* 2000; Smith *et al.* 2001) our findings suggest that either the anisotropic fabric in real-world wedges is much stronger than in our models, or else measurements in excess of  $\sim 1$  s in subduction zones arise from anisotropy beneath the supraslab wedge, e.g. in the slab or the asthenosphere beneath it (Peyton *et al.* 2001).

Our tests of the performance by two common methods of shear wave birefringence estimation showed the value of cross-checks afforded by the simultaneous use of both. We conclude from our study that considerable divergence in  $\delta t$  and  $\phi$  values may be used

as a quick data-based diagnostic of the applicability of simple anisotropic models. In our simulations the results of the two methods agree well when the birefringence signal was largely from a single source of anisotropy (in Model 1, and also in western-most parts of Models 2 and 3 when rays arrived from the southwest). The largest disagreement was seen for signals where  $S$  wave pulse shape was perturbed by other aspects of the simulation, for example, propagation through the wedge at  $90^\circ$  to the direction of anisotropic symmetry axis.

Also, our results suggest the importance of stating clearly the definition of ‘null’ measurements in real data studies. For example, in Model 3 simulations many observations from the eastern end of the array could have qualified for a ‘null’ designation under a different definition. Such ‘null’ results, if seen from one backazimuth only (e.g. due to a limited range of observations in a data set), could have been interpreted as evidence of anisotropic symmetry along that backazimuth. We stress again the ambiguity of a ‘null’ measurement in individual observations of shear wave birefringence: either no birefringence is present or else waveform complexity confounds the birefringence estimate.

While the wedge geometry of the anisotropic region affects shear wave splitting results most strongly, other effects are noticeable as well. Our calibration test results, and results for the thicker side of the wedge models (i.e. farther from the trench, typically in the backarc) show evidence for the influence of the crust and/or the orientation of the wave front with respect to the fabric in the model. The cause for concern lies in the fact that splitting estimation results, while deviating from ‘true’ values, can be quite stable, as evidenced by good agreement of *minT* and *Xcor* techniques. Thus, in real data with similar properties the formal metrics of the birefringence estimate will be good, while the orientation and intensity of the anisotropy inferred will be wrong. Recognition of this sort of modest alteration of the birefringence in observed data appears impossible without explicit waveform modelling as done in our ‘group inversion’ solutions. Even those solutions do not always return correct results where the anisotropic signal is weak. In our simulations using anisotropic wedges (Models 2, 3) fast polarization direction deviates from the expected E–W orientation (Fig. 7). One plausible cause for that is the asymmetry of our simulated observations set that contains incoming waves between backazimuths of  $90^\circ$  and  $270^\circ$ , but no ‘northerly’ backazimuths. Presence of observations ‘reversed’ in backazimuth would probably lead to a better constraint on the fast polarization direction. A greatly unbalanced set of incoming waves is the rule, rather than the exception, in birefringence studies, however. Therefore, one should expect some bias in the fast polarization associated with wavefield distortion by 2-D mantle-wedge geometry. Such distortion can mimic the effects of a tilted-axis or multilayer anisotropic structure.

To test whether this would be the case we used synthetic seismograms simulated in Model 3 (with the corner flow in the wedge) to obtain group solutions for two additional classes of anisotropic models: (a) a model with two layers of horizontal-axis anisotropy and (b) a model with one layer of anisotropy that can have an arbitrary orientation of the symmetry axis. The nine synthetic seismograms computed for an observing array element at model position 180 km (see Fig. 3) were used as ‘data’. Table 1 lists parameters of best-fitting models in each class, including the 1-layer model (see Fig. 7), and compares their data misfit statistics. Notably, the East–West orientation of fabric ( $\phi = 90^\circ$  strike) is recovered in all cases. In the best two-layered model most of the delay comes from the layer with a fast direction  $80^\circ$ , while the best model with inclined symmetry axis has a fast direction of  $100^\circ$ . In case of the inclined

