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PHYSICS OF THE EARTH AND PLANETARY INTERIORS

Physics of the Earth and Planetary Interiors 158 (2006) 174-189

www.elsevier.com/locate/pepi

Texture of mantle lithosphere along the Dead Sea Rift: Recently imposed or inherited?

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Received 20 June 2005; received in revised form 12 December 2005; accepted 8 May 2006

Abstract

Seismic anisotropy, a property linked to the texture of the mantle rock, should be distributed with depth along the trace of the Dead Sea Rift (DSR), owing to a combination of present day and ancient tectonics. Using data from four permanent and one temporary seismic observatories we evaluate birefringence (splitting) of 91 teleseismic core-refracted shear waves, primarily SKS phases. We find significant levels of birefringence in the bulk of observed phases. We also find that birefringence parameters (fast directions and delays) vary as a function of source-receiver geometry. Notably, the pattern of this directional variation in birefringence is quite similar at all sites we have examined. We interpret observed birefringence in SKS phases in terms of one- and two-layer models. Single-layer models for all stations exhibit a fast polarization oriented 12–19° east of north, with anisotropy sufficient to generate 1.3-s time delay. We find strong evidence for at least two distinct anisotropic layers. For the two layer models, the upper layers resemble the single-layer models, showing near-north fast polarizations and delays 0.3–0.6 s. One site, at the northern end of the DSR, displays a higher degree of anisotropy in the lower layer, and a more northerly fast polarization. Overall, the lower layer is sub-parallel to the present-day transcurrent motion on the DSR, but also matches the typical orientation of lithospheric seismic anisotropy in the Arabian shield. Our overall conclusion is that the impact of the DSR on the rock fabric of the mantle lithosphere is probably quite weak. © 2006 Elsevier B.V. All rights reserved.

Keywords: Anisotropy, Lithosphere, Upper mantle, Body waves

1. Introduction

The complex structure of the continental lithosphere typically causes directional dependence of such basic physical properties as rigidity and electric conductivity. This anisotropy arises through a variety of mechanisms.

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Some act on the scale of mineral grains, while others emerge only when large volumes of rock are sampled. Proper understanding of anisotropic properties may be key to unraveling both tectonic history and present-day lithosphere deformation.

Most stable continental regions on Earth are characterized by anisotropy in seismic wavespeed (Silver, 1996; Park and Levin, 2002). The intensity of anisotropic indicators, e.g., roughly 1 s of shear wave birefringence in an average continent, precludes the affected rock from residing exclusively in the crust. Major candidate mech-

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Fig. 1. A regional map showing sites (triangles) where previous studies found fast polarizations of birefringent SKS phase to be approximately north–south. Included in the compilation are studies by Wolfe et al. (1999), Levin and Park (2000), Schwartz and Rodgers (2003), Schmid et al. (2004) and Hansen et al. (2005). Labels: DSR stands for Dead Sea Rift; MS for Mediterranean Sea. A box shows the outline of the map in Fig. 3.

anisms for anisotropy in the upper mantle are the rock fabric formed in the shear zone between the lithospheric plate and the underlying mantle, and the fabric that is preserved in the mantle lithosphere from past episodes of deformation. Both mechanisms have a significant body of evidence in their favor (Vinnik et al., 1992; Silver, 1996; Park and Levin, 2002). Electrical anisotropy can accompany wavespeed anisotropy if the aligned mantle minerals are bounded by thin conductive films, e.g., of graphite (Mareschal, 1995).

Given that seismic anisotropy is pervasive in the lithosphere of continents, its persistence is an important question. What processes can disrupt an anisotropic fabric? What tectonic regime would either erase or overwhelm a pre-existing fabric? If rock fabric, once created, is hard to disrupt, maps of rock fabric in mantle lithosphere should document major tectonic upheavals of the past. On the other hand, if new rock fabric easily develops an anisotropic signature, only the most recent episode will be important.

The Dead Sea Rift (Fig. 1) offers a convenient setting to probe the longevity of the mantle lithosphere fabric in the continent. The Dead Sea Rift (DSR), in spite of its name, is an active left-lateral transform boundary between the Arabian and African plates, with over 100 km of offset accumulated since ~ 15 Ma (Garfunkel, 1981). DSR cuts through the ~ 500 My old Arabian–Nubian shield (presently exposed on both sides of the Red Sea), and thus modifies lithosphere that has experienced numerous episodes of rock-fabric formation prior to the inception of the transform.

The main sense of motion on the Dead Sea Rift is transcurrent, but there is evidence that in the last 2–5 Ma a component of divergence has developed, leading to the formation of pull-apart basins within the transform, e.g., the Dead Sea and the Gulf of Aqaba (Garfunkel and Ben-Avraham, 1996).

In this paper we examine how seismic anisotropy is distributed with depth at a number of locations along the trace of the DSR. We find evidence for at least two distinct anisotropic layers on both sides of the rift. In the upper layer the fabric is relatively strong, and consistent in its orientation with the present-day transcurrent motion across the DSR. It is also consistent with the overall pattern of lithospheric seismic anisotropy throughout the Arabian platform. The strength of this fabric makes the lithospheric upper mantle a most likely locus for it. The lower layer of anisotropy has weaker fabric. Where well developed, the orientation of this fabric appears to agree with the deformation caused by the plate motion relative to the asthenosphere. Our findings lead us to question whether either of the observed fabrics reflect deformation associated with DSR activity. Instead, we propose an interpretation of the lithospheric component of anisotropy in terms of the inherited rock fabric.

2. Tectonic and geophysical background

The Arabian-Nubian shield that consolidated \sim 500 My ago (Stoeser and Camp, 1985) included the present-day Arabian plate as well as parts of Northern Africa that presently lie on the other side of the Red Sea. On the basis of geochemical data, Stein and Goldstein (1996) argued that the original lithospheric mantle of that shield had never separated from the overlying crust. Areas along the present-day eastern shore of the Mediterranean experienced a rifting episode and became a part of the passive margin in late Paleozoic time (Patton et al., 1994). Starting \sim 34 My ago, the shield has been subject to rifting along the present Red Sea and the Gulf of Suez (Omar and Steckler, 1995). Since ~ 15 Ma, extension in the Gulf of Suez has largely ceased, and active tectonics have shifted onto the Dead Sea Rift (Joffe and Garfunkel, 1987; Steckler et al., 1998; Ghebreab, 1998). This major transcurrent shear zone has since accumulated $\sim 105 \text{ km}$ of displacement (Garfunkel, 1981). Geomorphologic evidence suggests that since 2-5 My ago Dead Sea Rift may have acquired a trans-tensional component of motion which is thought

to be responsible for the opening of the pull-apart basins of the Dead Sea and in the Gulf of Aqaba (Garfunkel and Ben-Avraham, 1996). A present-day rate of relative transcurrent motion across the DSR is relatively low, with estimates based on GPS observations ranging between 1 and 7 mm/yr (Péeri et al., 2002; McClusky et al., 2003; Wdowinski et al., 2004). There is a disagreement between various GPS studies on whether there is also a measurable tensional component across the DSR. McClusky et al. (2003) posit the opposite, interpreting their data as suggestive of compression across the northern part of the DSR. Notably, the estimate of the total lateral displacement on the DSR, and its overall interpretation as a transform fault has its detractors. An alternative advocated recently by Mart et al. (2005) holds that the sinistral displacement on the DSR is small, and interprets the DSR as a nascent spreading center.

