Global Mantle Flow and the Development of Seismic Anisotropy: Differences Between the Oceanic and Continental Upper Mantle

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Viscous shear in the asthenosphere accommodates relative motion between the Earth’s surface plates and underlying mantle, generating lattice-preferred orientation (LPO) in olivine aggregates and a seismically anisotropic fabric. Because this fabric develops with the evolving mantle flow field, observations of seismic anisotropy directly constrain flow patterns if the contribution of lithospheric fossil anisotropy is small. We use global viscous mantle flow models to characterize the relationship between asthenospheric deformation and LPO, and constrain the asthenospheric contribution to anisotropy using a global compilation of observed shear-wave splitting measurements. For asthenosphere \( >500 \text{ km} \) from plate boundaries, simple shear rotates the LPO toward the infinite strain axis (ISA, the LPO after infinite deformation) faster than the ISA changes along flow lines. Thus, we expect ISA to approximate LPO throughout most of the asthenosphere, greatly simplifying LPO predictions because strain integration along flow lines is unnecessary. Approximating LPO~ISA and assuming A-type fabric (olivine \( a \)-axis parallel to ISA), we find that mantle flow driven by both plate motions and mantle density heterogeneity successfully predicts oceanic anisotropy (average misfit = 12°). Continental anisotropy is less well fit (average misfit = 39°), but lateral variations in lithospheric thickness improve the fit in some continental areas. This suggests that asthenospheric anisotropy contributes to shear-wave splitting for both continents and oceans, but is overlain by a stronger layer of lithospheric anisotropy for continents. The contribution of the oceanic lithosphere is likely smaller because it is thinner, younger and less deformed than its continental counterpart.
1. Introduction

Convection in the Earth’s mantle drives the tectonic motions of Earth’s lithospheric plates as well as viscous deformation of the mantle rocks beneath them [e.g., Turcotte and Oxburgh, 1967]. Above a depth of ~300 km [e.g., Hirth and Kohlstedt, 2003], this deformation occurs as dislocation creep [Karato and Wu, 1993], which aligns olivine crystals into a lattice preferred orientation (LPO) [McKenzie, 1979; Ribe, 1989]. Because olivine crystals are anisotropic [e.g., Verma, 1960], the development of LPO produces a macroscopic anisotropy in the uppermost upper mantle that can be detected seismically [e.g., Hess, 1964; Forsyth, 1975; Montagner, 1994]. This implies that observations of seismic anisotropy can in principle be used to constrain the sub-lithospheric mantle deformation that produces this anisotropy. However, the development of the LPO in upper mantle rocks is dependent on the finite strain history of olivine fabric [e.g., Ribe, 1992] as well as the effects of dynamic recrystallization, sub-grain rotation, and grain-boundary migration [e.g., Zhang and Karato, 1995]. In addition, the presence of water may change the anisotropic fabric that olivine aggregates develop in response to flow deformation [Jung and Karato, 2001]. Taking these effects into account potentially introduces significant complexity into the prediction of LPO from a given mantle flow field [e.g., Kaminski and Ribe, 2001; Blackman et al., 2002a; Kaminski et al., 2004]. This complexity has generated controversy over how observations of seismic anisotropy should be used to constrain geodynamic models of mantle flow [e.g., Savage, 1999].

Despite the uncertainty over the relationship between mantle flow and LPO, several studies have used observations of seismic anisotropy to constrain geodynamic models. Some studies simply use the direction of observed anisotropy as a direct indicator of the direction of mantle flow. Doing so implicitly assumes that the anisotropic A-type fabric, which orients anisotropy subparallel to the direction of maximum shear, dominates for upper mantle olivine [Jung and Karato, 2001]. Basic patterns of flow around slabs [e.g., Russo and Silver, 1994; Peyton et al., 2001], hotspots [e.g., Walker et al., 2005; Xue and Allen, 2005], and ridges [e.g., Wolfe and Solomon, 1998] have been deduced in this way. An alternative approach is to use seismic anisotropy to evaluate forward models of mantle flow. Recent studies have used global models of mantle flow driven by plate motions and (tomographically-inferred) mantle density heterogeneity to predict seismic anisotropy. Despite employing different degrees of complexity in the prediction of LPO from flow, these studies tend to successfully predict the direction of anisotropy in oceanic regions. For example, several studies have simply used the horizontal projection of the instantaneous (strain-rate) maximum shear direction to predict anisotropy observed from SKS shear-wave splitting measurements [Behn et al., 2004; Hammond et al., 2005] or surface wave anisotropy [Gaboret et al., 2003]. Becker et al. [2003] used a tracer method to account for finite strain accumulation when predicting anisotropy inferred from surface wave inversions. More recently, Becker et al. [2006a; 2006b] added the effects of dynamic recrystallization and subgrain rotation [Kaminski and Ribe, 2001; Kaminski et al., 2004] to the prediction of anisotropy. Despite differences in how LPO direction is inferred from the geodynamic flow field, these global studies all do an excellent job of predicting anisotropy in oceanic regions. In the analysis below, we will show why this is the case and in doing so introduce some simplicity into the discussion over how to best predict LPO direction from models of mantle flow.