**Table 1.** Results of a group inversion performed using synthetics computed for the case of corner flow in a wedge. The ‘data’ come from an array element at model coordinate 180 km. Note an order of magnitude difference in the measure of fit between simple models and a full synthetic comparison.  $\phi$  is fast direction (an expected value is  $90^\circ$ );  $\delta t$  is delay;  $H$  is thickness of an anisotropic layer with a tilted symmetry axis, where tilt is measured from the vertical towards the azimuth of the fast direction;  $B$  is the magnitude of anisotropy (same for both  $P$  and  $S$  waves). See ‘group analysis’ section for the definition of a data misfit measure, and ‘Discussion’ section for the detailed description of the comparison exercise.

| Model class                                                                      | Model parameters                                                                                               | Data misfit measure |
|----------------------------------------------------------------------------------|----------------------------------------------------------------------------------------------------------------|---------------------|
| One-layer anisotropy                                                             | $\phi = 97^\circ$ ; $\delta t = 0.3$ s                                                                         | 0.102900            |
| Two-layer anisotropy                                                             | $\phi$ upper = $150^\circ$ ; $\phi$ lower = $80^\circ$ ;<br>$\delta t$ upper = 0.3 s, $\delta t$ lower = 0.5 s | 0.100348            |
| Tilted anisotropic<br>Symmetry axis                                              | $H = 60$ km, $\phi = 100^\circ$ ,<br>Tilt = $150^\circ$ ; $B = 0.04$                                           | 0.100906            |
| A synthetic seismogram from a nearby array node computed using the ‘true’ model. |                                                                                                                | 0.0135388           |

axis model, the fast axis (tilted  $150^\circ$  from vertical) is near-normal to the plane bounding the anisotropic wedge, an orientation that is not surprising given the shape of the flow lines in it (see Fig. 2). Another noteworthy fact is that there is a clear preference (in terms of smaller values of data misfit measure) for a model that is more complex than a single layer of anisotropy with horizontal axis of symmetry. The difference in data misfit measure values for ‘single layer tilted’ and ‘two layers’ solutions is small, and would hardly help distinguish them in a real-world study.

To obtain an anisotropic model on the basis of a group of observed seismograms, we convolve observations with synthetic seismograms predicted from simple models. In the present study we have a possibility to evaluate performance of such simple models relative to the true complex model that was used to generate the seismograms. For such a comparison we convolved ‘observations’ at one array element (at 180 km) with the synthetic seismograms at an adjacent array element (at 183 km). For shear waves with 5 s period there should be little difference in birefringence observed 3 km apart, while details of synthetic wave propagation computations are different enough to avoid numerical problems. In essence, we replicated a scenario where a complex waveform simulation is used to explain collected data. We achieved an order of magnitude improvement in the data misfit measure with respect to all group solutions that use simple anisotropy models (Table 1). This last result is encouraging as it shows that once a proper model is constructed, the quality of data fit will be obviously superior to other, improper models. It also suggests that most simple models developed to match shear wave splitting observations are likely far from capturing the full complexity of fabric at depth.

The approach taken in this paper is a practical way of evaluating expected values of shear wave birefringence in specific geodynamic conditions. Two key aspects of the approach are a full wavefield simulation, and the use of multiple measurement techniques for resulting wavefields. The earth structure models used in this paper are simplistic, but there are no fundamental obstacles to the implementation of more realistic models, based on anisotropic properties of real rocks, results of regional tomographic inversions, and geodynamic modelling. The computational expense will not change with the degree of model complexity, except if increased spatial sampling of the model is needed.

With respect to measurement techniques, our experiments demonstrate the value of cross-checks with different techniques, and an overall superiority of techniques that use multiple observations.

Finally, our experiment comparing data fit achievable with simple and ‘true’ models of anisotropy implies that our understanding of

seismic anisotropy at depth will advance considerably once we stop relying on simplistic models implicit in commonly used metrics of shear wave birefringence, and turn to more realistic representations of anisotropic structure.