The Arabian plate is presently the smallest of Earth's cratonic areas, characterized by $\sim 200 \text{ km}$ thick lithosphere (Artemieva and Mooney, 2002). It is composed of the (western) Arabian Shield and the (eastern) Arabian Platform. The crust is generally thicker in the Platform than in the Shield (Mooney et al., 1985; Prodehl, 1985; Sandvol et al., 1998a,b; Rodgers et al., 1999; Kumar et al., 2002). In the DSR area the crust becomes progressively thinner from east to west, with reported values ranging from 33 to 39 km on the eastern side, and on the order of 25 km at the Mediterranean shoreline (Al-Zoubi and Ben-Avraham, 2002; Weber, 2004; Mohsen et al., 2005).

Studies of seismic anisotropy indicators carried out on the Arabian shield show a very consistent pattern. Overall, long-period shear waves tend to travel faster if they are polarized in the north–south direction (Wolfe et al., 1999; Levin and Park, 2000; Schwartz and Rodgers, 2003). Similar values are reported along the Dead Sea Rift (Ruempker et al., 2003; Ryberg et al., 2005; Schmid et al., 2004). However, some evidence points to the likely presence of multiple distinct anisotropic regions at depth (Levin and Park, 2000). Fig. 1 shows locations of sites where shear wave splitting studies find fast polarizations close to the north–south direction.

3. Methodology

To detect and characterize anisotropy of seismic wave propagation within the upper mantle beneath the DSR we use observations of birefringence (also known as splitting) of shear waves arriving from distant earthquakes. The shear–wave splitting technique is arguably the most popular among methods designed to detect upper-mantle anisotropy, due both to its ease of use, and to the lateral resolution it offers (Savage, 1999). The technique is based on the tendency of a shear wave to partition into two components as it propagates through an anisotropic medium. One component becomes polarized in the plane of the "fast" polarization of the anisotropic material, while the other component becomes polarized in the orthogonal plane (the "slow" polarization). As a consequence, a delay in arrival time develops between these components, proportional to the amount of anisotropy encountered along the path. The pair of values φ and $\Delta \tau$ define the polarization azimuth of the fast wave and the delay time between the fast and the slow polarizations and are known as "splitting parameters". Assuming an average anisotropy of 4% and a subvertical raypath, $\Delta \tau = 1$ s roughly corresponds to a 100 km thick anisotropic layer (Silver, 1996).

Core-refracted teleseismic shear waves (SKS, SKKS, PKS and similar phases) are useful for studies of the upper mantle since they are affected by anisotropy only on the "receiver-side" of their path. Their passage through the liquid outer core involves a conversion to compressional waves, and thus removes any "sourceside" birefringence signal. Furthermore, core-refracted waves have a known initial polarization. At the start of their ascent from the core-mantle boundary these phases are radially polarized, with particle motion in the vertical plane of the source and receiver. Finally, they ascend almost vertically through the mantle, and thus sample upper-mantle anisotropy at near-normal incidence.

On their near-vertical ascent through the upper mantle, SKS-type phases integrate anisotropic properties along their path, with little constraint on the depth interval in which anisotropy is present. In the case of a single homogeneous anisotropic region with either horizontal or vertical symmetry of the anisotropic elasticity tensor the birefringence of shear waves would not depend much on their propagation parameters. Since Earth structure most likely deviates from these restrictive conditions, a complex dependence of birefringence on the azimuth of wave propagation is expected (Savage, 1999). As a consequence, individual observations of birefringence become "effective" measures of anisotropy at depth. A number of strategies, all based on the study of directional variations in such measurements, have been explored in the literature (Silver and Savage, 1994; Rumpker and Silver, 1998; Levin et al., 1999; Hartog and Schwartz, 2000; Saltzer et al., 2000). Menke and Levin (2003) proposed a new technique that does not rely on effective measurements of anisotropy, but rather seeks to satisfy the data through explicit matching of observed and simulated seismograms. The technique depends critically on availability of high-quality data distributed in approach azimuth, and offers a numerical comparison of the quality of fit between various classes of anisotropic structures used in simulating synthetic seismograms. In this paper we use this last approach, together with a more traditional method of estimating splitting parameters, to derive depth-dependent distributions of anisotropic properties for a number of sites along the trace of the Dead Sea Rift.

4. Data and measurements

We selected 91 records of SKS phases recorded by seismic observatories of the GEOFON network (KSDI, MRNI, JER, EIL), and a portable observatory (HIT) operated by Lawrence-Livermore National Laboratory (Rodgers et al., 2003). Station locations are shown in Figs. 3 and 5, and in Table A1 of Appendix A. Observations were collected over a period spanning 8 years, from 1996 to 2004. However, very few events were recorded at more than two sites at once. Source parameters of all observations are given in Table A2 of Appendix A.

For each observation we determined birefringence parameters, fast polarization (φ) and delay ($\Delta \tau$), using a cross-correlation estimator described in Levin et al. (1999). This tool searches for a combination of the rotation and the delay that yields the most similar waveforms on both horizontal components. Fig. 2 shows examples of data and illustrates the measurement procedure. Note that phases of very similar quality observed from different back-azimuths yield significantly different estimates of fast direction. Results of all measurements are presented in Fig. 3 as black bars aligned with fast polarizations and scaled with delay. Positions of individual measurements on the map correspond to piercing points at 150 km depth for respective rays. The arrangement of individual measurements around their respective receivers illustrates data distribution in terms of their approach azimuths. Values of all individual measurements of φ and $\Delta \tau$, and their respective error estimates are given in Table A3 of Appendix A.

