



Influence of observed mantle anisotropy on isotropic tomographic models

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[1] There is a potential for isotropic tomographic models to be biased by ignored anisotropy; anisotropy could be mapped as artificial isotropic velocity perturbations. We investigate the distribution and strength of this potential bias for three tomographic *S* velocity models based on regional *S* and Rayleigh waveforms. We use observed *SKS* delay times and fast-axis orientations to compute equivalent *S* velocity perturbations for each wave path used for the tomographic models. These synthetic perturbations are then combined into a three-dimensional isotropic model which reflects the potential bias in the actual tomographic model. We quantify anisotropic bias and find that it indeed exists in the isotropic tomographic models investigated here. This bias gets weaker with increasing depth of the anisotropic material, and compared to the isotropic velocity anomalies typically interpreted by tomographers, the bias from anisotropy is small.

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1. Introduction

[2] Inversion of traveltimes or waveforms of seismic waves for three-dimensional seismic velocity models is a common procedure in seismology, finding widespread application. The resulting tomographic models provide insights into the Earth's interior structure and constrain the temperature, composition, and associated dynamics of the mantle. Interpreting the seismic velocities requires a thorough understanding of the complex uncertainties and errors. One such error could come from ignoring azimuthal anisotropy. Although observations of split *SKS* phases, for example, provide evidence for anisotropy, many tomographic models have been obtained under the assumption that the seismic velocity structure of the Earth is isotropic. This anisotropy might thus have biased the isotropic velocity models. *Sobolev et al.* [1999] investigated this bias in isotropic P velocity models inferred from teleseismic P wave traveltimes. They found that anisotropy-induced artifacts appear for a variety of anisotropic structures. We investigate this bias in isotropic S velocity models inferred from regional S and surface waveforms.

[3] *Sobolev et al.* [1999] illustrate that quantifying anisotropic bias is not straightforward. This is because the location and intensity of the anisotropic



effects depend on how azimuthally uniform the wave path coverage is in an anisotropic region. If seismic waves traverse an anisotropic body predominately parallel to the fast velocity axis, seismic velocity at the location will be overestimated in the tomographic model. Conversely, if the waves travel predominantly in the slow-velocity direction, velocities there will be underestimated. Only if an anisotropic region is perfectly uniformly sampled by the seismic waves used for the tomographic inversion can its mean velocity be well estimated. In reality, study regions are imperfectly sampled. Thus, the strength of anisotropic bias at a given location in a tomographic model is expected to not only depend on the strength of the observed anisotropy at that location but also on the wave path coverage. To estimate the bias, it is necessary to separate the data into an isotropic and an anisotropic component. In this study we investigate and quantify the potential anisotropic bias in isotropic S velocity models due to ignoring azimuthal anisotropy [e.g., Marone and Romanowicz, 2007; Montagner and Griot, 2000]. To this end we convert the delay times and orientations of observed shear wave splitting into equivalent S velocity perturbations for each wave path. We invert these perturbations to image potential bias and subtract them from observed data before the inversion for an anisotropy corrected isotropic model. We do so for three 3-D tomographic S velocity models: NA04 for North America [van der Lee and Frederiksen, 2005], EAV03 for the Mediterranean region [Marone et al., 2004], and SA99 for South America [van der Lee et al., 2001] to quantify the possible bias caused by ignoring azimuthal anisotropy.

2. Method

2.1. Inversion

[4] The Partitioned Waveform Inversion (PWI) [Nolet, 1990; van der Lee and Nolet, 1997] is a two-step method, which was used to obtain the three tomographic models under investigation. First, a nonlinear waveform inversion is performed for each wave path separately, to determine uncorrelated path averaged linear constraints on the velocity structure between respective source and receiver pairs. In a second step, these constraints are combined in an inversion to infer a 3-D S velocity model. This system of linear equations inverted in this second step is where **m** is the unknown model vector, representing either SA99, EAV03, or NA04, depending on which data was used in the first step of the PWI to determine the data vector \mathbf{d}_{obs} . Ideally the waveforms fitted in the first step constrain purely isotropic signal, such that the resulting linear equations are free from anisotropic effects. In reality this is not likely the case, and equation (1) may be written as

$$\mathbf{G}(\mathbf{m}_{iso} + \mathbf{m}_{bias}) = \mathbf{d}_{iso} + \mathbf{d}_{aniso}.$$
 (2)

where

$$\mathbf{m}_{iso} + \mathbf{m}_{bias} = \mathbf{m} \tag{3}$$

and

$$\mathbf{d}_{iso} + \mathbf{d}_{aniso} = \mathbf{d}_{obs}.$$
 (4)