## CONCLUSIONS

Modest wavefield complications associated with the non-planar geometry of a subduction zone can significantly bias the determination of the fast polarization and time delay associated with shear wave birefringence. Splitting associated with circulation in the mantle wedge near the trench (e.g. the forearc) is likely to be weak, owing to short path lengths. In such situations, popular splitting estimators are vulnerable to waveform distortions caused by  $S$ -wave interaction with the dipping slab interface. Even for a simplified scenario of anisotropy with a uniform horizontal symmetry axis, trench-parallel birefringence can be inferred (mistakenly) from isolated measurements in a trench-normal setting, especially when birefringence is estimated by minimizing the transverse component.

A more reliable and robust birefringence measurement can be obtained in a mantle wedge setting via a group inversion of splitting observations. We apply the inversion technique of Menke & Levin (2003) in this study, but improvement can be expected from other multiple-event estimators as well. The Menke and Levin estimator gives a reliably small amount of splitting at locations above the forearc mantle wedge, but is still susceptible to modest bias ( $10$ – $20^\circ$ ) in fast polarization if the distribution of backazimuth is not symmetric about the trench-normal direction. Our study suggests that multiple-event splitting analysis and careful interpretation of mantle-wedge anisotropy are necessary. Older interpretations based on single-record splitting estimates, particularly the minimize-the-transverse algorithm, may need re-evaluation.

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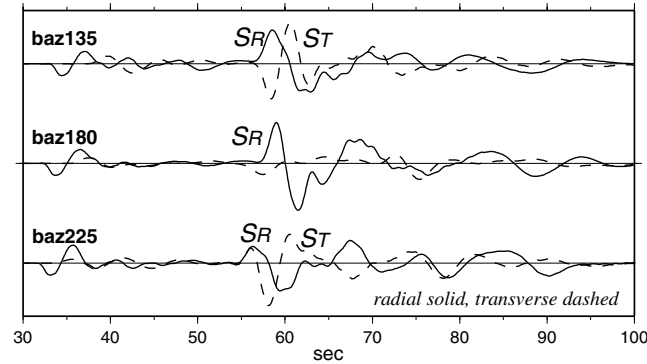
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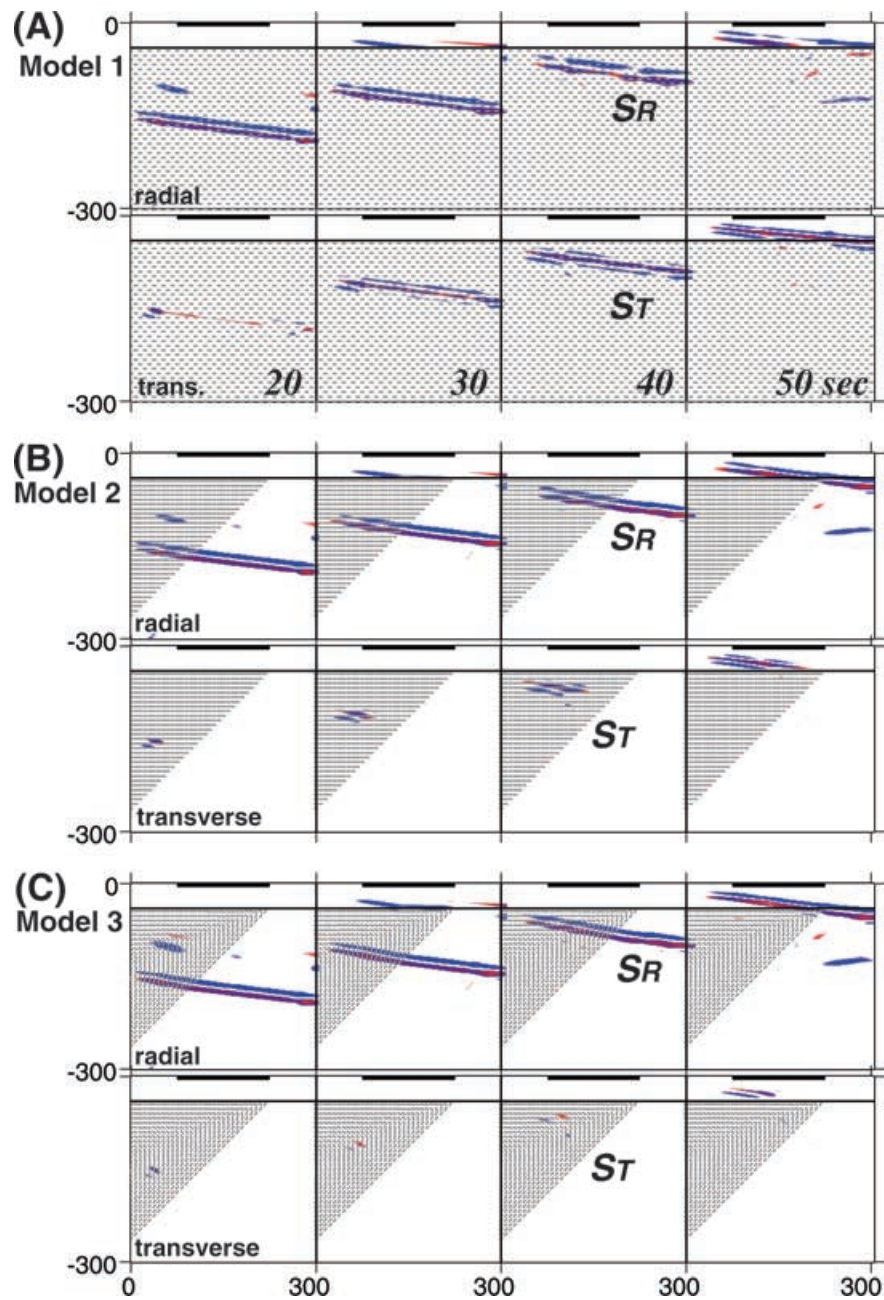
# APPENDIX: WAVEFIELDS AND SEISMOGRAMS OF SYNTHETIC PLANE WAVES COMPUTED IN ANISOTROPIC MANTLE MODELS