To explore depth dependence of anisotropic properties beneath our observing sites, we selected an azimuthally distributed subset of best-quality data for



Fig. 2. Examples of data used in the study. Two SKS phases recorded by station JER (Jerusalem) illustrate the single-event measurement procedure. (Left) Radial (solid) and transverse (grey) components of recorded waveforms. Data are bandpass filtered between 0.02 and 0.2 Hz. Event labels and back-azimuth values are noted on the plot (middle) Waveforms rotated into fast (grey) and slow (solid) components show close similarity of pulse shapes. Note different values for fast direction. (Right) Enlarged sections show ~ 0.5 s of time lag between fast and slow components for both observations.



Fig. 3. A topographic relief map of the study area showing measurements of shear wave birefringence arranged according to the geometry of observations. Individual measurements are plotted as bars aligned with the azimuth of fast polarization in the shear wave, and scaled by the delay value. Centers of the bars correspond to map projections of their rays' piercing points at 150 km depth.

each (entries marked by stars in Table A3 of Appendix A). Selection was based mainly on signal-to-noise ratio, with preference given to observations with simple pulse shapes and lower likelihood of contamination from other phases. We performed grid-search inversions for best-fitting models using cross-convolution technique of Menke and Levin (2003). We combined observa-

Table 1
Results of the inversion for one and two layer models

Station	One-layer		Two-layer				
	$\varphi\left(^{\circ} ight)$	$\Delta \tau$ (s)	<i>φ</i> 1 (°)	$\Delta \tau 1$ (s)	<i>φ</i> 2 (°)	$\Delta \tau 2$ (s)	
EIL	14.0	1.2	88.0	0.6	10.0	1.6	
HIT	19.0	1.3	98.0	0.7	18.0	1.8	
JER	12.0	1.2	94.0	1.8	10.0	2.9	
North	13.0	1.6	30.0	0.8	0.0	0.9	

Values for two layer models are those of the model with a smallest misfit value in the entire search. In the two-layer models layer 2 is on top. See Appendix A for the discussion of relative quality of these solutions.

tions from stations KSDI and MRNI into one group due to their geographical proximity. We explored two classes of models: one-layer models with a horizontal axis of symmetry, and two-layer models with horizontal axes of symmetry. Due to the inherent trade-off between anisotropic intensity and layer thickness, our models are described by their fast polarizations and delays in respective layers. In conducting the searches, we limited delay values within individual layers to 3 s. Fast polarization searches were conducted with a step of 2° between values of 0° and 180°.

Results of group inversions for each site are presented in Fig. 4. Full ranges of one-layer model parameters for each station are shown in the left column as shaded surfaces that quantify goodness of fit. Global minima on these surfaces are marked, and values are noted on the plots. For two-layer models we have combinations of four parameters (two fast polarizations and two delays) to search. On plots in the right column of Fig. 4 we show surfaces composed of the smallest misfit values found for corresponding combinations of fast polarizations. In other words, for each pair of fast polarizations we examine all possible delay pairs, identify the combination with the lowest misfit measure, and assign that value to the corresponding position on the plot. Two-layer misfit values that are smaller than the best corresponding one-layer model misfit are shaded.

The interpretation of a two-layer inversion requires care, because they can include layers with near-normal fast polarizations (e.g., HIT, see Table 1) in which delay times in one layer cancel the delay in the other. As Menke and Levin (2003) show, shear wave splitting data with signal from one layer of anisotropy can be fit as well, or

Fig. 4. Results of group inversions for one-layer (left) and two-layer (right) structures. (Left column) Misfit value surfaces for one-layer group inversions. Global minima of the surfaces are marked by white stars, and corresponding values are noted on the plot. (Right column) Plots of best (smallest) misfit values for combinations of fast directions in two layers. Values less then those of a global minimum for a corresponding one-layer model are shaded. Green stars mark the smallest value of misfit found in a search of two-layer model parameters (see Table 1 for specific values). Lines show the fast direction value of the one-layer model. Noted on the plots are ranges of values for two-layer models that fit data better then corresponding one-layer solutions. See text for a description of how these models are selected.





Fig. 5. Parameters of final one-layer (left) and two-layer (right) models of seismic anisotropy derived for individual sites are shown as bars aligned with fast polarization within a layer and scaled with the amount of delay this layer contributes. For one-layer models, values of the model with the smallest misfit are shown. For two-layer models, two bars are shown per layer to delimit a range of values that yield similar solutions. Parameters of the lower layer are depicted by wider darker bars. Open arrows illustrate various models of plate motion in the region. Plate motion model values are taken from the UNAVCO plate motion calculator at http://sps.unavco.org/crustal_motion/dxdt/model. Absolute plate motion (model NUVEL1A in no-net-rotation frame) is shown on the right panel. Relative plate motion between Africa and Arabia is shown by red arrows on the left panel (values from Global Strain Rate Map project, Kreemer et al., 2003). The direction of relative motion across the DSR from continuous GPS observations (Prawirodirdjo and Bock, 2004) is shown by green arrows. This later estimate takes into account independent motion of the Sinai block. Red arrows on both plots show motion on the order of 30 mm/yr, while green arrows (GPS-derived motion) show direction of motion on the order of 1 mm/yr.

even better, by a two-layer model where one of the layers has "true" fast polarization direction, and the other is near-normal to it. Delay values in such layers suffer complete trade-off, i.e., an increase in one layer's delay can be compensated by an increase in the other layer's delay by the same value. In practice data inversion models with 2 (or more) "crossed" layers may fit stochastic noise in the observations better, but not improve the fit in a statistically significant manner. In view of the above complication, a full range of solutions for the two-layer model search needs to be examined, and solutions of the "crossed" type must be excluded. For inversions of DSR data we assigned models to a "crossed" type if the acute angle between fast directions exceeded 80°. Ranges of model parameters (both fast polarizations and delays) that fit selected data groups better than corresponding one-layer models are noted within plots, and also depicted in Fig. 5.

5. Results

Most SKS phases observed along the trace of the Dead Sea Rift are birefringent. When analyzed with a single-event splitting estimator, they yield $\Delta \tau$ values ranging from near zero to almost 2 s. Values of φ close to north-south alignment predominate. At all sites φ varies with the azimuth of wave propagation. At the same time, for individual events we find considerable similarity in splitting parameters at our stations (e.g., event 2001.013.17 yields fast polarizations of 121.95, 120.25 and 118.29 at EIL, JER and MRNI, respectively). Thus, two noticeable features of our observations are their overall similarity over the entire region, and the systematic dependence of φ and $\Delta \tau$ values on the arrival direction (back-azimuth) of the wave. This quasi-periodic variation in splitting parameters is documented for all observing sites in Fig. A1 in Appendix A. Obviously, values of φ and $\Delta \tau$ obtained for individual phases are not representative of the structure at depth, but rather reflect a path-averaged effect.

