

Dynamics of the upper mantle in light of seismic anisotropy

Thorsten W. Becker^{1,2} and Sergei Lebedev³

¹Institute for Geophysics, Jackson School of Geosciences, The University of Texas at Austin, USA

²Department of Geological Sciences, Jackson School of Geosciences, The University of Texas at Austin, USA

³Dublin Institute for Advanced Studies, Dublin, Ireland

Corresponding author: T. W. Becker, twb@ig.utexas.edu

Abstract

Seismic anisotropy records continental dynamics in the crust and convective deformation in the mantle. Deciphering this archive holds huge promise for our understanding of the thermo-chemical evolution of our planet, but doing so is complicated by incomplete imaging and non-unique interpretations. Here, we focus on the upper mantle and review seismological and laboratory constraints as well as geodynamic models of anisotropy within a dynamic framework. Mantle circulation models are able to explain the character and pattern of azimuthal anisotropy within and below oceanic plates at the largest scales. Using inferences based on such models provides key constraints on convection, including plate-mantle force transmission, the viscosity of the asthenosphere, and the net rotation of the lithosphere. Regionally, anisotropy can help further resolve smaller-scale convection, e.g. due to slabs and plumes in active tectonic settings. However, the story is more complex particularly for continental lithosphere, and many systematic relationships remain to be established more firmly. More integrated approaches based on new laboratory experiments, consideration of a wide range of geological and geophysical constraints, as well as hypothesis-driven seismological inversions are required to advance to the next level.

1 Introduction

Anisotropy of upper mantle rocks records the history of mantle convection and can be inferred remotely from seismology. Seismic anisotropy refers to the orientational dependence of propagation velocities for waves traveling at different azimuths, or a difference in velocities for waves that are polarized in the horizontal or vertical plane such as Love and Rayleigh waves, respectively. “Anisotropy” without any qualifier shall here refer to the seismic kind caused by an anisotropic elastic stiffness tensor unless noted otherwise. Anisotropy is a common property of mineral assemblages and appears throughout the Earth, including in its upper mantle. There, anisotropy can arise due to the shear of rocks in mantle flow. As such, it provides a unique link between seismological observations and the evolution of our planet. However, given the need to resolve more parameters for an anisotropic than for an isotropic solid, seismological models for anisotropy are more uncertain, and the interpretation and link to flow necessarily non-unique.

Our personal views of seismic anisotropy have oscillated from a near-useless can of worms to the most useful constraint on convection ever, and we strive to present a more balanced view here. Anisotropy matters for all layers of the Earth, and there exist a number of excellent reviews covering the rock record, seismological observations, and laboratory constraints [e.g. *Nicolas and Christensen*, 1987; *Silver*, 1996; *Savage*, 1999; *Mainprice*, 2007; *Skemer and Hansen*, 2016; *Romanowicz and Wenk*, 2017], as well as comprehensive treatments in textbooks [e.g. *Anderson*, 1989]. Also, most of what was said in the overview of *Long and Becker* [2010] remains relevant. However, here we shall focus our discussion on the upper mantle within and underneath oceanic plates, the seemingly best understood part of mantle convection. We will highlight some of the insights afforded by seismic anisotropy within a convective context, and discuss selected open questions and how to possibly answer them.

2 Observations of seismic anisotropy

A range of seismic observations show the presence of anisotropy in the Earth. In tomographic imaging of its three-dimensional distribution, anisotropy must normally be resolved simultaneously with the isotropic seismic velocity heterogeneity, typically greater in amplitude. Substantial non-uniqueness of the solutions for anisotropy can arise [e.g. *Laske and Masters*, 1998], and other sources of uncertainties include the treatment of the crust [e.g. *Ferreira et al.*, 2010] and earthquake locations [*Ma and Masters*, 2015]. In fact, the very existence of intrinsic anisotropy (e.g. due to lattice preferred orientation (LPO)

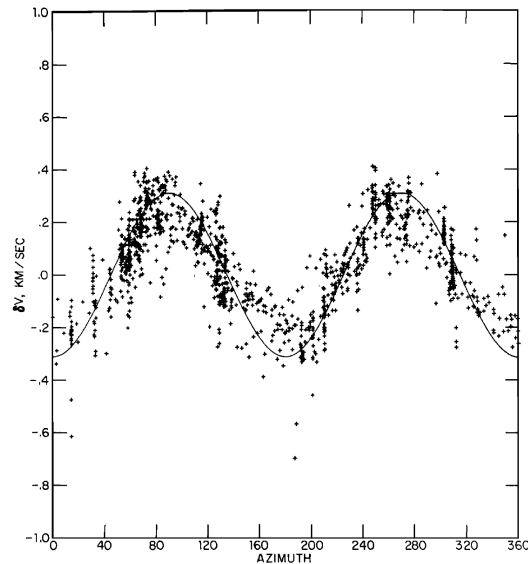


Figure 1. Velocity deviation for Pn from the mean ($v_P = 8.159$ km/s) in the central Pacific N and NW of Hawaii as a function of propagation azimuth along with a 2φ fit (eq. 1). Modified from *Morris et al.* [1969].

of anisotropic mantle peridotite minerals) as opposed to apparent anisotropy [e.g. caused by layering of isotropic material of different wave speeds; e.g. *Backus*, 1962] has been debated [*Fichtner et al.*, 2013; *Wang et al.*, 2013].

At least regionally, the occurrence of anisotropy is, of course, not really in doubt since different lines of seismic evidence for it are corroborated by observations from mantle rocks [e.g. *Ben Ismail and Mainprice*, 1998; *Mainprice*, 2007]. However, accurate determination of anisotropy is clearly not straightforward. Models based on data of different types, or even of the same type, are often difficult to reconcile and only agree on large spatial scales. Improvements in data sampling and anisotropy analysis methods are therefore subjects of active research, aimed at yielding more accurate and detailed information on the dynamics of the lithosphere and underlying mantle.

In order to ground the dynamics discussion, we first address the scales of resolution and distribution of seismic anisotropy coverage that are currently available to guide global mantle circulation assessment.

2.1 Pn anisotropy

Historically, the detection of P wave anisotropy just below the Moho from refraction experiments in the Pacific Ocean was important in terms of establishing the existence of seismic anisotropy in the upper mantle and linking it to plate tectonics [e.g. *Hess*, 1964; *Morris et al.*, 1969]. It can be shown that the azimuth, φ , dependence of P wave speed anomalies, δv , for small seismic anisotropy at location \mathbf{x} can be approximated by

$$\delta v(\varphi, \mathbf{x}) \approx A_0(\mathbf{x}) + A_1(\mathbf{x}) \cos(2\varphi) + A_2(\mathbf{x}) \sin(2\varphi) + A_3(\mathbf{x}) \cos(4\varphi) + A_4(\mathbf{x}) \sin(4\varphi) \quad (1)$$

[*Backus*, 1965]. The simplest form of azimuthal anisotropy is due to the 2φ terms alone, and the corresponding 180° periodic pattern seen in *Morris et al.*'s [1969] results, for example (Figure 1). Based on such patterns, *Hess* [1964] concluded that the oceanic lithosphere and uppermost mantle must have undergone convective flow and made the connection to seafloor spreading [*Vine and Matthews*, 1963].

Anisotropy beneath continents was also detected, using both refraction and quarry-blast data [e.g. *Bamford, 1977*]. More recently, *P_n* and *S_n* waves propagating from earthquakes have been used for mapping azimuthal anisotropy in the uppermost mantle, just beneath the Moho [e.g. *Smith and Ekström, 1999; Buehler and Shearer, 2010*], where they form a connection between shallow, crustal anisotropy and the deeper mantle observations such as from *SKS* splitting which we discuss next.

2.2 Shear wave splitting

In the presence of azimuthal anisotropy, a shear wave pulse traveling into an anisotropic layer will be separated into two orthogonal pulses, one propagating within the medium's fast polarization plane (containing its "fast axis", or fast-propagation azimuth), and the other within the orthogonally-oriented, slow propagation plane. At a seismic station, those split pulses will arrive separated by a delay time, δt , that is proportional to the integral of anisotropy strength and path length, assuming a uniform anisotropy orientation within the anisotropic layer [e.g. *Silver and Chan, 1988; Vinnik et al., 1989*]. Such "splitting" is akin to optical birefringence and observed for local shear wave arrivals in the shallow crust ($\delta t \lesssim 0.2$ s) where it mainly reflects anisotropy due to aligned cracks, whose opening is controlled by tectonic stresses [*Crampin and Chastin, 2003*]. For teleseismic shear waves, $\delta t \sim 1.2$ s, on average, and the splitting measurements can be related to whole-crustal and mantle anisotropy [*Vinnik et al., 1992; Silver, 1996*]. *SKS* splitting due to anisotropic fabric within the crust is typically $\lesssim 0.3$ s, much smaller than that accumulated in the mantle. Areas with anomalously thick crust, for example Tibet, are the exception where crustal delay times have been estimated to be up to ~ 0.8 s [e.g. *Agius and Lebedev, 2017*].

The popular shear-wave splitting method yields a direct indication of anisotropy in the Earth [e.g. *Savage, 1999*]. Outer-core-traversing waves such as *SKS* and *SKKS* are often used for the splitting measurements because they can yield information on receiver side anisotropy; source effects are excluded because of the *P*-to-*S* conversion upon exiting the core. The advantages of the method are its ease of use and its high lateral resolution. Figure 2 shows the current distribution of teleseismic shear wave splitting measurements with fairly dense sampling in most of the actively deforming continental regions.

The main disadvantage of *SKS* splitting is its poor vertical resolution; anisotropy may arise anywhere along the path. In the presence of one dominant anisotropic layer (say, the asthenosphere) with azimuthal anisotropy, the splitting parameters (delay times and fast azimuth) will characterize this layer directly. However, if multiple layers with different fast axes or more complex types of anisotropy are present, the net splitting will depend non-linearly on back-azimuth and the depth-variable anisotropy [e.g. *Silver and Savage, 1994; Rümpler and Silver, 1998; Saltzer et al., 2000*]. Resolving some of the depth-dependence is possible with dense spatial coverage but requires long station deployment times and good back-azimuthal sampling [e.g. *Chevrot et al., 2004; Long et al., 2008; Abt and Fischer, 2008; Montellier and Chevrot, 2011*].