We estimate a likely anisotropic contribution to the data vector \mathbf{d}_{aniso} from delay times and orientations of the fast velocity axes of split shear waves. We then consider the following two equations for the anisotropic bias \mathbf{m}_{bias} and an anisotropy-corrected isotropic model \mathbf{m}_{iso} :

$$\mathbf{G}\mathbf{m}_{bias} = \mathbf{d}_{aniso} \tag{5}$$

$$\mathbf{G}\mathbf{m}_{iso} = \mathbf{d}_{obs} - \mathbf{d}_{aniso}.$$
 (6)

2.2. Anisotropy Data Vector

[5] The inversion described above requires the anisotropy data vector \mathbf{d}_{aniso} . It is derived from the observed azimuthal anisotropy in three steps: (1) interpolation of observed anisotropy over a regular grid (see auxiliary material Text S1¹), (2) conversion to equivalent orientation-dependent *S* velocity perturbations that produce the observed splitting delay times and orientations, and (3) averaging of the orientation-dependent equivalent *S* velocity perturbations along each wave path. The result of these three steps is a path averaged 1-D velocity model containing the anisotropic effects for each wave path. It is described by uncorrelated parameters, which, together with the constraints for the other wave paths, form \mathbf{d}_{aniso} .

[6] In constructing our anisotropy data vector we limit ourselves to the assumptions made in the studies that provide the input data (observed

¹Auxiliary materials are available in the HTML. doi:10.1029/2008GC001997.



anisotropy). Thus we consider only orthorhombic anisotropy (which is the configuration of, e.g., olivine, wadsleyite and ringwoodite [Mainprice et al., 2000]) with a horizontal symmetry axis as assumed in many SKS studies [Savage, 1999]. Similarly we do not consider variations in azimuth of anisotropy with depth as the observations typically constrain such variations insufficiently. The depth and the thickness of the anisotropic region cannot be well constrained from analysis of split shear waves. It is possible that all the inferred anisotropy is in the lithosphere [e.g., Silver, 1996], in the sublithospheric mantle [e.g., Vinnik et al., 1992], or distributed over lithosphere and asthenosphere [e.g., Fouch et al., 2000; Marone et al., 2007]. We explore scenarios for the distribution of anisotropy using natural boundaries of the Earth. Because independent constraints on crustal thickness were used in the isotropic tomographic inversions, crustal structure is effectively separated from mantle structure. Therefore we ignore crustal anisotropy and pick the Moho (CRUST2.0 [Bassin et al., 2000]) as the upper boundary for possible anisotropy occurrence. However, since SKS splitting is affected by crustal anisotropy our projection of all SKS splitting into the mantle tends to overestimate anisotropy in the mantle slightly. The lithosphere-asthenosphere boundary defines an intermediate depth boundary, for which we use the thermal model TC1 [Artemieva, 2006]. TC1 requires xenolith data to determine the geotherm in tectonically active regions; since these are not always available, for example in the Mediterranean, it is heterogeneously resolved. Nonetheless it is a better approximation than a constant depth boundary. The diffusion-dislocation creep transition marks the deepest level to which anisotropy might easily reach [e.g., Karato and Wu, 1993]. We position this transition at 200 km below the bottom of the lithosphere, but nowhere deeper than 300 km [Podolefsky et al., 2004; Mainprice et al., 2005]. We define two end-member cases with all the anisotropy either in the mantle lithosphere or entirely below the lithosphere. In addition we define an intermediate case, with the anisotropy being equally strong in both regions. In regions where the lithosphere is thin (e.g., the western margin of North America) its thickness is too small to physically accommodate all of the observed anisotropy. In order to reproduce the observed SKS delay times, the lithospheric velocity in the fast direction would have to be up to $\delta V = 900$ m/s greater than in the slow direction. If all olivine crystals were aligned, this difference would be at most 500 m/s [e.g., Abramson et al., 1997; Brugger, 1965]. Since olivine only accounts for about half of the composition of the lithosphere we limit the maximum possible velocity difference to 250 m/s and assume the remainder of the observed anisotropy originates below the lithosphere. However, this limit appears to have a negligible effect on our results and conclusions.