A periodic variation in splitting parameters is expected if the anisotropic medium is horizontally stratified (Silver and Savage, 1994; Rumpker and Silver, 1998; Levin et al., 1999; Saltzer et al., 2000), or else if the symmetry axis of anisotropy is uniformly inclined throughout the region (Plomerova et al., 1998; Hartog and Schwartz, 2000), although the rate of variation is different (90° and 180° for layering and axis dip, respectively). Variation observed in the DSR data set clearly follow the 90° periodicity, motivating the exploration of layered models.

In group inversions of data subsets that solve for parameters of a single layer of anisotropy we obtain remarkably similar models at all sites, with fast polarizations that strike 12-19° CW from north and delays of roughly 1.3s (Table 1, Fig. 4 (left)). At all sites inversions for two-layer models of anisotropy return goodness-of-fit measures that improve upon those achieved by one-layer models. The preference for a two-layer solution is stronger (i.e., fit measures improve more) for station JER, and for the northern group of observations (sites KSDI and MRNI). A description of the procedure used to assess the relative quality of inversion results is provided in Appendix A. In all cases except EIL the statistical confidence that the splitting parameters of the second layer are non-random exceeds 99%. For EIL the confidence level for nonrandomness is 93.7%. The backazimuthal variation of fast polarization for a typical station (Fig. A1) exhibits clearly the behavior of multiplelayer anisotropy.

Two-layer models selected for sites EIL, JER and HIT are very similar. The lower layer has small (0.3–0.6 s) delay time and fast polarization in the range $50-80^{\circ}$. The upper layer has higher delay time (1–1.5 s) and fast polarization sub-parallel to north–south. For the northern duo of stations, parameters of the upper layer are similar to those found elsewhere (fast polarization within a few degrees of north, delay ~1 s), but the lower layer has higher delay (0.6–1.0 s) and fast polarization in the range 25–35 °CW from north.

6. Discussion

Our measurements of shear wave birefringence parameters at sites along the trace of the DSR agree reasonably well with previous efforts. For instance, the SKS phase from a $M_w = 7.6$ earthquake on 28 March 2000 was observed by a dense array of short-period geophones crossing the Dead Sea Rift (Ruempker et al., 2003; Ryberg et al., 2005). The fast polarization $(0-15^{\circ})$ and delay of 1.2–1.5 s reported by Ruempker et al. (2003) are close to the splitting parameters we find for the same earthquake at EIL ($\varphi = 18^\circ$, $\Delta \tau = 1.25$ s). However, our estimate of birefringence parameters for the same phase observed further north at MRNI yields $(\varphi = 29.3^{\circ}, \Delta \tau = 1.03 \text{ s})$. This is consistent with our overall finding that birefringence at the northern end of the DSR differs somewhat from the more homogeneous signature further south.

In a study of shear wave splitting throughout the Mediterranean, Schmid et al. (2004) reported coherent values of shear wave splitting for sites EIL, JER and MRNI. At all three sites they find an average fast polarization $3-8^{\circ}$ east of north, and a delay near 1.3 s. While no individual measurements are reported by Schmid et al. (2004), their averaged values agree very well with estimates we obtained from group inversions for onelayer models (Figs. 4 and 5). Another probe for seismic anisotropy relies on measurements of travel time of the Pn phase. A recent study by Al-Lazki et al. (2004) includes the northern part of the region we study. The orientation of fast Pn propagation in the northern part of the DSR is highly oblique to the strike of the fault, and thus disagrees with what we and others observe with birefringent shear waves. However, such discrepancies are not uncommon (e.g., Park et al., 2004). Pn phases sample a relatively narrow depth range close to the crust-mantle interface, and integrate laterally. SKS phases, on the other hand, integrate a near-vertical path through the upper mantle, and may be weakly sensitive to the source of Pn anisotropy.

In our study of birefringence in core-refracted shear waves we have examined a larger data set (91 observations) than used in previous work. We paid particular attention to developing sets of measurements that would be distributed in back-azimuth. We are thus able to identify and compare patterns of azimuthal variation of birefringence along the trace of the DSR. Our main finding, illustrated in Figs. 3 and A1, is that these patterns are consistent with a two-layer anisotropic model throughout the region. Three of the four sites (EIL, JER and the KSDI/MRNI combination) are proximate to the DSR, and thus should feel the influence of the deformation associated with it. The fourth site, HIT, is located \sim 50 km from the DSR. Nearby investigations of the DSR structure with active source methods (Weber, 2004), as well as the study of a split shear wave observed on a dense array (Ryberg et al., 2005) both suggest that entire zone of deformation associated with the DSR is no wider than 20 km. We can therefore consider results for HIT to be representative of the Arabian shield lithosphere that was not involved in DSR deformation. If we interpret similarity in back-azimuthal patterns of shear wave birefringence to reflect a similarity in depth structure, the inference would be that anisotropic structure varies little beneath the DSR and away from it. This notion is further supported by the fact that observations from site EIL sample both sides of the DSR (see Fig. 3), and their pattern is not different from those of either HIT on the west flank or JER on the east flank.

On the basis of single-phase measurements of shear wave birefringence we documented a similarity of anisotropy-inducing fabric along the trace of the DSR, and also away from it. Given previous studies (e.g., Schmid et al., 2004) this is not surprising, except for the directional changes in birefringence documented throughout the region. Levin et al. (2000a) documented a similar scenario of a region-wide coherent directional pattern of birefringence parameters in northeastern North America. There a model composed of two distinct layers of anisotropic properties successfully captured the observed behavior. In developing models for depthdependent anisotropy, we tested one- and two-layer models with horizontal symmetry axes. Given the tectonic history of the region as a stable continent, and later as a passive margin, we feel that a vertically stratified structure of anisotropic properties is a reasonable first guess. Also, the nature of the observed pattern of shear wave birefringence (see Fig. A1) appears consistent with vertical layering. Values of fast polarization display periodic change with back-azimuth with an approximate period of 90°. Such a pattern is well recognized as diagnostic of some form of fabric stratification (Silver and Savage,

1994; Rumpker and Silver, 1998; Levin et al., 1999; Saltzer et al., 2000; Menke and Levin, 2003). Alternatives, e.g., a regionally consistent plunge of the symmetry axis (e.g., Plomerova et al., 1998; Hartog and Schwartz, 2000) would yield different patterns of directional variation.