When considering the uncertainty in the mantle depths where teleseismic splitting arises, we can focus on high stress/low temperature boundary layers where dislocation creep might dominate [*Karato, 1992; Gaherty and Jordan, 1995; McNamara et al., 2001*]. For *SKS* splitting this means uncertainty on whether the delay times are caused by anisotropy in the lithosphere, asthenosphere, the transition zone between the upper and lower mantle [e.g. *Fouch and Fischer, 1996; Wookey and Kendall, 2004*], and/or the core-mantle boundary/D" region (reviewed elsewhere in this volume).

The integrated anisotropy of the lithosphere alone is typically not enough to fully explain *SKS* splitting delay times [e.g. *Vinnik et al., 1992; Silver, 1996*]. Comparisons between local and teleseismic splitting from subduction zones are usually consistent with an origin of most *SKS* splitting observations within the top ~ 400 km of the mantle [e.g. *Fischer and Wiens, 1996; Long and van der Hilst, 2006*]. Together with surface wave mod-

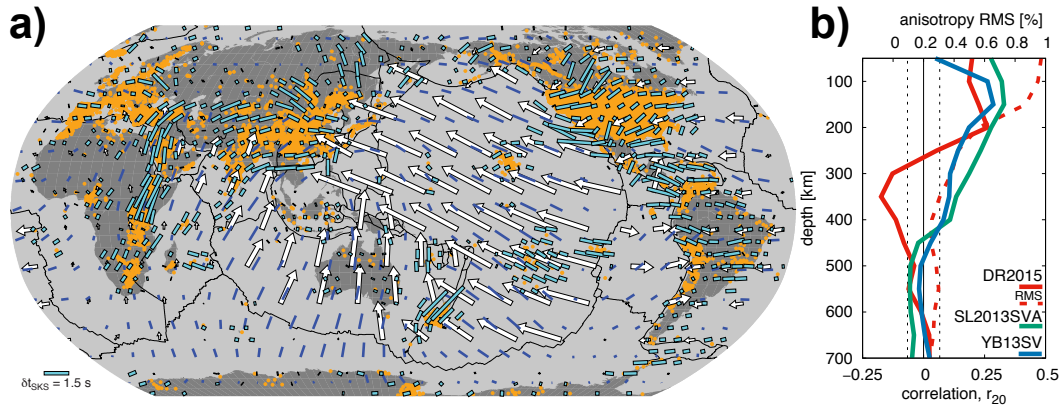


Figure 2. Azimuthal anisotropy of the upper mantle. **a)** Non-zero *SKS* splitting observations (orange dots) fit using spherical harmonics up to degree, $\ell = 20$ [cyan sticks, processed as in *Becker et al., 2012*, and updated as of 01/2019], and compared to the global azimuthal anisotropy model SL2013SVA at 200 km depth [blue sticks; *Schaeffer et al., 2016*], and MORVEL [*DeMets et al., 2010*] plate motions in a spreading-aligned reference frame [white vectors; *Becker et al., 2015*]. **b)** Correlation up to $\ell = 20$, r_{20} (solid lines), between *SKS* splitting (a) and three seismological models: DR2015 [*Debayle and Ricard, 2013*, RMS anisotropy also shown with dashed line], SL2013SVA [*Schaeffer et al., 2016*], and YB13SV [*Yuan and Beghein, 2013*]. Dashed vertical lines are 95% significance levels for r_{20} [cf. *Becker et al., 2007a, 2012*; *Yuan and Beghein, 2013*].

els of anisotropy (Figure 2b) as well as mineral physics and dynamics considerations discussed below, this suggests a dominant asthenospheric cause of *SKS* splitting.

2.3 Surface waves

There are a range of other approaches used for mapping anisotropy, including study of *P*-wave polarization [*Schulte-Pelkum et al., 2001*], body-wave imaging [e.g. *Plomerová et al., 1996*; *Ishise and Oda, 2005*; *Wang and Zhao, 2008*], receiver-function anisotropy analysis [e.g. *Kosarev et al., 1984*; *Farra and Vinnik, 2002*; *Schulte-Pelkum and Mahan, 2014*], and normal-mode measurements [e.g. *Anderson and Dziewoński, 1982*; *Beghein et al., 2008*]. However, for global-scale imaging of the upper mantle, surface wave analysis holds most promise for making the link to depth-dependent convection scenarios.

Just as the response of the Earth to a seismic event can be expressed as a superposition of normal modes (standing waves), it can be decomposed into a sum of surface waves [traveling waves; *Dahlen and Tromp, 1998*]. The depth sensitivity of surface waves depends on their period; the longer the period, the deeper they sample. Global maps of surface-wave phase velocities at periods from ~ 35 to 150 s, sampling the mantle lithosphere and asthenosphere, have been available for over two decades [e.g. *Ekström et al., 1997*; *Trampert and Woodhouse, 2003*]. More recently, global models have been constructed with surface waves in broadening period ranges, up to ~ 25 -250 s [*Ekström, 2011*] and even up to 10-400 s [*Schaeffer and Lebedev, 2013*], although at the shortest of the periods the event-station measurements can no longer cover the entire globe.

Using the ambient noise wave field, speeds of the surface waves excited by ocean waves are routinely measured in a 1-35 s period range, i.e. sensing from the uppermost crust to the uppermost mantle [*Shapiro et al., 2005*; *Bensen et al., 2007*; *Ekström et al., 2009*]. Anthropogenic noise yields measurements at frequencies of a few Hz to a few tens of Hz, sampling within the shallowest, sedimentary layers [*Mordret et al., 2013*]. Cross-

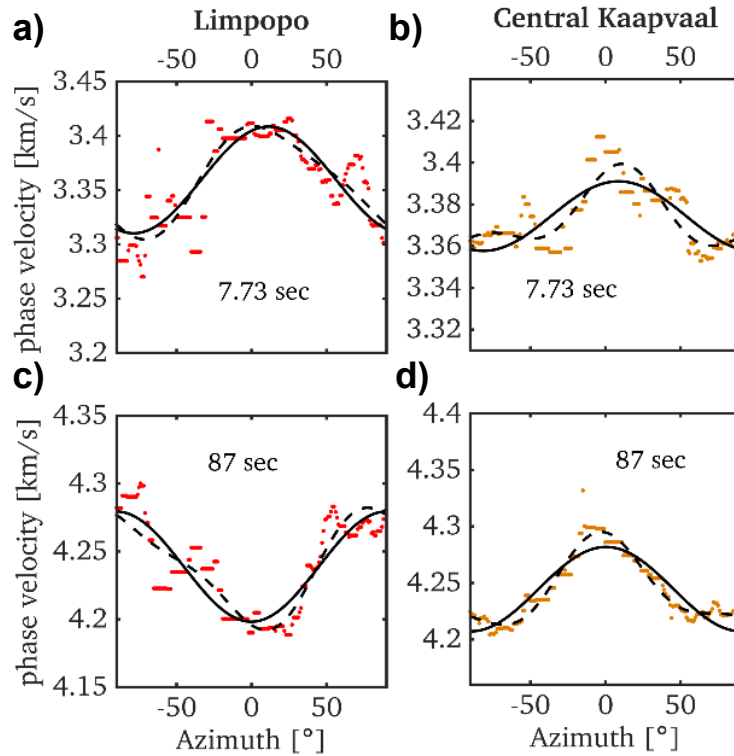


Figure 3. Azimuth, φ , dependent anisotropy of Rayleigh-wave phase velocities in different regions (a and c vs. b and d) in southern Africa. Rayleigh waves at the 7.73 s period (a and b) sample primarily the upper and middle crust, and at 87 s (c and d), the lower part of the cratonic lithosphere, respectively. Dots show the phase-velocity measurements, binned and smoothed with a 30° sliding window. Solid black lines: best-fitting models with isotropic and 2φ terms (see eq. 1). Dashed black lines: best-fitting models with isotropic, 2φ , and 4φ terms. Modified from *Ravenna* [2018], using measurements from *Adam and Lebedev* [2012].

correlations of seismograms from teleseismic earthquakes yield phase-velocity measurements down to periods as short as 5-10 s, sampling the upper and middle crust [*Meier et al.*, 2004; *Adam and Lebedev*, 2012] (Figure 3) and up to periods over 300 s [e.g. *Lebedev et al.*, 2006], sampling the deep upper mantle and transition zone.

Rayleigh waves are mainly sensitive to vertically polarized shear wave speed, v_{SV} , with smaller, although non-negligible, sensitivity to horizontally polarized shear wave velocity, v_{SH} , and v_P [e.g. *Montagner and Nataf*, 1986; *Romanowicz and Snieder*, 1988; *Dahlen and Tromp*, 1998]. The azimuthal expansion of eq. (1) holds for surface waves as well [*Smith and Dahlen*, 1973], and in the olivine dominated upper mantle, the 2φ terms of eq. (1) are expected to be the main signature of azimuthal anisotropy for Rayleigh waves [*Montagner and Nataf*, 1986; *Montagner and Anderson*, 1989, cf. Figure 3]. At the same periods, Love waves are mainly sensitive to v_{SH} at shallower depths, and the 4φ terms of azimuthal anisotropy, depending on assumptions about petrology [*Montagner and Nataf*, 1986].

Radial anisotropy (the difference between v_{SV} and v_{SH}) was documented based on the finding that Love and Rayleigh waves could not be fit simultaneously by the same Earth model [*Anderson*, 1961; *Aki and Kaminuma*, 1963; *McEvelly*, 1964]. Azimuthal anisotropy of surface waves was also established early [*Forsyth*, 1975], and *Montagner and*

Tanimoto [1991] presented an integrated model of upper mantle anisotropy capturing both radial and azimuthal contributions.