Upper mantle anisotropy has been attributed to both asthenospheric and lithospheric sources. We expect anisotropy to be actively forming in the asthenosphere...
because the relative motion between the surface plates and the upper mantle is accommodated in this low viscosity zone by ongoing shear deformation [e.g., Park and Levin, 2002]. In contrast, because of its strength most mantle lithosphere is not currently experiencing active deformation. Thus, present-day plate motions probably do not induce lithospheric anisotropy, except in areas of active rifting or orogeny [e.g., Silver, 1996]. However, because the lithosphere is cold, it may preserve an anisotropic fabric associated with past deformation [e.g., Savage, 1999; Silver et al., 2001, 2004, 2006]. Thus, while recent mantle flow may control asthenospheric anisotropy, lithospheric anisotropy likely contains anisotropic fabric unrelated to flow in the present-day mantle. This basic difference between asthenospheric and lithospheric anisotropy may limit the usefulness of surface observations of anisotropy for constraining present-day mantle flow patterns in regions where the lithospheric contribution is large. However, if the influence of lithospheric anisotropy can be removed (e.g., by inverting shear-wave splitting data for a two-layer model [Silver and Savage, 1994] or utilizing the depth dependence of surface waves) the remaining asthenospheric anisotropy can place useful constraints on present-day mantle flow. First, however, the relative importance of lithospheric and asthenospheric anisotropy must be characterized for different tectonic regions.

The most obvious tectonic variation on the Earth’s surface is the one between oceanic and continental lithosphere. There are several differences between these types of lithosphere that may induce variations in anisotropy with depth. Continental lithosphere is typically older, thicker, and has a longer deformation history than oceanic lithosphere. For this reason, Silver [1996] hypothesized that preserved fossil anisotropy, unrelated to present-day mantle flow, is more prevalent in continental lithosphere than in oceanic lithosphere. The upper mantle viscosity structure is also probably simpler beneath the oceans. Although the thickness of oceanic lithosphere increases with age, it is thought to have a fairly uniform value of ~100 km for ages greater than ~50 Ma [e.g., Stein and Stein, 1992]. By contrast, continental lithosphere features “roots” that penetrate up to ~400 km beneath cratonic shields [e.g., Jordan, 1975; Ritsema et al., 2004]. These deeply penetrating roots are likely cold and highly viscous relative to the surrounding asthenosphere [Rudnick et al., 1998], and may extend all the way through the asthenosphere or may be underlain by a low-viscosity asthenospheric channel [Gung et al., 2003]. Such large lateral variations in viscosity will significantly influence the asthenospheric flow field beneath continents [Fouch et al., 2000]. By contrast, it is likely that the viscosity structure beneath oceans more closely resembles the layered structures that have been employed by most mantle flow models to date [e.g., Becker et al., 2003; Gaboret et al., 2003; Behn et al., 2004]. Although lateral viscosity variations have been used in studies that predict lithospheric stresses [Conrad and Lithgow-Bertelloni, 2006], the geoid and dynamic topography [Zhang and Christensen, 1993; Çadek and Fleitout, 2003], net lithosphere rotation [Zhong, 2001; Becker, 2006], and plate motions [Wen and Anderson, 1997; Becker, 2006]; only recently have upper mantle lateral viscosity variations been included in geodynamic models that predict seismic anisotropy [Becker et al., 2006a]. In this study, we investigate the influence of lateral viscosity variations in an effort to better predict asthenospheric anisotropy. These predictions are then used to characterize the relative importance of asthenospheric and lithospheric anisotropy in both continental and oceanic regions.
2. Development of Seismic Anisotropy in the Asthenosphere