One-layer solutions determined in our study (Table 1, Figs. 4 and 5) are very similar, and thus confirm our notion of the overall similarity of anisotropic structure at depth. At all sites the orientation of fast shear wave polarization inferred assuming there is only one source of anisotropic signal is very close to the north-south direction. Models with two layers yielded significantly better fits to data at all sites (see Table 1), particularly at JER and at the northern sites KSDI and MRNI. Once again, at all four sites the models chosen in the search are very similar. Top layers have fast polarizations close to north-south direction, and prescribe >1-s of delay time. In models for EIL, HIT and JER, the lower layers have northeastern fast polarizations (35-80°NE), and prescribe a relatively small (0.3-0.6 s) delay. For the northern site the lower layer of the model has more northerly orientation of fast polarization (25-35°NE), and prescribes a delay of 1 s. It should be noted that differences between top-layer fast orientations and those found for one-layer models are small, implying that improvement in data fit comes largely from including the contribution of the lower layer.

To interpret the meaning of layered anisotropic models it is useful to consider possible sources of anisotropic fabric. Given the size of measured delays (over 2 s in individual phases) we are confident that causes of this signal reside in the mantle, either within the lithosphere or beneath it. Because the region has formed from oceanic plateau accretion (Stoeser and Camp, 1985), and has been a stable continent for over 500 My, it is likely that rules of fabric formation in the lithosphere are those appropriate for dry olivine (e.g., Ribe, 1992; Zhang and Karato, 1995). Similarly, as no obvious sources of volatiles are present in the region, sublithospheric mantle is likely to be dry, obviating the deformation mechanisms reported by Jung and Karato (2001). Therefore, we can relate fast polarization of shear waves to the preferred alignment of a-axes of olivine grains in the mantle rock. In turn, this alignment should be subparallel to the direction of flow in the mantle, whether the deformation is fossil or ongoing.

The presence of two layers of anisotropy in stable continental environment argues that the top layer is located in the mantle lithosphere (cf. Levin et al., 2000a,b) while the lower one may be related to the relative motion between the lithosphere and the asthenosphere. This latter interpretation appears reasonable in view of the close alignment between the lower-layer fast polarization at the three northern sites (JER, KSDI, MRNI) and the direction of absolute plate motion (Fig. 5 (right)). At two southern sites (EIL, HIT) the second layer is defined with less confidence (see Table A4 in Appendix A), and interpreting its meaning is harder. We note that the nature of the plate motion in this part of the region is especially complex, with potential independent motion of the Sinai block, and possible local flow regime in the upper mantle (Steckler et al., 1998). Our results for the northern part of the DSR appear to rule out a model of mantle dynamics that involves upward motion of asthenospheric material beneath the DSR, as advocated by Rabinowitz and Mart (2000) and Mart et al. (2005). It would not be consistent with either the directions or the uniform nature of the sublithospheric fabric implied by our results. We cannot make an equally clear determination for the southern part of the DSR.

The mechanism responsible for the upper layer of anisotropy is less clear. Obviously, the polarization of fast shear wave propagation in the region is very close to the strike of the DSR. Previously this similarity was interpreted as evidence for fabric associated with transcurrent deformation on the DSR (Schmid et al., 2004). However, we find that station HIT, which is farther from the DSR, also has the same fast polarization. Results of the DESERT project (Ruempker et al., 2003; Ryberg et al., 2005; Weber, 2004) suggest that the influence of the DSR is concentrated in a very narrow (20 km or so) zone, and thus would not extend to the location of HIT. Furthermore, at locations further east, within the Arabian shield, fast shear-wave polarization is also nearly north-south (Wolfe et al., 1999; Levin and Park, 2000; Schwartz and Rodgers, 2003).

Anisotropy in subsurface resistivity bears an interesting relation to both seismic anisotropy and tectonic trends, but electrical anisotropy induced by the penetration of saline groundwater into Dead Sea Rift structures is a confounding factor. Ritter et al. (2003) analyze a magnetotelluric (MT) transect across the Dead Sea Rift midway between stations JER and EIL. The MT data indicate a geoelectric strike 17° east of north near the rift, similar to shear-wave fast polarization in the upper layer of our models. Their analysis is restricted to periods <100 s, limiting resolution to the shallow crust. Low apparent resistivity in the TM mode for stations east of the DSR was interpreted in terms of conductive sediments in the crust west of the rift. A possible connection to seismic anisotropy is unclear, because graphite along olivine grain boundaries in the Mareschal (1995) model for electrical anisotropy should induce lower resistivity

in the TE mode. Ritter et al. (2003) concluded that the rift is a structural barrier to lateral fluid flow in the upper crust. In Northern Israel and the Golan Heights region, near seismic stations KSDI and MRNI, Rotstein and Goldberg (1981) determined that the TE mode was less resistive than the TM mode and showed for some stations that large differences in apparent resistivity occurred for EM wave periods approaching 800 s. Low resistivity parallel to the Dead Sea Rift would parallel the shear–wave fast polarization and support the Mareschal (1995) mechanism, but the periods analyzed were too short to offer significant penetration of the lithospheric mantle, which requires 500 s < T < 10000 s (Bahr and Duba, 2000).

Relative motion between two sides of the DSR. and the associated deformation at depth, are presently debated. Fig. 5 (left) shows directions for relative plate motion (Africa-Arabia) determined in the framework of the Global Strain Rate Map (Kreemer et al., 2003). Clearly, a divergence across the DSR is implied. On the other hand, examination of the regional deformation field using continuous GPS measurements at sites on both sides of the DSR led Wdowinski et al. (2004) to conclude that the rate of left-lateral motion is low, not exceeding 4 mm/yr. This latter study assumed an independent motion of the Sinai plate. Similarly, a global model of crustal velocity based on GPS observations (Prawirodirdjo and Bock, 2004) yields ~1 mm/yr of motion across the DSR, with motion direction roughly along its strike (green arrows in Fig. 5 (left)).

The left-lateral motion on the DSR, inferred from both geology and continuous GPS, is likely to form a sub-horizontal fabric in the lithosphere consistent with observed shear wave birefringence.

However, the presence of a similar birefringence signature away from the DSR poses a question in the spirit of Occam's Razor—is it a necessary interpretation? We favor the minimalist alternative: the lithosphere of the Arabian platform is characterized by a fossil north–south anisotropic fabric that dates from the time of its assembly.

Acknowledgements

VL and JP were supported by NSF grant EAR0208031. Data from the GEOFON network was acquired via the Data Management System of IRIS. Figures were produced using GMT software (Wessel and Smith, 1991). This work was performed in part under the auspices of the U.S. Department of Energy by University of California Lawrence Livermore National Laboratory under contract no. W-7405-Eng-48. This is LLNL contribution UCRL-JRNL-213087.