A full description of seismic anisotropy is achieved by an elastic stiffness tensor with 21 independent components instead of the isotropic two [e.g. Anderson, 1989], but often hexagonal symmetry (or “transverse isotropy”) is assumed. In this case, five parameters fully specify the tensor, for example the vertically and horizontally polarized S and P wave speeds, v_{SV} , v_{SH} , v_{PV} , and v_{PH} , respectively, and a parameter η which determines how waves polarized between the horizontal and vertical plane transition from v_{SH} to v_{SV} [e.g. Dziewoński and Anderson, 1981; Kawakatsu, 2016]. In the case of radial anisotropy imaging, the hexagonal symmetry axis is assumed vertical, and $\xi = (v_{SH}/v_{SV})^2$ is commonly used as a measure of anisotropy strength. For the case of azimuthal anisotropy, the hexagonal symmetry axis is in the horizontal plane and its azimuth determines the 2φ terms of eq. (1), e.g. for the Rayleigh wave, v_{SV} , anisotropy or the fast axes of SKS splitting.

The construction of large waveform datasets over the last two decades has enabled increasingly detailed surface-wave tomography of upper-mantle anisotropy on global scales. A number of 3-D radial [e.g. Nataf et al., 1984; Ekström and Dziewoński, 1998; Panning and Romanowicz, 2006; Kustowski et al., 2008; French and Romanowicz, 2014; Auer et al., 2014; Moulik and Ekström, 2014; Chang et al., 2015] and azimuthal [e.g. Tanimoto and Anderson, 1984; Montagner, 2002; Debayle and Ricard, 2013; Yuan and Beghein, 2013; Schaeffer et al., 2016] (Figure 2) anisotropy models have been presented.

Many features of anisotropic structure are now consistently mapped for the upper mantle on continent scales. The mutual agreement of different anisotropy models, however, remains well below that of models of isotropic heterogeneity [Becker et al., 2007a; Auer et al., 2014; Chang et al., 2015; Schaeffer et al., 2016]. Given the typical period range for fundamental mode surface wave measurements, both radial and azimuthal anisotropy are best constrained in the uppermost ~ 350 km of the mantle, even though comprehensive waveform analysis [e.g. Lebedev et al., 2005; Priestley et al., 2006; Panning and Romanowicz, 2006] or the explicit use of overtones [e.g. Trampert and van Heijst, 2002; Beghein and Trampert, 2004] extends the depth range to the bottom of the transition zone (~ 700 km) and beyond, at least theoretically.

Dense arrays of seismic stations enable higher lateral resolution surface wave anisotropy imaging regionally [e.g. Shapiro et al., 2004; Deschamps et al., 2008a; Lin et al., 2011; Takeo et al., 2018; Lin et al., 2016]. On those scales, it is also easier to explore uncertainties, and probabilistic 1-D profiles obtained with Monte Carlo inversion schemes can be used, for example, to explore the trade-off between the radial and azimuthal anisotropy layer imaging [e.g. Beghein and Trampert, 2004; Agius and Lebedev, 2014; Bodin et al., 2016; Ravenna et al., 2018].

Uncertainties aside, array measurements can present unambiguous evidence of anisotropy in the crust and upper mantle beneath the array footprint. Figure 3 shows an example for a continental plate site. The measurements of phase velocities for different period Rayleigh waves clearly indicate seismic azimuthal anisotropy of the 2φ kind (cf. Figure 1), and a change in the fast propagation azimuth from the shallow to the deeper layers.

3 Interpretation of seismic anisotropy

3.1 Origin of upper mantle anisotropy

Shear due to convective flow is expected to lead to the formation of lattice (or, more appropriately “crystallographic”) preferred orientation anisotropy in the olivine-dominated upper mantle, meaning that anisotropy should be a record of mantle flow [e.g. McKenzie, 1979; Tanimoto and Anderson, 1984; Ribe, 1989]. The foundations for this common

assumption include the observation that natural xenolith and exhumed mantle massif samples show such alignment [e.g. *Nicolas and Christensen, 1987; Ben Ismail and Mainprice, 1998*], and that laboratory experiments indicate a link between the orientation and amount of shear induced deformation and the resulting LPO [e.g. *Karato et al., 2008; Skemer and Hansen, 2016*]. For olivine single crystals $\sim 75\%$ of the total elastic anisotropy is hexagonal, while most of the remainder is of orthorhombic symmetry [e.g. *Browaeyns and Chevrot, 2004*]. For assemblages, the hexagonal contribution ranges from $\sim 80\%$ for peridotites from spreading centers to $\sim 55\%$ in xenoliths from kimberlites in the compilation of *Ben Ismail and Mainprice [1998]*. This apparent predominance of hexagonal anisotropy for mantle assemblages motivates the approximations usually made in seismology.

LPO development is usually assumed to require not just solid state convection but deformation within the dislocation creep regime. For typical olivine grain sizes of order mm, this implies that LPO formation and hence seismic anisotropy will be enhanced in the mantle's boundary layers [e.g. *Karato, 1998; Podolefsky et al., 2004; Becker, 2006*]. Thus, shear within the asthenosphere underneath the lithospheric plates, say within the top ~ 400 km of the mantle, is expected to dominate the upper mantle signal of geologically recent anisotropy formation. The more slowly deforming lithosphere may record past episodes of deformation or creation in the case of continental and oceanic lithosphere, respectively [e.g. *Vinnik et al., 1992; Silver, 1996*].

There are possible other contribution to anisotropy besides LPO due to past and present mantle flow, such as preserved shape preferred fabrics or LPO within the crust [e.g. *Godfrey et al., 2000; Brownlee et al., 2017*], or the effects of partial melt [e.g. *Blackman et al., 1996; Holtzman and Kendall, 2010; Hansen et al., 2016a*]. An effectively anisotropic partial-melt layer at the base of the lithosphere can explain observed impedance contrasts, for example [e.g. *Kawakatsu et al., 2009*]. However, it is commonly held that regions of large partial melt fraction are of limited spatial extent away from spreading centers and continental rifts. This will be revisited below.

When deforming olivine aggregates in the laboratory, anisotropy strength due to LPO saturates at linear strains, γ , of $\lesssim 5 \dots 10$ [e.g. *Zhang and Karato, 1995; Bystricky et al., 2000; Hansen et al., 2014*]. Pre-existing textures likely require larger strain values for reorientation, in broad accordance with observations from the field [e.g. *Skemer and Hansen, 2016*]. For strain-rates that might be typical for the asthenosphere, say $\sim 5 \times 10^{-15} \text{ s}^{-1}$ (e.g. a plate moving at 5 cm/yr inducing shear over a ~ 300 km thick layer), $\gamma = 5$ is achieved in ~ 30 Myr. Using circulation computations and finite strain tracking, one arrives at similar numbers; times of advection in mantle flow are commonly between 10 and 30 Myr over path lengths between 500 km to 1500 km, respectively [*Becker et al., 2006a*]. In the highly deforming asthenosphere, these relatively short saturation or reworking times of order of 10s of Myr then determine the “memory” of seismic anisotropy, i.e. how much convective history and changes in plate motions are recorded. Within the cold and hence slowly deforming lithosphere, older episodes of deformation may be partially frozen-in for very long times, say $\gtrsim 300$ Myr in continents. This is longer than the characteristic lifetime of an oceanic plate, though it is most likely not a continuous record that is being preserved [e.g. *Silver, 1996; Boneh et al., 2017*].

In strongly and coherently deforming regions of the upper mantle, we therefore expect that the amplitude of anisotropy is mainly governed by the orientation of olivine LPO near saturation. Exceptions include spreading centers and subduction zones where a transition from simple to pure shear during vertical mass transport will lead to strong reworking of fabrics [e.g. *Blackman and Kendall, 2002; Kaminski and Ribe, 2002; Becker et al., 2006a*]. Such reworking is where different mineral physics approaches regrettably diverge in their predictions [e.g. *Castelnau et al., 2009*], and constraints from the lab and field indicate a mismatch with widely used LPO modeling approaches [*Skemer et al., 2012; Boneh et al., 2015*].

Irrespective of the details of the LPO formation mechanism, we note that anisotropy strength is not expected to scale with absolute plate or slab velocity, rather it is spatial variations in velocities (i.e. strain-rates) that control the rate of anisotropy saturation. Any relationship between plate speed and the signature of anisotropy is thus likely indirect, for example such that LPO formation under plate-motion induced shear is more efficient compared to other processes like small-scale convection for faster plates with higher strain-rates [van Hunen and Čadek, 2009; Husson *et al.*, 2015].

3.2 Anisotropy and plate motions

Given the link between LPO induced anisotropy and mantle flow, a first order constraint on convection can thus be provided by the existence of significant radial anisotropy in the upper mantle [e.g. Dziewoński and Anderson, 1981; Nataf *et al.*, 1986; Beghein *et al.*, 2006; Wang *et al.*, 2013]. Due to the alignment of the fast symmetry axis of an LPO aggregate in the vertical or horizontal direction, a simple mantle convection cell with an oceanic plate forming at its top limb should display $v_{SH} > v_{SV}$ within and below the plate's interiors (dominating the global average), and $v_{SV} > v_{SH}$ within the up- and downwelling limbs underneath spreading centers and subduction zones, respectively [e.g. Montagner and Guillot, 2000].

Relatively few studies have addressed the distribution of average radial anisotropy in light of mantle dynamics [e.g. Regan and Anderson, 1984; Montagner and Tanimoto, 1991; Chastel *et al.*, 1993; Montagner, 1994; Babuška *et al.*, 1998; Plomerová *et al.*, 2002; Gung *et al.*, 2003]. Both average and broad-scale patterns of radial anisotropy can be shown to be consistent with the predictions from mantle convection computations with dislocation/diffusion creep olivine rheologies at grain sizes of order mm [Becker *et al.*, 2008; Behn *et al.*, 2009]. Amplitudes of radial anisotropy appear under-predicted within the lithosphere by convective LPO models, particularly within continental regions [Becker *et al.*, 2008]. This hints at an additional contribution, e.g. due to frozen in anisotropy similar to what has been suggested for oceanic plates [e.g. Beghein *et al.*, 2014; Auer *et al.*, 2015].