To study the development of LPO in the sub-lithospheric mantle, we first consider the case of a simple shear flow (Fig. 1A). This is the flow pattern that we expect throughout most of the asthenosphere, and accommodates the differential motion between the lithospheric plates and the sub-asthenospheric upper mantle. Jung and Karato [2001] have shown that olivine aggregates may develop different anisotropic fabrics in response to shear flow. The A-type fabric is found in the majority of naturally occurring rocks [Ismail and Mainprice, 1998]. However, other fabric types may occur in water-rich and/or high stress environments, and produce anisotropy that is not parallel to the maximum shear direction [e.g., Jung and Karato, 2001; Katayama et al., 2004]. Although these more unusual fabrics may be important in certain regions such as the mantle wedge [e.g., Kneller et al., 2005], the A-type fabric is expected to dominate for most regions of the asthenosphere [e.g., Ismail and Mainprice, 1998]. For A-type fabric, Zhang and Karato [1995] showed that the fast axis of olivine initially aligns with the axis of principal extension when exposed to simple shear. For small strains the LPO direction will be oriented 45° to the direction of maximum shear (Fig 1A [This is statement is really from ZK95 not PL02]). However, for larger strains (>~100%), the LPO will rotate into the maximum shear direction [Zhang and Karato, 1995]. Therefore for asthenospheric simple shear (horizontal shear plane), the horizontal projection of LPO will be identical at both small and large strain, and this approximation for LPO has been used by several authors [e.g., Gaboret et al., 2003; Behn et al., 2004] to predict observed anisotropy. It is important to remember that these studies implicitly assume that asthenospheric anisotropy forms quickly and does not depend on deformation history. However, shear flow in the asthenosphere may be more complex due to variations in lithospheric thickness, mantle density structure, and variations in plate motion with time and near plate boundaries. In this case, finite strain accumulation must be considered, as it has been by several recent studies [e.g., Becker et al., 2003; 2006a; 2006b].

To help evaluate under what conditions it is necessary to integrate LPO along flow lines, Kaminski and Ribe [2002] introduced the Infinite Strain Axis (ISA). The ISA is defined as the direction of the LPO after exposure to an infinite amount of deformation within a specific velocity gradient field. If the shear flow does not change with time, then the LPO direction will rotate toward the ISA. The rate at which this rotation occurs, which we define as $\Omega_{\text{ISA}}$, is approximated by the absolute value of the largest eigenvalue of the strain-rate tensor. $\Omega_{\text{ISA}}$ is the reciprocal of the timescale $\tau_{\text{ISA}}$ defined by Kaminski and Ribe [2002]. For simple shear (Fig. 1A), the ISA is parallel to the direction of flow (i.e., the direction of maximum shear). Thus, studies that use observations of anisotropy to infer a direction for mantle flow [e.g., Russo and Silver, 1994; Wolfe and Solomon, 1998; Peyton et al., 2001; Walker et al., 2001; Silver and Holt, 2002; Xue and Allen, 2005] are implicitly assuming infinite deformation under simple shear conditions. Note that the horizontal components of the initial LPO and the final ISA are parallel because finite strain effects only the orientation of LPO in the vertical plane (Fig. 1A). This is probably why studies that ignore finite strain deformation [e.g., Gaboret et al., 2003; Behn et al., 2004] obtain good fits to observed oceanic anisotropy.

A more complicated situation arises when the flow pattern changes with time, either due to spatial variations in the deformation field that olivine fabric experiences as it travels along streamlines, or because the flow is time-dependent. Spatial variations in the deformation field occur if mantle density heterogeneity imparts a vertical component to
asthenospheric flow (Fig. 1B), as we would expect above an upwelling or downwelling mantle. Time-dependent flow occurs if either surface plate motions or the mantle density heterogeneity field evolve with time. Because temporal changes in plate motions are poorly constrained in some areas and the mantle's past density heterogeneity field is poorly constrained everywhere, the integration of finite strain for time periods longer than a few million years [e.g., Becker et al., 2003; 2006a] may introduce significant uncertainty into calculations of LPO development.