Table A2 (Continued)

Table A1 Seismic stations used in the study

Name	Latitude (°)	Longitude (°)
EIL	29.67	34.95
HIT	29.74	35.84
JER	31.77	35.19
KSDI	33.19	35.66
MRNI	33.12	35.39

Table A2

Earthquakes used in this study

Event	Latitude (°)	Longitude (°)	Depth (km)
1996.162.15	51.48	-176.85	26.00
1996.197.21	17.60	-100.96	18.00
1997.011.20	18.22	-102.76	33.00
1997.142.08	18.68	-101.60	70.00
1997.168.21	51.35	-179.33	33.00
1997.222.09	-16.01	124.33	10.00
1997.301.06	-4.37	-76.68	112.00
1997.332.23	-13.74	-68.79	586.00
1997.356.02	-5.50	147.87	179.00
1998.010.08	14.37	-91.47	33.00
1998.093.22	-8.15	-74.24	165.00
1998.142.05	-17.73	-65.43	24.00
1998.216.19	-0.59	-80.39	33.00
1998.232.15	51.62	175.25	33.00
1998.235.14	11.66	-88.04	55.00
1998.242.01	17.09	148.13	33.00
1998.245.08	5.41	126.76	50.00
1998.264.06	0.26	122.47	147.00
1998.281.05	-16.12	-71.40	136.00
1998.300.21	2.92	128.62	61.00
1998.301.16	0.84	125.97	33.00
1998.312.07	-9.14	121.42	33.00
1998.350.17	1.12	126.18	33.00
1999.079.11	51.59	-177.67	33.00
1999.090.06	5.83	-82.62	10.00
1999.093.06	-16.66	-72.66	87.00
1999.136.01	-4.75	152.49	74.00
1999.137.10	-5.17	152.88	27.00
1999.166.21	18.39	-97.44	70.00
1999.192.14	15.78	-88.33	10.00
1999.258.03	-20.93	-67.28	218.00
1999.273.16	16.06	-96.93	61.00
1999.289.10	34.59	-116.27	0.00
1999.291.03	-56.12	-26.58	33.00
1999.334.04	-18.90	-69.17	128.00
2000.088.11	22.34	143.73	127.00
2000.109.00	-52.46	13.54	10.00
2000.125.04	-1.10	123.57	26.00
2000.166.17	4.54	127.72	90.00
2000.168.08	-33.88	-70.09	120.00
2000.207.03	-53.55	-3.17	10.00
2000.219.07	28.86	139.56	395.00
2000.234.09	-53.02	-45.97	10.00
2000.272.23	-0.21	-80.58	22.00
2001.010.16	57.08	-153.21	33.00

Event	Latitude (°)	Longitude (°)	Depth (km)
2001.013.17	13.05	-88.66	60.00
2001.055.07	1.27	126.25	35.00
2001.059.19	47.15	-122.73	51.00
2001.103.15	-59.72	-25.59	26.00
2001.177.04	-17.75	-71.65	24.00
2001.182.02	-4.31	152.96	28.00
2001.346.14	-42.81	124.69	10.00
2002.010.11	-3.21	142.43	11.00
2002.087.05	-21.66	-68.33	125.00
2002.116.16	13.09	144.62	85.00
2002.158.00	-0.88	148.33	10.00
2002.164.01	-47.80	99.75	10.00
2002.167.03	8.78	-83.99	35.00
2002.226.14	14.10	146.20	30.00
2002.251.19	-3.30	142.95	13.00
2002.346.08	-4.79	153.27	34.00
2002.351.04	-56.95	-24.83	10.00
2003.021.03	13.63	-90.77	24.00
2003.070.07	-4.69	153.24	40.00
2003.076.16	51.27	177.98	33.00
2003.174.12	51.44	176.78	20.00
2004.028.22	-3.12	127.42	17.00
2004.036.21	-3.62	135.54	16.00
2004.039.09	-3.67	135.34	25.00
2004.124.05	-37.70	-73.41	21.00
2004.180.10	54.80	-134.25	20.00
2004.250.13	-55.37	-28.98	10.00

Appendix A

Coordinates of stations and earthquakes used in this study are presented in Tables A1 and A2. Results of measurements of birefringence parameters using the cross-correlation technique of Levin et al. (1999) are shown in Table A3.

A.1. Statistics of data fit

To assess the quality of data fit for a given anisotropic model \mathbf{m} we compute a quantity $E(\mathbf{m})$ that evaluates a fit between an ensemble of N observed and predicted time series. As discussed in Menke and Levin (2003), this quantity is computed as follows

$$E(\mathbf{m}) = \left(\frac{1}{N}\right) \Sigma_{i=1}^{N} ||x_i(t) - y_i(t)||^2,$$

where

$$||x_{i}(t) - y_{i}(t)||^{2} = \frac{\int [x_{i}(t) - y_{i}(t)]^{2} dt]}{\int x_{i}^{2}(t) dt + \int y_{i}^{2}(t) dt}$$

$$x_{i}(t) = h^{\text{pre}}(\mathbf{m}, t) * V_{i}^{\text{obs}}(t),$$

$$y_{i}(t) = v^{\text{pre}}(\mathbf{m}, t) * H_{i}^{\text{obs}}(t)$$

Table A3 Splitting measurements using a cross-correlation algorithm of Levin et al. (1999)