We now proceed to discuss the large-scale origin of azimuthal seismic anisotropy (Figure 2) in light of oceanic plate boundary dynamics [cf. Montagner and Guillot, 2000]. Within the small-strain-rate lithosphere, we expect azimuthal anisotropy to record past deformation during creation of the plate. This deformation may be inferred from the spreading directions and rates that are recorded in the gradients of seafloor age [e.g. Conrad and Lithgow-Bertelloni, 2007]. We can then compare the fast axes with paleo-spreading orientations [e.g. Hess, 1964; Forsyth, 1975; Nishimura and Forsyth, 1989].

Figure 4a shows a typical result for such a comparison. Spreading orientations overall represent a good first-order model of azimuthal anisotropy in the lithosphere. They appear recorded more clearly in anisotropy in younger rather than in older seafloor, particularly in the Pacific plate [e.g. Smith *et al.*, 2004; Debayle and Ricard, 2013; Becker *et al.*, 2014], perhaps due to small-scale reheating at ages older than ~80 Ma [cf. Nagihara *et al.*, 1996; Ritzwoller *et al.*, 2004]. Seafloor that was generated during higher spreading rate activity shows smaller orientational misfits with lithospheric azimuthal anisotropy than regions that were generated by slower spreading [Becker *et al.*, 2014], possibly indicating variations in the degree of ductile to brittle deformation [Gaherty *et al.*, 2004], asymmetry or non-ridge-perpendicular orientation of slow spreading, or the relative importance of small-scale convection [e.g. van Hunen and Čadek, 2009].

Besides controlling factors such as spreading rate and seafloor age which may have general relevance for the creation of oceanic lithosphere, there are also geographic differences (Figure 4a); the Atlantic displays larger misfits than the Pacific, for example. This might be an overall reflection of tectonics (Atlantic spreading rates are slower than Pacific ones). However, the resolution of surface wave anisotropy imaging is also spatially

variable [e.g. *Laske and Masters, 1998; Becker et al., 2003*] and in particular earthquake source location errors are mapped into larger variations in fast azimuths in the Atlantic than the Pacific domain [*Ma and Masters, 2015*].

If we seek an explanation for deeper, asthenospheric, layers, we can consider the orientation of azimuthal anisotropy compared to plate motions. The underlying assumption for such comparisons is that the direction of surface velocities in some absolute reference frame, e.g. as based on hotspots [e.g. *Minster and Jordan, 1978*] are indicative of the orientation of shear due to motion of the lithosphere with respect to a relatively stationary deep mantle. This is called an absolute plate motion (APM) model.

Even in the absence of convective contributions due to density anomalies, plate-induced mantle flow can lead to regionally significant deviations from the shear deformation that is indicated by the APM model. This is true in terms of the velocity magnitude, i.e. if the plate is leading the mantle or vice versa in simple shear (Couette) type flow (with possible effects on anisotropy dip angle), and it is also important in that the orientation of mantle flow may be very different from that of plate motion [*Hager and O'Connell, 1981*]. The sense of asthenospheric shear may thus be at large angles to APM orientations. Moreover, the degree to which asthenospheric flow is of the plug (Poiseuille) type matters because the depth distribution of strain-rates will be different for each case [*Natarov and Conrad, 2012; Becker, 2017; Semple and Lenardic, 2018*]. These effects are likely most relevant for slowly moving plates.

Setting aside these complexities, the comparison between APM and azimuthal anisotropy in the asthenosphere can provide some guidance as to how much of the pattern of anisotropy might be related to convection and, importantly, it does not require any further modeling assumptions. Comparisons with APM have thus been used extensively to explore how anisotropy might be related to mantle flow [e.g. *Montagner and Tanimoto, 1991; Debayle and Ricard, 2013*].

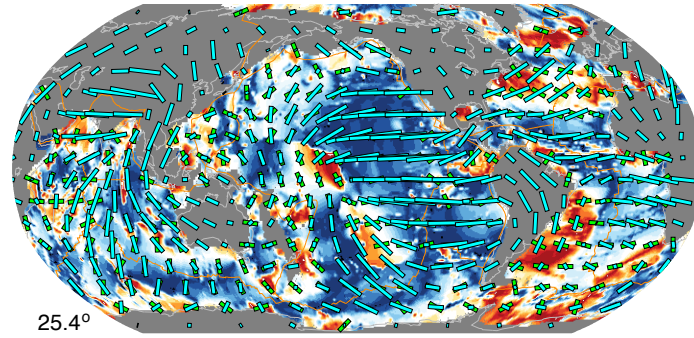
Figure 4b shows such a comparison of azimuthal anisotropy with APM orientations at nominally 200 km depth. Much of the patterns of azimuthal anisotropy in the oceanic regions can be matched by APM alignment, indicating a relationship between flow-induced LPO and seismological constraints. The global oceanic misfit is smaller than for the lithospheric match to paleo-spreading, at average angular misfit $\lesssim 20^\circ$. This is of the order of orientational uncertainties for surface wave studies for azimuthal anisotropy [e.g. *Laske and Masters, 1998; Becker et al., 2003; Ma and Masters, 2015; Schaeffer et al., 2016*]. In this sense, the APM model, its inherently non-physical nature notwithstanding, provides a plausible explanation for asthenospheric anisotropy and confirms that plates are an integral part of mantle convection.

However, there appear to be systematic geographic variations in misfit in the APM asthenospheric match of Figure 4b whose origin is unclear. Moreover, any use of crustal kinematics in an absolute sense, of course, requires a choice of reference frame. Figure 4b uses the spreading-aligned reference frame, which was argued by *Becker et al.* [2015] to provide a parsimonious explanation to a range of constraints for geologically recent plate dynamics. This reference frame is similar to hotspot reference frames with relatively small net rotation of the lithosphere with respect to the deep mantle [e.g. *Ricard et al., 1991; Becker, 2006; Conrad and Behn, 2010*].

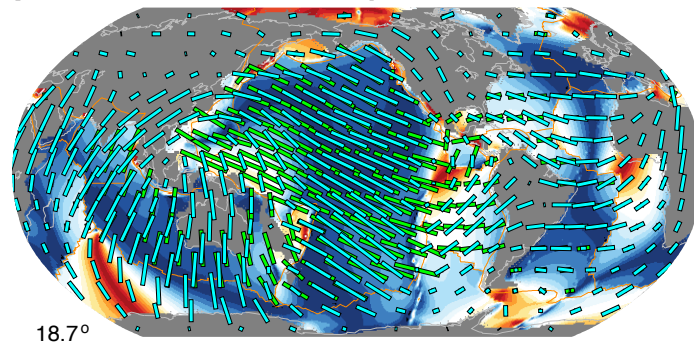
3.3 Mantle circulation modeling

If we seek to make use of our understanding of the physics of mantle circulation instead of comparing anisotropy to APM, we need to approximate the details of mantle flow and LPO formation. In particular, we need to make choices as to how to infer density anomalies and viscosity variations within the mantle. In fact, comparisons of azimuthal

a) spreading - lithosphere



b) APM - asthenosphere



c) LPO from flow - asthenosphere

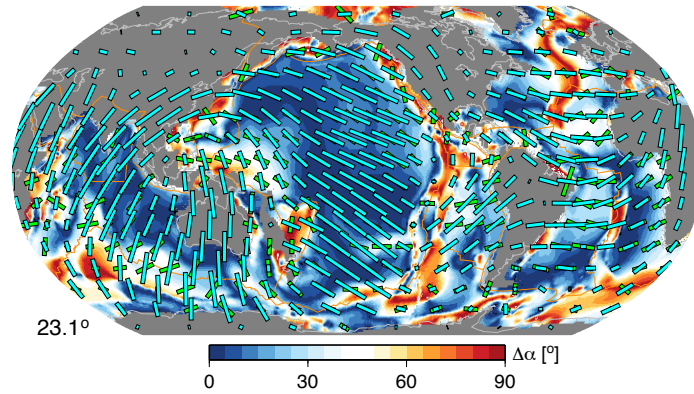


Figure 4. Angular orientational misfit, $\Delta\alpha$, in the oceanic plate regions, computed between azimuthal anisotropy from SL2013SA [cyan sticks; *Schaeffer et al.*, 2016] and geodynamic models (green). **a)** Seismology at 50 km depth vs. paleo-spreading orientations inferred from seafloor age gradients. **b)** SL2013SA at 200 km depth vs. absolute plate motions in the spreading-aligned reference frame [*Becker et al.*, 2015]. **c)** SL2013SA at 200 km depth vs. synthetic anisotropy based on LPO formed in mantle flow [model of *Becker et al.*, 2008]. Numbers in lower left indicate average angular misfit in the oceanic regions. See *Becker et al.* [2014] for more detail on the analysis.

anisotropy with the seminal mantle circulation model of *Hager and O’Connell* [1981] followed soon after [*Tanimoto and Anderson*, 1984].

To arrive at estimates of mantle flow, typically slab structure from seismicity [*Hager*, 1984] or isotropic seismic tomography is scaled to temperature using simplified approximations to what would be inferred from mineral physics and assumptions as to mantle composition [e.g. *Hager et al.*, 1985]. Such circulation model predictions can, for example, explain geoid anomalies as long as there is an increase in viscosity toward the lower mantle [e.g. *Richards and Hager*, 1984; *King and Masters*, 1992], and the associated mantle tractions also provide a powerful explanation for the patterns and rates of plate motions [e.g. *Ricard and Vigny*, 1989; *Forté et al.*, 1991; *Lithgow-Bertelloni and Richards*, 1998; *Becker and O’Connell*, 2001]. However, mantle velocities are strongly dependent on the variable force transmission that results from lateral viscosity variations [e.g. *Conrad and Lithgow-Bertelloni*, 2002; *Becker*, 2006; *van Summeren et al.*, 2012; *Alisc et al.*, 2012], and those will affect strain-rates and hence anisotropy development. In the case of seismic anisotropy, we can thus ask if geodynamic models of mantle flow that are constructed based on other constraints (e.g. geoid or plate motions) also fit seismic anisotropy, and we can use anisotropy to further refine such models.