If asthenospheric deformation varies either spatially or temporally, the ISA will change orientation as crystals pass through the flow field. Kaminski and Ribe [2002] showed that the rate of ISA rotation along flow lines, which we define here as $\Omega_{\text{flow}}$ (the reciprocal of Kaminski and Ribe’s [2002] timescale $\tau_{\text{flow}}$), is given by the rate of change of the angle, $\Theta$, between the local flow direction and the local ISA. If the local ISA direction changes along flow lines faster than it can be formed, then the net LPO will depend on the finite strain history in a complex way. However, if the rate of ISA formation ($\Omega_{\text{ISA}}$) is faster than the rate of ISA rotation ($\Omega_{\text{flow}}$), then the finite strain history is irrelevant, and the LPO will approach the ISA. To determine which rate is faster, Kaminski and Ribe [2002] introduced the “Grain Orientation Lag” parameter:

$$\Pi = \frac{\Omega_{\text{flow}}}{\Omega_{\text{ISA}}}$$  \hspace{1cm} (1)

which compares the rates of ISA rotation along flow lines to the rate of ISA formation due to local deformation. They find that if $\Pi < 0.5$, the ISA is a good approximation for LPO, which greatly simplifies predictions of anisotropy from a given flow field.

### 3. Models for Global Asthenospheric Flow

Kaminski and Ribe [2002] calculated $\Pi$ for several idealized flow geometries, however to date no studies have evaluated $\Pi$ globally throughout the upper mantle. In this study, we construct a series of models of global mantle flow, driven by various combinations of mantle density heterogeneity and surface plate motions, to predict LPO within the asthenospheric layer. We use the ISA as an approximation for LPO, and then test this approximation by determining if $\Pi < 0.5$.

#### 3.1 Flow Field Calculations

We model global mantle flow using the spherical finite element code CitComS, which has been benchmarked extensively and can handle more than 4 orders of magnitude variation in viscosity [Moresi et al., 1996; Zhong et al., 2000], including the lateral variations important for this study [e.g., Zhong, 2001]. We employ a free slip condition at the core-mantle boundary, but our choice of surface boundary conditions depends on how mantle flow is driven (described below). We employ both layered and laterally-varying viscosity structures for the lithosphere and asthenosphere (described below), which allow us to probe the influence of variable continental thickness on asthenospheric flow and anisotropy.

#### 3.1.1 Finite Element Grid

We set up a global finite element grid with 874800 elements, including 24300 surface elements corresponding to 157 km horizontal resolution at the surface. Vertical resolution is 150 km in the lower mantle, 50 km in the upper mantle, and 25 km above 350 km, which is significantly finer than the horizontal resolution near the surface. We chose to enhance the near-surface vertical resolution at the expense of horizontal
resolution in order to better resolve the horizontal velocity gradients associated with LPO formation in the asthenosphere. As a result, we can accurately measure vertical gradients in both horizontal \((\partial \mathbf{v}_h/\partial r, \partial \mathbf{v}_h/\partial \theta)\) and vertical \((\partial \mathbf{v}_v/\partial r)\) velocity (\(\theta, \psi,\) and \(r\) are the spherical coordinates), but measurement of horizontal gradients in velocity are less accurate. This is not a problem for our calculation because the vertical length scale of the asthenosphere (about 200 km) is much shorter than its horizontal length scale (typically 1000s of km). This causes the asthenospheric strain-rate field to be dominated by its vertical components, which are typically 10-1000 times larger in magnitude than its horizontal components (Fig. 2). Exceptions occur near plate boundaries (Nazca-Pacific) and immediately above upwelling (Africa) or downwelling flow (South America) flow. In these areas, vertical strain-rates are still larger than horizontal strain-rates, but by less than an order of magnitude (Fig. 2). Because our enhanced vertical resolution allows us to resolve these gradients to comparable degrees, we are able use our finite element grid to reproduce the flow fields produced by Behn et al. [2004], who used a semi-analytical spectral method to calculate mantle flow [e.g., Hager and O’Connell, 1981].