Station	Event	BAZ ($^{\circ}$)	POL (°)	FAST (°)	DELAY (s)	Fast_er (°)	Del_er (s)	Xcor
EIL*	1997.301.06	277.12	99.14	143.39	0.21	4.60	0.09	0.976
EIL	1997.332.23	264.42	91.52	48.40	0.24	4.81	0.10	0.926
EIL^*	1998.010.08	302.70	124.59	174.21	0.81	8.98	0.18	0.978
EIL^*	1999.028.08	14.30	197.83	152.64	0.10	1.99	0.01	0.992
EIL	1998.093.22	272.32	94.66	50.41	0.21	4.38	0.09	0.974
EIL^*	1998.216.19	282.65	108.22	63.50	0.24	5.16	0.11	0.984
EIL	1998.281.05	263.36	91.18	48.00	0.38	5.72	0.18	0.936
EIL^*	1999.093.06	263.40	85.86	41.56	0.27	5.10	0.11	0.961
EIL.	1999 258 03	256.89	79.69	35.44	0.71	4 79	0.09	0.973
EIL	1999.273.16	307.95	134.63	176.27	0.79	5.34	0.21	0.958
EIL	1999 291 03	209.88	209.08	166.81	0.74	4.86	0.04	0.980
EIL*	1999.334.04	259.67	79.62	35.31	0.34	4.03	0.08	0.975
EIL*	2000 088 11	61 39	64 19	17.80	1.25	10.00	0.12	0.931
FII	2000.125.04	91.64	94.22	53 30	0.65	7.65	0.12	0.930
FII	2000.125.04	84.68	80.80	45 37	0.05	5.85	0.09	0.950
EIL *	2000.168.08	244.78	70.40	28.42	0.40	5.02	0.11	0.907
EIL *	2000.108.08	4 00	8.06	20.42	0.05	5.02	0.00	0.887
	2000.195.01	4.99	121.05	166.00	0.55	5.12	0.09	0.97
EIL EIL*	2001.015.17	299.12	121.93	100.90	0.74	5.12	0.07	0.994
EIL EU *	2001.055.07	00.23 244.76	90.24	43.00	0.57	5.75	0.08	0.970
EIL FIL *	2001.059.19	344.70	107.48	28.05	1.15	0.87	0.09	0.968
EIL FIL *	2001.346.14	128.91	133.15	-1.02	0.69	5.05	0.17	0.950
EIL	2002.010.11	84.02	83.85	39.33	0.95	3.97	0.08	0.963
EIL	2002.087.05	256.62	80.09	35.10	0.56	4.55	0.11	0.943
EIL	2002.116.16	68.62	70.68	20.51	0.93	11.74	0.31	0.903
EIL'	2002.316.02	209.86	32.23	5.00	1.00	32.69	0.74	0.994
EIL*	2003.174.12	22.67	201.76	156.95	0.42	5.08	0.07	0.981
HIT	2000.109.00	193.43	23.86	65.93	0.07	6.39	0.25	0.902
HIT	2000.207.03	201.96	190.28	148.02	0.16	5.99	0.19	0.911
HIT	2000.219.07	58.46	55.49	11.72	1.08	4.83	0.05	0.938
HIT	2000.234.09	218.97	214.44	170.85	0.83	7.31	0.10	0.961
HIT	1998.242.01	63.81	56.04	13.44	0.24	6.71	0.26	0.915
HIT	1998.245.08	84.85	92.77	27.83	1.15	13.49	0.48	0.986
HIT	1998.264.06	91.45	72.64	22.51	1.24	6.50	0.12	0.907
HIT	1998.300.21	86.09	101.02	53.09	0.47	8.47	0.09	0.931
HIT	1998.301.16	89.21	93.14	20.02	1.29	6.49	0.35	0.970
HIT	1998.312.07	100.12	111.15	154.22	0.10	5.61	0.06	0.949
HIT	1998.350.17	88.86	105.72	63.18	0.35	5.65	0.07	0.916
JER [*]	1996.162.15	19.31	15.61	57.24	1.10	6.46	0.26	0.920
JER	1996.197.21	313.16	142.09	4.77	1.32	5.66	0.11	0.957
JER	1997.011.20	315.05	145.29	17.11	1.11	12.05	0.28	0.985
JER [*]	1997.142.08	314.44	143.44	7.39	1.07	6.08	0.09	0.937
JER*	1997.168.21	20.74	202.65	157.44	0.45	6.54	0.05	0.988
JER [*]	1997.222.09	104.14	109.04	152.67	0.43	5.33	0.07	0.963
JER	1997.301.06	278.05	109.13	153.48	0.09	5.98	0.12	0.978
JER*	1997.356.02	82.52	87.97	44.47	0.51	5.87	0.16	0.964
JER [*]	1998.010.08	303.57	128.43	175.42	0.81	7.17	0.14	0.975
JER*	1998.142.05	259.92	89.66	45.36	0.62	3.96	0.06	0.970
JER*	1998.216.19	283.63	113.56	157.62	0.03	5.37	0.11	0.985
JER	1998.232.15	23.50	207.26	163.52	0.08	4.73	0.10	0.956
JER	1998.235.14	299.02	124.91	168.54	0.96	5.46	0.06	0.977
JER	1998.301.16	88.88	150.46	20.85	2.73	7.24	0.21	0.944
JER	1999.079.11	19.71	184.89	142.90	0.04	7.46	0.30	0.900
JER	1999 090 06	290.60	114 94	158.83	0.31	4 99	0.10	0.968
JER	1999,093.06	264 42	91.75	46.70	0.52	4.84	0.09	0.967
IER*	1999 289 10	335 52	152 79	17 12	1.00	6.73	0.17	0.950
JER	1999,291,03	210.18	32.42	-2.18	1.06	13.24	0.23	0.978
IFR	2000 272 23	284.08	105 27	61.80	0.11	476	0.14	0.976
IFR [*]	2001.013.17	300 56	120.25	165 35	0.56	5.95	0.07	0.040
JUIN	2001.013.17	500.50	120.23	105.55	0.50	5.75	0.07	0.775

Table A3 (Continued)

Station	Event	BAZ ($^{\circ}$)	POL (°)	FAST (°)	DELAY (s)	Fast_er (°)	Del_er (s)	Xcor
JER*	2001.059.19	345.02	172.72	30.77	1.01	10.40	0.17	0.967
JER*	2001.103.15	207.00	209.04	174.66	0.86	16.58	0.29	0.968
JER	2001.177.04	262.89	89.61	42.06	0.55	6.16	0.09	0.987
JER [*]	2001.182.02	78.43	86.34	39.70	0.53	6.93	0.07	0.951
JER	2002.158.00	78.10	78.95	33.38	0.78	6.08	0.15	0.929
JER*	2002.167.03	293.97	115.40	157.91	0.72	6.39	0.07	0.991
JER	2002.346.08	78.68	87.32	36.24	0.63	14.62	0.18	0.972
JER	2002.351.04	208.89	210.16	174.83	1.14	14.30	0.29	0.973
JER	2003.021.03	302.49	126.46	169.04	0.47	5.78	0.11	0.934
KSDI*	2002.164.01	142.26	143.70	11.51	1.51	4.88	0.09	0.928
KSDI [*]	2002.226.14	66.75	60.45	0.82	2.15	5.60	0.22	0.912
KSDI*	2002.251.19	83.17	85.85	39.28	0.72	7.81	0.13	0.978
KSDI	2002.351.04	209.22	204.48	165.03	1.17	12.29	0.17	0.989
KSDI*	2003.070.07	78.23	206.66	155.69	2.39	8.29	0.09	0.926
KSDI*	2003.076.16	22.49	202.02	156.29	0.26	5.23	0.04	0.967
KSDI*	2003.174.12	23.04	195.04	147.81	0.40	8.81	0.06	0.992
KSDI	2004.028.22	91.66	107.48	151.42	0.02	7.68	0.11	0.962
KSDI [*]	2004.036.21	87.61	91.81	47.96	0.18	5.01	0.06	0.965
KSDI	2004.039.09	87.77	95.37	54.45	0.70	7.57	0.20	0.888
KSDI [*]	2004.124.05	243.66	61.02	15.26	1.07	7.47	0.16	0.932
KSDI*	2004.180.10	354.20	168.92	7.08	2.05	29.20	2.78	0.973
KSDI*	2004.250.13	212.00	207.93	168.64	1.45	8.92	0.17	0.981
MRNI	1999.079.11	19.81	144.40	5.16	0.38	8.75	0.31	0.867
MRNI	1999.136.01	78.62	221.83	177.47	0.53	7.08	0.16	0.928
MRNI	1999.137.10	78.75	44.06	4.09	0.29	7.01	0.19	0.824
MRNI	1999.166.21	311.54	95.41	141.47	2.04	6.68	0.23	0.918
MRNI	1999.192.14	303.00	75.66	125.20	1.10	7.76	0.16	0.965
MRNI	1999.273.16	309.48	68.44	113.07	0.99	4.85	0.14	0.976
MRNI	2000.088.11	61.47	80.74	29.33	1.03	9.78	0.16	0.952
MRNI*	2001.010.16	4.66	188.76	145.07	0.16	6.94	0.16	0.954
MRNI [*]	2001.013.17	301.11	118.29	165.72	0.73	8.64	0.15	0.990
MRNI [*]	2001.177.04	263.56	90.95	45.86	0.20	5.19	0.09	0.980
MRNI	2002.010.11	83.26	86.76	43.63	0.29	4.68	0.11	0.951