Assuming that velocities of mantle flow are estimated, we need to make the link to seismic anisotropy. This can be done by simply examining shear in a certain layer of the mantle [i.e. velocity differences; e.g. *Tanimoto and Anderson*, 1984], computing the finite strain ellipsoid (FSE) accumulated along a particle path [e.g. *McKenzie*, 1979; *Ribe*, 1989], or estimating LPO using more complex micro-physical models [e.g. *Ribe and Yu*, 1991; *Wenk and Tomé*, 1999; *Tommasi*, 1998; *Kaminski and Ribe*, 2001; *Blackman et al.*, 2002]. Such approaches have the capability to incorporate laboratory results which indicate the importance of recrystallization during LPO anisotropy formation under sustained shear [e.g. *Zhang and Karato*, 1995; *Bystricky et al.*, 2000]. Experiments also suggest that olivine slip system strength and hence the type of LPO being formed depends on deformation conditions and volatile content [e.g. *Jung and Karato*, 2001; *Katayama et al.*, 2004].

The most common, A-type LPO regime [*Karato et al.*, 2008; *Mainprice*, 2007] appears most prevalent among xenolith and mantle massif samples [*Ben Ismail and Mainprice*, 1998; *Bernard et al.*, 2019]. The corresponding modeled LPO predictions of best-fit hexagonal symmetry axis alignment in flow are broadly consistent with the orientation of the longest FSE axis. Exceptions are regions of strong fabric reworking such as underneath spreading centers or other complex flow scenarios [*Ribe and Yu*, 1991; *Blackman et al.*, 2002; *Kaminski and Ribe*, 2002; *Becker et al.*, 2006a; *Conrad et al.*, 2007]. Other approximations of the LPO such as the infinite strain axis [*Kaminski and Ribe*, 2002] appear to perform less well in comparisons with surface wave based anisotropy than LPO estimates [*Becker et al.*, 2014]. These tests indicate that anisotropy from mantle flow may perhaps be best modeled by either using the FSE [equivalent to whisker orientation in analog experiments; *Buttles and Olson*, 1998] or by computing bulk-approximate [*Goulding et al.*, 2015; *Hansen et al.*, 2016b] or grain-oriented [e.g. *Kaminski et al.*, 2004; *Castelnau et al.*, 2009] descriptions of actual LPO formation, on which we will focus here.

Once LPO is estimated for olivine or olivine-orthopyroxene assemblages by some scheme [e.g. *Kaminski et al.*, 2004], we then need to assign elastic tensors to each virtual grain to compute effective anisotropy. Choices as to the pressure and temperature dependence of elasticity tensor components as well as the averaging scheme have noticeable effects [*Becker et al.*, 2006a; *Mainprice*, 2007], but are likely smaller than uncertainties in seismological imaging on global scales.

Given dramatic improvements in seismological constraints during the twenty years after the fundamental comparison of *Tanimoto and Anderson* [1984], a number of groups revisited mantle circulation modeling in light of azimuthal anisotropy ~ 15 years ago. *Gaboret et al.* [2003] and *Becker et al.* [2003] focused on Pacific and global-scale surface

wave models, respectively, while *Behn et al.* [2004] and *Conrad et al.* [2007] explored matching *SKS* splitting in oceanic plate regions and globally. These models usually find that moving from APM models to mantle flow computations that respect the return flow effects caused by plate motions alone does not improve, or sometimes rather significantly degrades, the fit to seismologically inferred anisotropy. The added physical realism of estimating flow and LPO does come into play once density anomalies are considered for the flow computations, and such models typically outperform APM approaches [*Gaboret et al.*, 2003; *Becker et al.*, 2003; *Behn et al.*, 2004; *Conrad et al.*, 2007; *Conrad and Behn*, 2010; *Becker et al.*, 2014].

Figure 4c shows an example of how LPO formed under dislocation creep in a global circulation model that includes density anomalies [as used in *Becker et al.*, 2008, to study radial anisotropy] matches azimuthal anisotropy [*Schaeffer et al.*, 2016] at asthenospheric depths. While the average misfit for the LPO model is larger than for the comparison with APM (Figure 4b), the regions of large misfit appear now more easily associated with tectonic processes. In particular, large misfits are found underneath spreading centers, where LPO is expected to be reworked [e.g. *Blackman and Kendall*, 2002; *Kaminski and Ribe*, 2002], a process that is as of yet fairly poorly constrained experimentally [*Skemer et al.*, 2012; *Hansen et al.*, 2014, 2016b]. In the models, a consequence of this reworking is that elastic anisotropy locally displays slow axis hexagonal symmetry as well as significant non-hexagonal contributions in regions with pure shear type of flow [*Becker et al.*, 2006a]. Besides, non LPO contributions due to partial melting is expected to matter close to the spreading centers [*Blackman et al.*, 1996; *Blackman and Kendall*, 1997; *Holtzman and Kendall*, 2010; *Hansen et al.*, 2016a]. However, given that regions of large misfit appear confined to “special” places and that all oceanic basins are otherwise fit quite well (Figure 4c), we consider the match of LPO predictions from mantle flow and anisotropy a first-order achievement of “applied geodynamics” [*Gaboret et al.*, 2003; *Becker et al.*, 2003; *Behn et al.*, 2004; *Conrad et al.*, 2007].

The LPO model of Figure 4c relies on the approach of *Becker et al.* [2006a] who computed fabrics using the method of *Kaminski et al.* [2004] along particle paths. Tracers are first followed back in time until, upon iteration, their advective forward paths accumulate a critical finite strain, γ_c , at each observational point. The idea is that any existing textures will be overprinted, and in the case of the example in Figure 4c, $\gamma_c \approx 6$. This choice leads to a good match to radial anisotropy averages and patterns [*Becker et al.*, 2008] as well as regional *SKS* splitting delay times [e.g. *Becker et al.*, 2006b; *Miller and Becker*, 2012], and is consistent with overprinting strains from field and laboratory deformation [*Skemer and Hansen*, 2016].

Assuming that the LPO that is predicted from mantle flow modeling provides at least a statistically appropriate estimate of anisotropy in the upper mantle, we can then use geodynamic models to revisit the hexagonal approximation of seismological imaging. Globally, $\sim 80\%$ of LPO anisotropy is found to be of hexagonal character on average; within regional anomalies, the orthorhombic contribution can reach $\lesssim 40\%$ [*Becker et al.*, 2006a]. This is close to the orthorhombic fraction ($\sim 45\%$) invoked by the subducted asthenosphere model of *Song and Kawakatsu* [2012]. However, on global scales, geodynamic models confirm that the simplifying assumption of hexagonal anisotropy made by seismology appear justified if olivine LPO is the major source of anisotropy in the upper mantle.

The flow computation used in Figure 4c assumes that mantle circulation is stationary over the timescales needed to achieve γ_c . This is a potentially questionable approximation, and time-evolving scenarios expectedly produce larger complexity of LPO predictions, e.g. compared to steady-state subduction scenarios [*Buttles and Olson*, 1998; *Faccenda and Capitanio*, 2013; *Zhou et al.*, 2018]. However, reconstructing the time-evolution of convective flow introduces additional uncertainties due to having to use plate reconstructions and the non-reversibility of the energy equation [e.g. *Steinberger and O’Connell*, 1997; *Conrad*

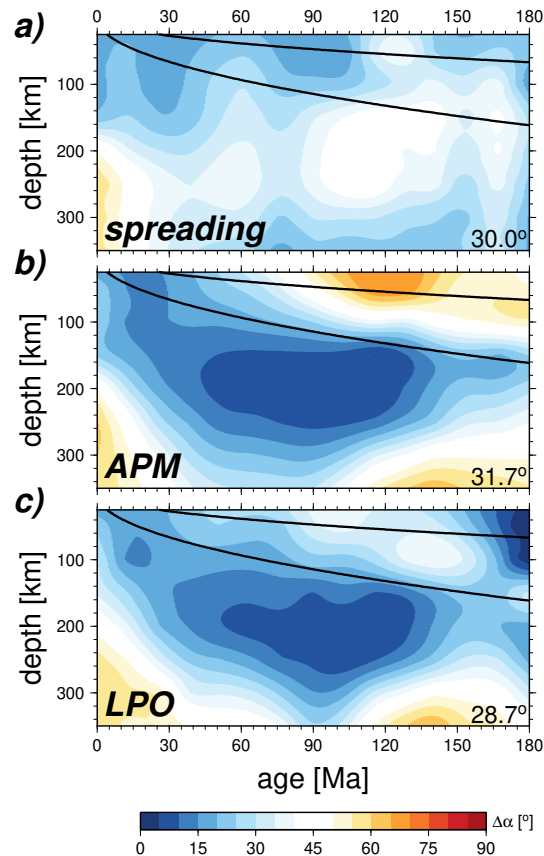


Figure 5. a) Angular orientational misfit, $\Delta\alpha$, underneath the Pacific plate, computed between fast propagation orientations from SL2013SA [Schaeffer *et al.*, 2016] and paleo-spreading orientations, as a function of depth and seafloor age bins (cf. Figure 4a). Black lines are 600°C and 1200°C isotherms from half-space cooling, respectively. b) Angular misfit between azimuthal anisotropy and absolute plate motions in the spreading-aligned reference frame [Becker *et al.*, 2015]. c) Angular misfit between azimuthal anisotropy and synthetics based on computing LPO formation in global mantle flow [model of Becker *et al.*, 2008]. Numbers in lower right indicate average angular misfit for each panel. See Becker *et al.* [2014] for more detail.

and Gurnis, 2003]. More to the point, Becker *et al.* [2003] found that the improvements in terms of the match of anisotropy predictions when allowing for time-dependent mantle circulation were ambiguous. Preliminary tests with newer models confirm that asthenospheric anisotropy predictions are not affected much compared to steady-state approximations as in Figure 5c, as expected given the relatively short advective times. However, the shallower regions within the lithosphere appear somewhat sensitive to which plate reconstruction is used. This provides an avenue for further research.

3.3.1 Boundary layer anisotropy

One of the major achievements of geodynamics is to have linked the bathymetry and heatflow of oceanic seafloor to the half-space cooling of a convective thermal boundary layer [Turcotte and Oxburgh, 1967; Parsons and Sclater, 1977]. Shear in the region below the mechanical boundary layer that is contained within the thermal lithosphere should determine LPO formation [Podolefsky *et al.*, 2004]. This is indeed seen when considering the amplitude of azimuthal anisotropy as a function of seafloor age [e.g. Burgos *et al.*, 2014; Beghein *et al.*, 2014] though alignment with APM is perhaps a better measure as

anisotropy orientations should be better constrained than amplitudes [Debayle and Ricard, 2013].