Examples of seismograms calculated for Model 1 (full anisotropic mantle, no slab) using backazimuths  $135^\circ$ ,  $180^\circ$  and  $225^\circ$  are shown in Fig. A1 for a station at  $(x, y) = (105, 150 \text{ km})$ , marked by the open triangle in Fig. 2(d). Nearly two seconds of splitting can be observed at backazimuth  $135^\circ$  in accordance with the 4 per cent shear wave velocity anisotropy (see Fig. 3).

Snapshots of wavefields propagating through the three models are presented in Fig. A2. Here the radial and transverse components of the particle velocity fields relative to the propagation direction are shown at 20, 30, 40 and 50 s propagation time. In Model 1 (anisotropic mantle, no slab) the shear plane-wave splits into strong radial and transverse phases ( $S_R$  and  $S_T$  in Fig. A2a). In contrast, for the wedge in Model 2 that has a uniform horizontal fabric, the split  $S_T$  is spatially limited to within the wedge (Fig. A2b) with diminishing magnitude and delay as the nose thins. For Model 3, the overturning flow field further diminishes the amount of shear wave splitting (Fig. A2c).



**Figure A1.** Synthetic seismograms at one station (open triangle in Fig. A1) for three backazimuths. The earth model was Model 1 (anisotropic mantle with uniform fabric, no slab).  $S_R$  and  $S_T$  denote the radial and transverse  $S$ -wave phases, respectively.



**Figure A2.** Snapshots of wavefield simulations for backazimuth  $225^\circ$  in (A) model 1 (anisotropic mantle, horizontal flowfield), (B) model 2 (anisotropic wedge, horizontal flowfield) and (C) model 3 (anisotropic wedge, overturned flowfield). For each model, the  $300 \times 300$  km  $x - z$  (east-depth) cross-sectional slice is located at  $y = 150$  km and thus includes the location of the synthetic 'recording array' (black bar). The wavefield snapshots are rotated into the radial and transverse directions of the propagating wave at time steps 20, 30, 40 and 50 s. The radial ( $S_R$ ) and transverse ( $S_T$ ) phases of the propagating S wave are identified. Note the diminished amount of the  $S_T$  phase due to the presence of the wedge and its overturned flowfield.