3.1.2 Plate-Driven and Density-Driven Flow

Following Behn et al. [2004] we drive instantaneous Stokes flow in the mantle using both mantle density heterogeneity inferred from seismic tomography (density-driven flow) and by imposing surface plate motions (plate-driven flow). For plate-driven flow, we impose NUVEL-1A plate motions [DeMets et al., 1994] for 13 plates in the no-net rotation reference frame as velocity boundary conditions on the surface of the finite element grid. For density-driven flow, we employ rigid surface boundary conditions and assign densities in the mantle by converting velocity anomalies in the S20RTSb seismic tomography model [Ritsema et al., 2004] to densities using a constant velocity-density conversion factor of 0.15 g cm\(^{-3}\) km\(^{-1}\) s. We chose this conversion factor because it is consistent with both laboratory data [e.g., Karato and Karki, 2001] and with previous studies [e.g., Behn et al., 2004]. Also following previous work [e.g., Lithgow-Bertelloni and Silver, 1998; Behn et al., 2004], we do not impose density anomalies above 325 km because near-surface topography will likely be partially controlled by isostatically-compensated density differences [e.g., Jordan, 1975].

The actual flow field will be a combination of plate-driven and density-driven flow. Although it would be preferable to drive plate motions dynamically from density-driven flow, this is impractical for this study. Because asthenospheric shear flow is extremely sensitive to plate motion directions, it is important that our models include the proper plate motions to correctly compare predictions of anisotropy to observations. Behn et al. [2004] provide a more complete justification for imposing plate motions, and present a method for combining the density-driven and plate-driven flow fields that we employ here. They noted that when plate motions are imposed, asthenospheric strain rates are set by the imposed plate velocities. By contrast, stresses are set for the density-driven flow model, making strain-rates proportional to the magnitude of mantle viscosity. Thus, the combined flow field depends on the absolute mantle viscosity assumed for the model. Behn et al. [2004] define a viscosity scale factor \(\beta\) by which to multiply the reference viscosity structure, which they define using lower mantle (below 670 km), asthenospheric (100-300 km) and lithospheric (0-100 km) layers that are 50, 0.1, and 30 times as viscous than the upper mantle (300–670 km), which has a viscosity of \(10^{21}\) Pa s prior to scaling by \(\beta\). By varying \(\beta\), Behn et al. [2004] solved for the viscosity structure that best fits the observed anisotropy field, and found that \(\beta = 0.35\) resulted in the best fit to 13 SKS
shear-wave splitting measurements made at oceanic island stations in the Atlantic and Indian oceans. This value of $\beta$ corresponds to an upper mantle viscosity of $3.5 \times 10^{20}$ Pa s and an asthenospheric viscosity 10 times smaller. Below (see Section 4.2) we employ a similar method to constrain the value of $\beta$ using a global data set, and find a similar best-fitting value ($\beta = 0.4$), which we then use when discussing the combined flow field.

3.1.3 Layered Viscosity vs. Laterally-Varying Viscosity

To investigate the influence of lateral variations in viscosity associated with cooling oceanic lithosphere and deeply-penetrating continental roots, we compare predictions of anisotropy made using a layered (spherically symmetric) viscosity structure with those made using a viscosity structure that includes lateral variations in viscosity. We employ the same layered viscosity profile used by Behn et al. [2004] (described above). In fact, except for the use of an updated tomography model to infer mantle densities (S20RTSb in this study vs. S20RTS in Behn et al. [2004], see Ritsema et al. [2004] for a description of differences), the setup for the calculations performed here using CitComS is identical to the setup employed by Behn et al. [2004] for a spectral code. Away from plate boundaries, which cannot be resolved using a spectral code, our predictions of asthenospheric strain rates closely match those of Behn et al. [2004].