Figure 5 shows a typical result where the misfit of the three geodynamic models of Figure 4 is shown for the Pacific plate as a function of age. As noted, paleo-spreading is only a good model for the shallowest oceanic lithosphere and relatively young ages. However, alignment with APM or LPO provides a good explanation of azimuthal anisotropy within a 150-200 km thick layer below the $\sim 1200^\circ\text{C}$ isotherm [cf. Burgos *et al.*, 2014; Beghein *et al.*, 2014], as expected given the depth distribution of deformation within the dislocation creep regime [Becker *et al.*, 2008; Behn *et al.*, 2009]. Alignment with both LPO and APM underneath the cold isotherm breaks down at ages older than ~ 150 Ma (cf. Figures 4b and c), perhaps a reflection of small-scale convection [van Hunen and Čadež, 2009].

Comparison of geodynamic predictions with different seismological models leads to similar conclusions [Becker *et al.*, 2014]. However, radial anisotropy does not appear to follow half-space cooling [Burgos *et al.*, 2014], and those discrepancies will be revisited below. The approach of computing mantle circulation and then inferring LPO anisotropy from it to constrain convection, of course, translates to the bottom boundary layer of the mantle as well [Romanowicz and Wenk, 2017], and a separate chapter in this volume is dedicated to this problem.

3.4 Examples of inferences that extend beyond the reference model

As the previous section illustrates, we can indeed use azimuthal seismic anisotropy as a constraint for mantle rheology and upper mantle convection, and in particular arrive at a consistent and quantitative, first-order description of lithosphere-asthenosphere dynamics underneath oceanic plates. The formation of olivine LPO within the “typical” A type slip system under convective flow and plates forming according to half-space cooling appears to provide a globally appropriate geodynamic reference model, and seismic anisotropy is another constraint for plate formation. We should keep in mind the relative success of this “reference” model (e.g. Figures 4 and 5) as we move on to briefly discuss some of the more indirect inferences based on seismic anisotropy, and in particular when we conclude by discussing regional or process level complications.

Mantle flow is driven by density anomalies and modulated by viscosity, and in theory both of these can be inverted for using seismic anisotropy assuming it is formed by the shear due to spatial variations in velocity. In practice, additional constraints are needed for all but the simplest tests. One important question in mantle dynamics is that of the appropriate reference frame for surface motions with respect to the lower mantle. Different reference frames yield a range of estimates for trench advance or rollback, for example [e.g. Chase, 1978; Funicello *et al.*, 2008] with implications for regional tectonics and orogeny.

Given that seismic anisotropy due to LPO is formed under the shear that corresponds to the motion of the surface relative to the stagnant lower mantle, one may thus postulate that the best APM is that which minimizes the misfit to anisotropy. This was addressed by Kreemer [2009] based on SKS splitting and explored by Montagner and Anderson [2015] for surface waves and individual plate motions with focus on the Pacific. The spreading-aligned reference of Figure 4b naturally minimizes the misfit with a number of surface-wave based estimates of azimuthal anisotropy and their individual best-fit poles are very similar. This implies that the anisotropy-constrained reference frame may have general relevance, with implications for the relative strength of transform faults, for example [Becker *et al.*, 2015].

One can also use mantle circulation modeling to explore the depth-distribution of the shear that corresponds to different degrees of net rotation of the lithosphere [Zhong,

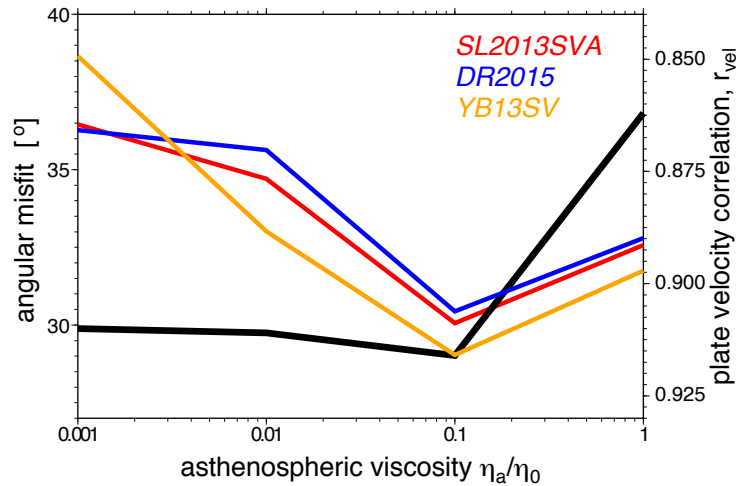


Figure 6. Angular misfit (minimum with depth) between flow model predictions and azimuthal seismic anisotropy for oceanic domain for three different seismological models: SL2013SVA [Schaeffer *et al.*, 2016, as in Figure 4], DR2015 [Debaille and Ricard, 2013], and YB13SV [Yuan and Beghein, 2013] (colored lines), and misfit of predicted plate velocities (black line), as a function of asthenospheric viscosity reduction for a 300 km thick layer. Modified from Becker [2017], see there for detail.

2001; Becker, 2006], and then test how such a shear component would affect the match of global circulations models to seismic anisotropy. This exploits the fact that the match to anisotropy is sensitive to where in the mantle shear is localized [Becker *et al.*, 2003; Conrad and Behn, 2010]. Becker [2008] used the match to surface wave based azimuthal anisotropy to argue that net rotation should be less than $\lesssim 0.2^\circ/\text{Myr}$. Conrad and Behn [2010] considered both *SKS* splitting and surface wave anisotropy and further explored this “speed limit” on net rotation. They find a permissible net rotation of $\sim 0.25^\circ/\text{Myr}$ for an asthenospheric viscosity that is one order of magnitude smaller than that of the upper mantle.

Using models which self-consistently generate plate motions, Becker [2017] showed that anisotropy constraints on asthenospheric viscosity are consistent across different modern azimuthal anisotropy models, and that even slab-driven flow alone leads to perturbations of the large-scale match of LPO anisotropy from flow that is seen in Figure 4c. A moderate sub-oceanic viscosity reduction to $\sim 0.01 \dots 0.1$ times the upper mantle viscosity is strongly preferred by both the model match to azimuthal anisotropy and the fit to plate motions (Figure 6), even though there exists a typical trade-off with layer thickness [Richards and Lenardic, 2018]. In particular, suggested high partial melt, lubricating zones underneath oceanic plates [e.g. Kawakatsu *et al.*, 2009; Schmerr, 2012] appear to not be widespread enough to affect large scale mantle shear, or else it should be seen in the match to seismic anisotropy [Becker, 2017].

On a regional scale, *SKS* splitting provides better lateral resolution than traditional surface wave analyses and is thus widely used to infer the role of mantle flow for tectonics, particularly within continental plates (Figure 2). When combined with flow models, we can exploit the sensitivity of mantle circulation to density anomalies and viscosity variations [e.g. Fouch *et al.*, 2000; Hall *et al.*, 2000; Behn *et al.*, 2004; Becker *et al.*, 2006b]. This was done by Miller and Becker [2012] in a quasi-inverse sense, exploring a large number of global mantle flow computations with a range of density and viscosity models to test which (in particular with respect to continental keel geometry and strength) are consistent with *SKS* splitting in NE South America. A similar approach was used on

a larger-scale for South America by *Hu et al.* [2017], and *Faccenna et al.* [2013] to infer a low viscosity channel underneath the Red Sea, for example.

Another possible approach to explore the effects of mantle convection, helpful in the absence of good isotropic tomography for example, is to test different forward models of the effect of density anomalies, e.g. compared to plate-scale flow for plumes [e.g. *Walker et al.*, 2005; *Ito et al.*, 2014] or details of subduction and delamination scenarios [e.g. *Zandt and Humphreys*, 2008; *Alpert et al.*, 2013]. In such regional contexts, mantle flow models provide the capability to explore the impact of depth variations in seismic anisotropy, as those are ideally recorded in the back-azimuthal dependence of *SKS* splitting [e.g. *Blackman and Kendall*, 2002; *Hall et al.*, 2000; *Becker et al.*, 2006b]. Subduction zone *SKS* splitting anisotropy is, however, complex to the extent that the correct background model becomes questionable [e.g. *Long*, 2013], as discussed in a different chapter of this volume.

A question to ask whenever seismic anisotropy observations are used to infer mantle dynamics is how consistent any model is with a range of observations besides the anisotropy data, e.g. in terms of the geoid, dynamic topography, or plate motions. Some studies invoke different effects that may impact mantle flow (e.g. a small-scale plume, inherited structure in the lithosphere, volatile variations in the asthenosphere) for nearly every single different *SKS* split, often without any consistent flow modeling, and so trivially explain all data perfectly in the extreme case. Other studies, such as the approach illustrated in Figure 4c, strive for a broad-scale match to the observations, within an actual geodynamic framework that respects continuum mechanics conservation laws. This can then invite further study as to which effects (e.g. intraplate deformation, volatile variations of frozen in structure) may be required regionally on top of the reference model. Clearly, there is a continuum between those quasi end-members.

4 Open questions

4.1 Regional complexities and scale-dependent resolution

Navigating between the extremes of a possibly very complicated model or simulation that matches all data, and a simple model which may or may not be a good reference given large regional misfits is, of course, not an uncommon challenge in the Earth sciences. However, the complexities of anisotropy, both in terms of spatially variable resolution and in terms of possible mechanisms for anisotropy generation, seem to make these trade-offs more acute for efforts of linking anisotropy to mantle flow.