[7] The accumulated *SKS* delay time caused by the anisotropic velocity difference δV within one azimuthally anisotropic layer of thickness δz is

$$\delta t = \frac{\delta z}{V_{S_0} - \frac{1}{2}\delta V} - \frac{\delta z}{V_{S_0} + \frac{1}{2}\delta V}.$$
(7)

After reordering and dropping δV^2 terms (since in the case of weak anisotropy $V_{S_0}^2 \gg \delta V^2$) we obtain

$$\delta t = \frac{\delta z \delta V}{V_{S_0}^2 - \frac{1}{2} \delta V^2} \approx \frac{\delta z \delta V}{V_{S_0}^2},\tag{8}$$

which is the same expression as given by *Montagner and Griot* [2000]. The total SKS delay time for multiple layers with identical fast axis direction is

$$\delta t = \sum_{i=1}^{n} \frac{\delta z_i \delta V_i}{V_{S_{0_i}}^2}.$$
(9)

Our hypothesized anisotropy distributions use a maximum of two layers (n = 2) with different strengths (w_i) of the anisotropy. Substituting $\delta V_i = \delta \bar{V} w_i$ and assuming two layers we find an expression for the velocity difference between the fast and the slow axes:

$$\delta \bar{V} = \frac{\delta t V_{S_{0_1}}^2 V_{S_{0_2}}^2}{\delta z_1 w_1 V_{S_{0_2}}^2 + \delta z_2 w_2 V_{S_{0_1}}^2}.$$
 (10)

For each hypothesized anisotropy distribution we calculate δV at each point in the grid and calculate the azimuth dependant horizontal velocities as seen by surface waves according to *Montagner and Griot* [2000]. We then use our sensitivity matrix **G** to compute the effect of these velocities on our data for each wave path. The effects of anisotropy combined for all wave paths (in our data set) are grouped in a data vector **d**_{aniso}. This procedure takes into account that Rayleigh wave sensitivity diminishes with depth and that higher frequency Rayleigh waves. We do not account for the evidence that SKS splitting may be more strongly affected by upper layer anisotropy then by



Figure 1. (a) Resulting anisotropic bias, \mathbf{m}_{bias} , for the case where the anisotropic material is assumed to be entirely in the lithosphere. In regions with positive δV_{bias} velocities in NA04 could be overestimated; in regions with negative δV_{bias} they could be underestimated. (b) Same as Figure 1a, but for the anisotropy evenly distributed in the lithosphere and asthenosphere. (c) Same as Figure 1a, but for all the anisotropy in the asthenosphere. (d) The much stronger isotropic velocity perturbations (compared to \mathbf{m}_{bias}) in NA04 [van der Lee et al., 2001].

lower layer anisotropy [e.g., *Rümker and Silver*, 1998; *Saltzer et al.*, 2000], in part because we do not have details on the variation of frequency content of the individual SKS measurements from the more than 80 studies we used from the anisotropy database in step 1.

[8] Using the same parameterization and regularization as in the original models, this data vector may now be inverted for \mathbf{m}_{bias} , or can be used to remove the contribution from anisotropy from our observed data \mathbf{d}_{obs} . The corrected data is then inverted for an anisotropy-corrected model \mathbf{m}_{iso} .

3. Results

[9] Figure 1 shows the anisotropic bias \mathbf{m}_{bias} for NA04 for all three anisotropy distributions and the isotropic velocity perturbations in NA04. Figures 2 and 3 show \mathbf{m}_{bias} for EAV03 and SA99 only for the shallowest anisotropic case because this causes the strongest bias and thus represents a worst-case endmember. For these end-member scenarios, auxiliary material Figures S1–S4 compare the original models with the respective anisotropy corrected models \mathbf{m}_{iso} .

3.1. NA04

[10] Figure 1 shows that anisotropic bias could be present in model NA04 because the inversion of the equivalent S velocity perturbations yields nonzero values in \mathbf{m}_{bias} (henceforth we refer to these pseudovelocity perturbations as δV_{bias}). Positive δV_{bias} indicates regions where the velocity is possibly overestimated in NA04 because of ignoring anisotropy, while negative values indicate regions where the velocity may be underestimated. Anisotropic bias is observed for all hypothesized anisotropy distributions, but they differ in the magnitude of the bias introduced. The strongest bias occurs in the 90 km slice of Figure 1a, which was obtained assuming all the anisotropy is concentrated in the lithosphere. When the anisotropy is evenly distributed over the lithosphere and asthenosphere (Figure 1b), or if it is completely in the asthenosphere (Figure 1c), the associated anisotropic bias is much less. The distribution of positive and negative regions is the same for all depths throughout the different hypothesized anisotropy distributions. The largest positive (blue) bias is located beneath Northern California/southern Oregon. Here, δV_{bias} reaches values of up to





Figure 2. Same as Figure 1 but for the Mediterranean. We only show \mathbf{m}_{bias} for the case where all the anisotropy is in the lithosphere.