We introduce lateral viscosity variations into the lithospheric and asthenospheric layers (without changing the viscosities of the upper and lower mantle layers) following Conrad and Lithgow-Bertelloni [2006], who assign a characteristic length scale for lithospheric thickness at every surface point on a finite element grid (Fig. 3A). This “characteristic thickness” is used to define the depth-dependence of viscosity through the lithosphere (Fig. 3B). Oceanic lithosphere is assigned a thickness proportional to the square root of its age (taken from Müller et al., [1997]). Characteristic thickness is determined in continental areas following Gung et al. [2003], who employ the maximum depth for which the velocity anomaly (from S20RTSb) is consistently greater than +2%. We imposed 100 km as the minimum continental and maximum oceanic characteristic thickness (Fig. 3A). To introduce a smoothly-varying viscosity structure, we impose an error-function temperature profile above 325 km and use the characteristic lithosphere thickness (Fig. 3A) for the length scale in the error function [Conrad and Gurnis, 2003]. By invoking temperature-dependent viscosity for this boundary layer, we create stiff lithosphere that is thicker for increasing characteristic thickness and is underlain by a low-viscosity asthenosphere down to 300 km (Fig. 3B). To achieve these profiles, we assign temperature-dependent viscosity [e.g., Hirth & Kohlstedt, 1996] using the method of Conrad and Gurnis [2003] and a pre-exponential magnitude that produces an asthenosphere 10 times less viscous than the upper mantle (for direct comparison with the layered case), an activation energy of 200 kJ/mol for the temperature-dependence, and a maximum viscosity 1000 times that of the upper mantle. Note that at depths equivalent to the “characteristic thickness,” the viscosity is smaller than that of the upper mantle (Fig. 3B), which indicates asthenosphere. Thus, the lithosphere's effective (mechanical) thickness is typically smaller than the characteristic thickness defined here.

We also vary the asthenospheric viscosity by adjusting the pre-exponential viscosity magnitude to produce a range of asthenospheric viscosities that are 0.03 (low-viscosity), 0.1 (“reference”), and 1.0 (no asthenosphere) times those of the upper mantle [see Conrad and Lithgow-Bertelloni, 2006]. We adjust the activation energy (using values of 300, 200, and 100 kJ/mol, respectively) so that lithospheric viscosities remain similar between these models despite changes in the pre-exponential factor.
3.2 Predictions of the Lattice Preferred Orientation (LPO)

Kaminski and Ribe [2002] showed that the ISA is a good approximation for LPO if the Grain Orientation Lag parameter $\Pi$ (defined in Eq. 1) is less than 0.5. Thus, to characterize LPO, we calculate both ISA and $\Pi$ for the flow fields described above. To determine ISA, we follow the method described by Kaminski and Ribe [2002], which requires knowledge of the velocity gradient tensor $L_{ij} = \frac{\partial v_i}{\partial x_j}$, which is difficult to accurately measure directly from the flow field. In the asthenosphere, the horizontal strain-rates are significantly smaller than the vertical strain-rates (as in Fig. 2). This means that the elements in third column of $L_{ij}$ are much larger than the other two. We use this fact to form $L_{ij}$ from the vertical components of the strain-rate tensor: $L_{ij} = 0$ except for $L_{rr} = 2\epsilon_{rr}$, $L_{rr} = 2\epsilon_{rr}$, and $L_{rr} = \epsilon_{rr}$ [e.g., Malvern, 1969, chap. 4]. This approximation treats flow in the asthenosphere as horizontal shear flow (depicted in Fig. 1A and governed by the $L_{rr}$ and $L_{rr}$ terms) perturbed by any vertically-oriented component of flow (as shown in Fig. 1B and given by the $L_{rr}$ term). By following the method described in Kaminski and Ribe [2002, Appendix A; note that equation A3 should be $U = FF^T$ as defined by Malvern, 1969], we are able to calculate ISA for asthenospheric flow.

We next evaluate whether the ISA orientations are a valid approximation for LPO by calculating $\Pi$ throughout the mantle flow fields described above. This involves calculating $\Omega_{ISA}$ and $\Omega_{flow}$. The former is defined as the eigenvalue of the strain rate tensor with the largest absolute value. Kaminski and Ribe [2002] define the latter as:

$$\Omega_{flow} = \frac{\text{D}\Theta}{\text{D}t} = \frac{\partial \Theta}{\partial t} + u \cdot \nabla \Theta$$