4.1.1 Oceanic plates revisited

The previous discussion of convection dynamics as seen by seismic anisotropy focused on large spatial scales and seismic models that are derived from global surface wave datasets. *SKS* splitting for oceanic island stations (Figure 2) are also usually well fit by the density-driven mantle circulation models [e.g. *Behn et al.*, 2004; *Conrad et al.*, 2007]. Increasingly, we can interpret results from ocean bottom seismometer deployments which slowly infill the oceanic plates in terms of high resolution and high quality regional constraints [e.g. *Isse et al.*, 2019]. In particular, deployments that are designed to image “normal” or at least “melt free” oceanic plates are very valuable to further develop the thermo-mechanical reference model of plate generation that was alluded to previously [e.g. *Lin et al.*, 2016; *Takeo et al.*, 2018; *Kawakatsu and Utada*, 2017; *Russell et al.*, 2019].

Alas, the regional results are often at odds with inferences from global models, particularly when it comes to the variation in strength of radial and azimuthal anisotropy with depth. This has long been debated and results are sensitive to the dataset selection and applied corrections [e.g. *Ferreira et al.*, 2010; *Ekström*, 2011; *Rychert and Harmon*,

2017]. For presumably typical oceanic lithosphere, there is evidence for radial anisotropy with $v_{SH} > v_{SV}$ in the lithosphere [e.g. *Gaherty et al.*, 1996; *Russell et al.*, 2019], but regional [*Takeo et al.*, 2013] and many global models show a deeper peak in radial anisotropy [$\sim 80 \dots 150$ km, e.g. *Nettles and Dziewoński*, 2008; *French and Romanowicz*, 2014; *Auer et al.*, 2014; *Moulik and Ekström*, 2014]. The latter would be more consistent with a geologically recent, convective LPO origin of radial anisotropy.

Moreover, while azimuthal anisotropy strength appears to follow half-space cooling similar to the region of low angular misfit in Figure 5, radial anisotropy appears to have limited seafloor age dependence [*Burgos et al.*, 2014; *Beghein et al.*, 2014; *Auer et al.*, 2015; *Isse et al.*, 2019]. This might indicate that LPOs due to spreading and/or petrological heterogeneities [e.g. *Kawakatsu et al.*, 2009; *Schmerr*, 2012; *Sakamaki et al.*, 2013; *Ohira et al.*, 2017] are frozen in during the generation of plates [e.g. *Beghein et al.*, 2014; *Auer et al.*, 2015; *Russell et al.*, 2019], or that the effects of melt induced LPO mask the age dependence that would be expected [*Hansen et al.*, 2016a].

At least some mid-lithospheric structure appears required for oceanic plates that is unrelated to simple LPO anisotropy [*Rychert and Harmon*, 2017], perhaps indicating a mid lithospheric discontinuity similar to what has been discussed for continental plates [e.g. *Yuan and Romanowicz*, 2010; *Selway et al.*, 2015]. However, the mechanical effects of the lithospheric component of convection in terms of asthenospheric shear appear to be captured by half-space cooling and azimuthal anisotropy as reflected in Figure 5 [*Becker et al.*, 2014].

4.1.2 Anisotropy in continents

Referring to his analysis of P wave seismic anisotropy in oceanic plates, *Hess* [1964] noted that “the structure and history of the whole ocean floor can probably be worked out much more rapidly than the more complicated land surface of the Earth”, and this has been very much true. Thermal mantle convection explains the motions of the nearly rigid oceanic plates well, but the distributed and protracted deformation record of the continental lithosphere is strongly affected by rheological and compositional effects, and still presents many questions.

One way to explore the consistency between different ways of imaging anisotropy is to consider the match between SKS splitting estimates and surface wave based azimuthal anisotropy in continents where SKS splits are preferentially measured, because of station logistics (Figure 2). *Montagner et al.* [2000] conducted such a comparison using an approximate method and found that agreement in terms of the fast azimuth pattern was limited. A more positive assessment with a more complete SKS dataset was provided by *Wüstefeld et al.* [2009] who found a match at the longest wavelengths. *Becker et al.* [2012] revisited this comparison and showed that full wave form modeling of SKS splitting leads to only moderate differences in the apparent splitting compared to averaging for the current generation of global surface models. Indeed, the agreement in terms of fast azimuths is limited when SKS splits are smoothed over less than ~ 2000 km length scales. This does at least partly reflecting the inherent resolution limits of global models [e.g. *Debayle and Ricard*, 2013; *Schaeffer et al.*, 2016]. Most surface wave models also under-predict SKS delay times, likely because of the required regularization [cf. *Schaeffer et al.*, 2016].

Figure 2b shows that the long-wavelength correlation between SKS fast axis patterns and different surface wave models is best in the upper ~ 400 km of the mantle, where azimuthal anisotropy models show the largest power and are most in agreement with each other [*Becker et al.*, 2012; *Yuan and Beghein*, 2013]. This finding is consistent with a common origin, the suggested dominance of an uppermost mantle, asthenospheric anisotropy to typical SKS splitting results [e.g. *Fischer and Wiens*, 1996; *Silver*, 1996], and an LPO induced by mantle flow origin [Figure 5; e.g. *Becker et al.*, 2008, 2014]. Figure 2b also indicates a hint of a drop in correlation in the lithosphere, and regionally, it is

clear that there is both depth-dependence in the surface-wave imaged anisotropy and the match with *SKS* splitting [e.g. *Deschamps et al.*, 2008b; *Yuan et al.*, 2011; *Lin et al.*, 2016; *Takeo et al.*, 2018].

Such discrepancies motivate an alternative approach that exploits the difference in depth sensitivity between *SKS* and Rayleigh waves [*Marone and Romanowicz*, 2007; *Yuan and Romanowicz*, 2010]. One can use surface waves to constrain the shallow part of a model, say above ~ 250 km, and then use the complementary resolution of the path component of teleseismic waves beneath that region to exploit *SKS* constraints on anisotropy. Along with constraints from radial anisotropy [e.g. *Gung et al.*, 2003] and alignment of 2φ anisotropy with APM motions, such models have been used to infer a thermo-chemically layered structure of the North American craton [*Yuan and Romanowicz*, 2010; *Yuan et al.*, 2011], for example.

Given the now more complete coverage of the continental U.S. with *SKS* splitting thanks to USArray (Figure 2a), we recently revisited the question of agreement between surface wave models and *SKS*, and found that *SKS* remain poorly matched even by newer surface wave models. An exception is the model of *Yuan et al.* [2011] for most of North America, but that model attempts to fit *SKS* splits by design. This implies that the depth resolution of large-scale surface wave models is, on a continent-scale, not at the level where details in possible anisotropic layering could be consistently determined, particularly below ~ 200 km [*Yuan et al.*, 2011].

4.2 Uncertainties about microphysics

4.2.1 Formation of olivine LPO

There is now a range of experimental work that documents how olivine develops LPO under shear. Among the modern studies, *Zhang and Karato* [1995] and the large-strain experiments of *Bystricky et al.* [2000] showed how olivine *a*-axes cluster within the shear plane for the common, A type fabric, or the high-stress D type. A-type LPO is found most commonly in natural samples [*Ben Ismail and Mainprice*, 1998; *Bernard et al.*, 2019], and provides the most straightforward link between anisotropy and flow as was explored in section 3.3.

However, *Jung and Karato* [2001] and *Katayama et al.* [2004] found that additional LPO types can develop depending on deviatoric stress, temperature, and water content conditions, and mineral physics modeling approaches can reproduce those fabrics by assigning different relative strength to the major olivine slip systems [*Kaminski*, 2002; *Becker et al.*, 2008]. The predictions for anisotropy can be markedly different from A: The B (high water, high stress/low temperature type) would lead us to expect azimuthal anisotropy to be oriented perpendicular to shear, and this might explain some of the subduction zone complexities, though likely not all trench-parallel splitting [*Kneller et al.*, 2005; *Las-sak et al.*, 2006]. Effective B types are also seen in high partial melt experiments [*Holtzman et al.*, 2003]. The C (high water, low stress/high temperature) type is expressed such that $v_{SV} > v_{SH}$ under horizontal shear, implying a complete reorientation of the relationship between flow and radial anisotropy [*Karato et al.*, 2008; *Becker et al.*, 2008].

At present, it is not entirely clear how olivine LPO formation depends on deformation conditions, in particular in conjunction with ambient pressure [*Karato et al.*, 2008]. For example, *Ohuchi et al.* [2012] substantiated a transition from A \rightarrow B LPO as a function of water content at low confining pressures. However, *Ohuchi and Irifune* [2013] showed that the dependence on volatile content appears reversed at higher pressures (below ~ 200 km depth), such that A would be the high-volatile content LPO, and changes in LPO perhaps mainly depth-dependent [cf. *Mainprice et al.*, 2005; *Jung et al.*, 2009].

If we consider the range of natural xenolith and ophiolite samples, all of the LPO types documented in the lab are indeed found in global compilations [Mainprice, 2007; Bernard *et al.*, 2019]. However, samples from individual sites show a range of different LPOs, at presumably similar deformation conditions in terms of volatile content and deviatoric stress. Moreover, when such conditions are estimated from the samples, no clear systematics akin to the laboratory-derived phase diagrams arise [Bernard *et al.*, 2019]. This implies that the style and history of deformation may be more important in controlling natural sample LPO, and perhaps, by inference, seismic anisotropy, particularly in the lithosphere where deformation is commonly more localized and where preservation potential is high.

Besides these uncertainties regarding the formation of different LPO types under dislocation creep by changes in slip system activity due to ambient conditions, it is not straightforward to predict where dislocation creep should dominate [Hirth and Kohlstedt, 2004]. The estimate of predominant dislocation creep between $\sim 100 - 300$ km for grain sizes of order mm [Podolefsky *et al.*, 2004] is compatible with the depth distribution of radial anisotropy, for example [Becker *et al.*, 2008; Behn *et al.*, 2009]. However, using different assumptions about grain growth and evolution, Dannberg *et al.* [2017] constructed mantle convection models which are consistent with seismic attenuation, but would predict the dominance of diffusion creep within the asthenosphere. The assumption that LPOs only form under dislocation creep and are preserved or destroyed under diffusion creep has been challenged [e.g. Sundberg and Cooper, 2008; Miyazaki *et al.*, 2013; Maruyama and Hiraga, 2017], but it remains to be seen where most of the discrepancies arise.