Polarization of the particle motion "corrected" for the effect of birefringence is given in the fourth column. Stars denote events used in group inversions, except for site HIT where all observations were used for the group solution. Error estimates for individual measurements are derived using the shape of the cross-correlation surface, see Levin et al. (1999) for description.

 $V_i^{\text{obs}}(t)$ and $H_i^{\text{obs}}(t)$ are radial and tangential components of *i*th observation, $v^{\text{pre}}(\mathbf{m}, t)$ and $h^{\text{pre}}(\mathbf{m}, t)$ the radial and tangential impulse responses of the model \mathbf{m} , and the symbol (*) denotes convolution in the time domain. $E(\mathbf{m})$ behaves as a misfit criterion because an input SKS pulse convolved with $h^{\text{pre}}(\mathbf{m}, t)$ and $v^{\text{pre}}(\mathbf{m}, t)$ will be invariant to the order of convolution. If $P_i(t)$ is the incoming pulse for the *i*th observation, then $x_i(t) = h^{\text{pre}}(\mathbf{m}, t) * V_i^{\text{obs}}(t) = h^{\text{pre}}(\mathbf{m}, t) * v^{\text{pre}}(\mathbf{m}, t) * P_i(t) = v^{\text{pre}}(\mathbf{m}, t) * P_i(t) = v^{\text{pre}}(\mathbf{m}, t) * H_i^{\text{obs}}(t) = y_i(t)$ if the splitting model is an exact representation of the data. The denominator of $E(\mathbf{m})$ normalizes the misfit variance to equalize the influence of large and small SKS amplitudes.

To determine, given models $\mathbf{m_1}$ and $\mathbf{m_2}$, the degree to which the value of $E(\mathbf{m_1})$ is better than $E(\mathbf{m_2})$ we apply

Table A4

Data fit measures for one- and two-layered models (E1 and E2, respectively), degrees of freedom (M) and F-test critical values

Station	<i>E</i> 1	<i>E</i> 2	М	F-test variance ratio for second layer	Confidence level for nonrandomness (%)
EIL	0.165137	0.158250	134	2.83	93.7
HIT	0.193616	0.178350	129	5.35	99.4
JER	0.191648	0.161068	104	9.49	99.98
North	0.299110	0.268264	91	5.00	99.1



Fig. A1. Values of birefringence estimates for individual phases plotted as a function of event back-azimuth. Observations for each station are given distinct symbols: EIL (square); JER (circle); HIT (upright triangle); KSDI/MRNI combination (inverted triangle). Errors are from the cross-correlation algorithm of Levin et al., 1999. Fast polarization values are shown in (a) and delay values are shown in (b). Note the smooth and rapid change in value of fast polarization as a function of back-azimuth, an effect that arises from complex anisotropic structure at depth.

an *F*-test, using the following logic to assess the number of degrees of freedom *M* in the data set:

M = NKB.

Here N is the number of observations, K the number of samples in the observed seismogram, and

B is the bandwidth of the dominant signal in percent of the Nyquist bandwidth of the record. For a data set of 10 seismograms composed of 1000 samples each, sampled 10 times/s, and containing signals with a dominant period of 10 s, the quantity $M = NKB = 10 \times 1000 \times 0.1/(1/(2 \times 0.1)) = 200.$

If records used in the measurement have uneven lengths K_i (as is the case in this study), we define $M = \sum_{i=1}^{N} (K_i B)$ instead.

The number of retrieved model parameters is 2 for a one-layer model and 4 for a two-layer model. The significance of the model parameters can be assessed by comparing the improvement of the misfit criterion $E(\mathbf{m})$ with respect to a less complex model. Since splitting is a clear feature of the core-refracted phases in our data set, we focus on comparing the one-layer misfit $E(\mathbf{m}_1)$ with the two-layer misfit $E(\mathbf{m}_2)$.

If the one-layer model is correct, the misfit $E(\mathbf{m}_1)$ should follow a chi-squared distribution with M-2 degrees of freedom. If the two-layer anisotropic model is inappropriate for the data set, the variance per degree of freedom that the additional parameters explain $((E(\mathbf{m}_1) - E(\mathbf{m}_2))/2)$ should be statistically indistinguishable from the variance per degree of freedom in the residual misfit $(E(\mathbf{m}_2)/(M-4))$. To test for nonrandomness in the partition of variance we apply the *F* variance-ratio test with 2 and M-4 degrees of freedom. If the variance explained by the splitting parameters of the second anisotropic layer is too large to be explained by random fluctuations, we conclude that the two-layer model is preferred.

Table A4 documents the basis for significance level assessment for inversions performed in this paper. Inspection of data spectra showed that dominant bandwidth is 0.1 Hz for stations EIL, JER, KSDI and MRNI, and 0.2 Hz for station HIT. Sampling interval was 0.1 s for all sites. All sites except EIL indicated that the variance explained by the second layer of anisotropy was nonrandom with >99% confidence. For station EIL the variance explained by the second layer of anisotropy was nonrandom with >93% confidence..

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