140 m/s. The largest negative bias (red) is beneath the western Gulf of Mexico and southern Mexico. The δV_{bias} of this anomaly is around -120 m/s and connects to the less negative δV_{bias} that pervades the Atlantic Ocean. Several patches within the United States show weaker bias ($|V_{bias}| \le 100 \text{ m/s}$). Positive bias is imaged beneath Wyoming and Colorado, as well as the east coast of the US. Canada also contains both positive and negative δV_{bias} , but the amplitudes are relatively small compared to the ones observed in the US because the sparse SKS constraints on anisotropy there yield smooth anisotropic heterogeneity. Compared to the isotropic velocity perturbations δV_S in NA04 (Figure 1d), even the strongest anisotropic bias visible for the worst-case scenario (all the anisotropy in the lithosphere) the anisotropic bias is small.

[11] Because the effect of anisotropy is small, the validity of NA04 is not altered by our results. However, since several tectonic regions discussed by *van der Lee and Frederiksen* [2005] correlate with the distribution of anisotropy (i.e., similar geographic extent of isotropic and anisotropic-bias features) it is worth mentioning the effect of anisotropy in these regions. For example, they image higher S velocities in the Atlantic upper mantle compared to the Pacific upper mantle, high

upper mantle *S* velocities beneath the Aleutian arc, and the absence of high velocities in the upper mantle beneath Wyoming. In all three cases, taking the possible anisotropic bias into account would increase the effects leading to these observations. Conversely, accounting for anisotropy might have a weakening effect on the North American craton being disconnected from Greenland's lithosphere.

3.2. EAV03

[12] Shallow anisotropy in the Mediterranean region produces δV_{bias} at comparable strength as for North American model NA04 (Figure 2, left). Positive bias ($\delta V_{bias} \approx 100$ m/s) is found in the westernmost Mediterranean region, the northern Carpathians and in eastern Turkey. Regions with a negative δV_{bias} are beneath the British islands, most of the eastern Mediterranean Sea (from the Ionian Sea eastward) and most of Northern Africa. For the part of the Atlantic included in EAV03, δV_{bias} is mostly positive. This is likely due to the direction of the wave paths traversing this region and the fast *S* velocity direction both being predominantly east–west.

[13] Features discussed by *Marone et al.* [2004] that are stronger after anisotropic correction are low velocities beneath northern Algeria at 100–200 km depth and high velocities beneath 300 km,





Figure 3. Same as Figure 2 but for South America.

high velocities beneath the bay of Biscay, Italy, Adriatic Sea, Peloponnese and southern Greece, and low velocities beneath the Pannonian and Moesian basins. Features that become weaker are high velocities in the western Mediterranean and between 25°W and the European coast, low velocities at the Mid-Atlantic Ridge, Azores, and beneath the western part of the Iberian peninsula. However, as with NA04, we note that the bias is much weaker than the lateral variations in isotropic velocities (Figure 2, right).

3.3. SA99

[14] SA99 differs from NA04 and EAV03 in that δV_{bias} is weaker and reaches peak values of ±80 m/s (Figure 3, left). This weakness is most likely a result of the sparse constraints on anisotropy for this continent. We identify two regions with high δV_{bias} : the northwestern margin of south America and an east–west corridor in central South America. These two regions border a region of negative bias

in the eastern Amazon Basin. The observed variations in δV_{bias} are negligible compared to the isotropic velocity variations and maps of SA99 (Figure 3, right). We therefore conclude that SA99 is not significantly biased by known anisotropy.

4. Summary and Discussion

[15] For the three tomographic models investigated in this study anisotropic bias does not seem to introduce major artifacts. Anisotropic bias appears to be strongest if the observed anisotropy originates entirely in the lithosphere, in particular when the lithosphere is thin (e.g., western North America). If the anisotropy originates below the lithosphere, it would yield smaller anisotropic bias. Correcting the isotropic tomographic models for the estimated anisotropic bias yields small corrections at most. This is encouraging in that ignoring azimuthal anisotropy in Rayleigh and S wave tomographic inversions with decent wave path coverage does

not appear to introduce significant artifacts. Marone et al. [2007] reach similar conclusions for North America. In addition, they find that radial anisotropy predominantly originates below the lithosphere. If azimuthal and radial anisotropy share this relatively deep origin, the actual anisotropic bias in isotropic model NA04 would be less than that of the hypothesized worst-case end-member in this study. In eastern North America anisotropy seems to be distributed over the lithosphere and asthenosphere [Fouch et al., 2000; Marone et al., 2007], likewise yielding less bias than assumed in Figure 1a. Therefore we conclude that isotropic model NA04 is not significantly biased by ignored azimuthal anisotropy. Mediterranean model EAV03 and South-American model SA99 appear similarly robust with respect to ignoring azimuthal anisotropy.

Acknowledgments

Geochemistry

Geophysics Geosystems

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