While geodynamic modeling can easily incorporate stress, temperature, and pressure-dependent changes in slip systems for LPO predictions, for example, it is thus not clear if such a modeling approach is warranted. If volatile content is used as a free parameter, for example, we can construct upper mantle models with a range of seismic anisotropy predictions for the exact same convective model. This is not the most satisfying situation, unless other constraints such as from magneto-tellurics or phase boundary deflections provide further constraints on volatile variations. Moreover, the good match of the geodynamic reference model to azimuthal and radial anisotropy discussed in section 3.3 implies that A type LPO, formed under dislocation creep, may well be dominant in the upper mantle.

4.2.2 Mechanical anisotropy

Besides seismic, we also expect mechanical anisotropy as a result of the formation of LPO, i.e. the deformation behavior of olivine will depend on the sense and type of shear. Such viscous anisotropy is one potential mechanism for lithospheric strain-localization and deformation memory in plate boundaries [e.g. Tommasi *et al.*, 2009; Montési, 2013]. Mechanical anisotropy has been documented in the laboratory for olivine LPO [Hansen *et al.*, 2012], and was implemented in microstructural modeling approaches [Castelnau *et al.*, 2009; Hansen *et al.*, 2016c].

We expect viscous anisotropy to increase the wavelength of convection [Honda, 1986; Busse *et al.*, 2006] and localize flow in the asthenosphere if the weak plane is aligned with plate shear, possibly stabilizing time-dependent plate motions [Christensen, 1987; Becker and Kawakatsu, 2011]. The response of a mechanically anisotropic layer will also lead to a modification of the growth rates of folding or density driven instabilities [Mühlhaus *et al.*, 2002; Lev and Hager, 2008]. However, trade-offs between isotropically weakened layers and anisotropic viscosity may make any effects on post-glacial response, the geoid, or the planform of convection hard to detect [Han and Wahr, 1997; Becker and Kawakatsu, 2011], in analogy to the bulk seismic anisotropy of a layered medium with isotropic velocity variations [Backus, 1962].

On regional scales, the effect of viscous anisotropy may be more easily seen in tectonic or dynamic observables, and any treatment of the development of texture should in

principle account for the mechanical effects of LPO formation on flow for self-consistency and to account for possible feedback mechanisms [Knoll *et al.*, 2009]. The joint development of mechanical and seismic anisotropy may be of relevance in subduction zones, where seismic anisotropy is widespread, and time-dependent fluctuations in the mantle wedge temperature may result [Lev and Hager, 2011].

Self-consistent modeling of both seismic and viscous anisotropy was implemented by Chastel *et al.* [1993] for an idealized convective cell, and more recently by Blackman *et al.* [2017] for a more complete convection model of a spreading center. Models that include the LPO feedback show generally similar flow patterns than simpler models, but there can be up to a factor ~ 2 enhancement of predicted surface wave azimuthal anisotropy close to the ridge because of increased strain-rates, and the transverse isotropy symmetry axes are more horizontally aligned [Blackman *et al.*, 2017] than those of earlier one-way LPO predictions [Blackman and Kendall, 2002]. It remains to be seen if such effects of viscous anisotropy feedback are relevant for the interpretation of regional or global convection, or if other uncertainties such as the effects of temperature, composition, and volatile anomalies on isotropic olivine rheology swamp the signal.

5 Ways forward

Our general understanding of upper mantle convection thus appears to be reflected in seismic anisotropy, and anisotropy allows broad inferences on asthenospheric viscosity and regional tectonics, for example. Yet, many uncertainties remain and become most acute if the strength of anisotropy is to be exploited quantitatively. What can we do to raise the water levels of this glass half-full scenario?

For one, more data, and in particular more seafloor, or oceanic realm, observations, as well as dense continental seismometer deployments, certainly help. Higher density imaging should resolve many of the uncertainties including the depth distribution of seismic anisotropy in oceanic plates over the next decade. Seismometer arrays such as USArray have transformed our view of mantle structure under continental plates, even though much work is still to be done to integrate the newly imaged complexity into dynamic models of mantle evolution. Availability of high-density passive seismic data means better resolution for surface wave studies. Moreover, many deployments are now also sampling the upper mantle with strongly overlapping Fresnel zones (3-D sensitivity kernels) for *SKS* splitting [Chevrot *et al.*, 2004; Long *et al.*, 2008]. Using methods that make use of the array station density [Ryberg *et al.*, 2005; Abt and Fischer, 2008; Monteiller and Chevrot, 2011; Lin *et al.*, 2014; Mondal and Long, 2019] rather than presenting individual splits without concern as to the likely implications of back-azimuthal dependencies and overlapping sensitivity is still the exception, however. It should become the rule, even if the methodological burden is higher.

Yet, even for well-constrained regions, at least some trade-offs between structural model parameters will likely persist, which is when inversion choices as to parameterization and regularization become even more important. Any mixed- or under-determined inverse problem requires regularization, and often choices are made based on the intuition, or preconceptions, of seismologists as to the degree of isotropic and anisotropic heterogeneity. Different structural representations can result depending on the treatment of the preferred spectral character of isotropic and anisotropic heterogeneity. One way to approach the problem is to quantify the statistical character of heterogeneity of anisotropy and isotropic velocity anomalies (e.g. due to temperature and compositional variations) from field observations or convection modeling [e.g. Holliger and Levander, 1992; Becker *et al.*, 2007b; Kennett *et al.*, 2013; Alder *et al.*, 2017], and then have those properties guide regularization.

Another, more narrow, but perhaps in our context more productive, way to introduce a priori information is to use assumptions about the symmetry types of anisotropy and conditions for the formation of certain LPO types for imaging or joint seismological and geodynamic inversion. This spells out a project to integrate as much information about the link between seismic anisotropy and convection from laboratory and field work, to seismology, to geodynamic modeling for a better understanding of the evolution of plate tectonics. Once firm links are established, such an approach should, in principle, allow extending the use of seismic anisotropy much further back in time than the last few 10s of Myr, if we are able to capture the lithospheric memory of “frozen in” structure alongside the asthenospheric convection contribution.

The simplifications of hexagonal symmetry axes being oriented vertically (radial anisotropy) or horizontally (azimuthal anisotropy) is one example of imposing a priori assumptions to simplify imaging. More generally, we can solve for the overall orientation of the symmetry axes of a medium with hexagonal anisotropy, for example. This approach is called vectorial tomography [Montagner and Nataf, 1988; Chevrot, 2006] and has been in use for a long time [Montagner and Jobert, 1988], though not widely so. Vectorial tomography exploits the fact that individual minerals such as olivine show certain characteristics which link different elastic parameters, allowing reduction in the parameter space that has to be explored by a seismological inversion [Montagner and Anderson, 1989; Plomerová et al., 1996; Xie et al., 2015]. Similar relationships between parameters such as P and S wave anisotropy, for example, can be established for upper mantle LPO assemblages [Becker et al., 2006a, 2008] or crustal rocks [Brownlee et al., 2017], and the resulting scaling relationships can improve inversion robustness [Panning and Nolet, 2008; Chevrot and Monteiller, 2009; Xie et al., 2015; Mondal and Long, 2019].

Use of such petrological scalings limits the interpretation to, say, determining the orientation and saturation of a certain type of olivine LPO in the mantle, rather than being general and allowing for other causes of anisotropy. However, the images of lateral variations should be more robust than the general inversion which will itself be subject to other assumptions, even if it is just regularization. Moreover, different assumptions as to which types of LPOs might be present can also be tested in a vectorial tomography framework. In this context, surface wave anomalies in 2φ and 4φ patterns [e.g. Montagner and Tanimoto, 1990; Trampert and Woodhouse, 2003; Visser et al., 2008] also appear underutilized. Based on a simple petrological model for peridotite, Montagner and Nataf [1986] showed that mantle-depth Rayleigh and Love waves should be mainly sensitive to the 2φ and 4φ signal, respectively. However, there is evidence for additional Rayleigh and Love wave structure in 4φ and 2φ , respectively, for the mantle [e.g. Trampert and Woodhouse, 2003], and such signals are often seen for the crust [cf. Figure 3; e.g. Adam and Lebedev, 2012; Xie et al., 2015]. For many petrological models, there exist strong relationships between the 2φ and 4φ signature in each wave type, meaning that Rayleigh and Love waves can be inverted jointly for azimuthal anisotropy for certain a priori assumptions, yet this is not often done [e.g. Xie et al., 2015; Russell et al., 2019] as horizontal records are usually noisier than vertical seismograms. Moreover, the sensitivity of each surface wave type depends on the dip angle and olivine LPO type which might be diagnostic. This could be further utilized in vectorial tomography imaging, particularly for joint surface and body wave inversions [Romanowicz and Yuan, 2012].

In order to proceed with such joint inversions, we need to better understand the predictions of laboratory experiments as to LPO formation under different deformation conditions. This requires more experimental work, particularly under low deviatoric stress [Skemer and Hansen, 2016; Bernard et al., 2019]. Moreover, a better handle on the degree to which dislocation creep dominates LPO formation needs to be established. Using improved micro-structural relationships, we can also revisit the treatment of grain-size evolution to better explore which parts of the mantle should form anisotropy under what conditions. Among other improvements of time-dependent mantle flow models, such a con-

sistent picture of the across-scale deformation of the upper mantle will not only provide better constraints on convection, but also clarify the role of grain-size evolution and inherited fabrics for the formation and preservation of plate boundaries [e.g. *Tommasi et al.*, 2009; *Montési*, 2013; *Bercovici et al.*, 2015].

6 Conclusions

So, which is it, can of worms or the most useful constraint? We think that the pitfalls of non-uniqueness and uncertainties in the relationship between convection and seismic observations can be overcome. Using a combination of targeted laboratory experiments, further comparison with field analogs and samples, improved seismological imaging, and geodynamic modeling we can achieve a powerful, integrated interpretation of seismic and other data and models, including those from geodesy, the geological record, mineralogy and geodynamics. In particular, we look forward to seeing more work using vectorial tomography in multi-disciplinarily constrained, densely sampled key study areas. The promise of being able to potentially resolve the depth distribution of shear in convection, and hence mantle rheology, and the potential to unlock the deformation memory of both continental and oceanic plates for a new set of constraints on the mechanisms governing their evolution make the headaches worth while.

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