where $\Theta$ is the angle between the local flow direction ($u$) and the ISA. The second term on the right hand side of (2) represents changes in $\Theta$ as grains are advected along flow lines, and can be determined from the velocity field and the ISA direction. The first term represents changes in $\Theta$ due to the time-dependence of the flow field. For density-driven flow, we measure this quantity by differencing measurements of $\Theta$ across 10 time steps (corresponding to a time $\Delta t$ of ~5 Ma for most runs), and dividing by $\Delta t$. The resulting quantity is typically comparable in magnitude to the advective term. By contrast, plate motions are thought to be steady or only gradually changing, except during times of plate reorganization [e.g., Engebretson et al., 1984], which is not thought to be occurring presently [e.g., Sella et al., 2002]. Thus, time-dependent changes in $\Theta$ for plate-driven flow are likely to be small, and we have not included them here. We sum the advective and time-dependent changes in $\Theta$ to calculate $\Omega_{flow}$ throughout the mantle and divide by $\Omega_{ISA}$ (see Eq. 1) to estimate $\Pi$ for the present-day mantle.

3.3 The Grain Orientation Lag Parameter ($\Pi$)

To investigate the distribution of $\Pi$ throughout the upper mantle, we examine a sample cross section AB (which runs from the East Pacific Rise to the Mid-Atlantic Ridge, see track in Fig. 3A) for both plate-driven flow (Figs. 4A and B) and density-driven flow (Fig. 4C and D), and several different cross sections for the these two flows combined (using $\beta=0.4$, Fig. 5). For each of these cross sections, and for the global planform shown at 225 km depth, we find that $\Pi$ is less than 0.5 (corresponding to a log value of -0.3 shown in the center of the white band in Figs. 4 and 5) throughout most of the asthenosphere and upper mantle. Some regions of the asthenosphere show regions of $\Pi>0.5$. These are typically associated upwelling flow caused by divergence near ridges (Figs. 4A, 4B, and 5B at 5° and 85°, Fig. 5C at 75°, Fig. 5E at 25°) or downwelling flow associated with plate convergence (Figs. 4A, 4B, and 5B at 40°). Downwelling flow...
caused by slab descent (the Nazca slab in Figs. 4C, 4D and 5B at 35°-65°, the Farallon slab in Fig. 5C at 30°-50°, the Tethyan slab in Fig. 5E at 60°-75°) or upwelling flow associated with slow seismic velocity anomaly (the African “superplume” in Fig. 5D at 25°-75°) may also lead to $\Pi > 0.5$ in parts of the asthenosphere. However, away from plate boundaries, $\Pi < 0.5$ for the large majority of the asthenosphere (Fig. 5A).

The lower mantle (>660 km) shows large values of $\Pi$ because its high viscosities result in small strain rates, leading to small $\Omega_{ISA}$ and large values of $\Pi$. By contrast, $\Omega_{flow}$ does not differ significantly between the upper and lower mantle, which indicates that the spatial and temporal gradients that characterize lower mantle flow are similar to those of the upper mantle. Because rates of ISA rotation in the lower mantle are similar to those in the upper mantle, the slower development of the ISA in the lower mantle (indicated by small $\Omega_{ISA}$) may prevent the development of stable LPO there. This may be a possible explanation for the observed lack of a coherent anisotropic fabric in the lower mantle [e.g., Meade et al., 1995], despite recent laboratory observations indicating that lower mantle perovskite deforms according to dislocation creep, and thus should yield an anisotropic fabric [Cordier et al., 2004]. Thus, large $\Pi$ in the lower mantle may be partly responsible for its isotropic nature.

Values of $\Pi$ within the high-viscosity lithosphere are also typically several orders of magnitude larger than 0.5 (Figs. 4 and 5). In fact, most of the lithosphere can be distinguished from the asthenosphere by a large increase in $\Pi$. These large $\Pi$ values result from slow anisotropic fabric development associated with small lithospheric strain-rates. For these regions, prediction of anisotropic direction from a viscous flow model is difficult, if not impossible, because it requires an accurate integration of the strain history during the past 10-100 Myr or more. For this reason, we ignore anisotropy predictions in the lithosphere where $\Pi > 0.5$. This condition is generally not met for lithosphere near plate boundaries, where a discontinuity in plate motions leads to large strain rates and $\Pi < 0.5$ at ridges (mid-Atlantic ridge, Figs. 5B and 5C), subduction zones (Nazca slab, Fig. 5B), transform faults (San Andreas, Fig. 5C), and continent-continent collisions (India-Asia, Fig. 5E). Although these large strain rates generate small $\Pi$ within the lithosphere at plate boundaries, this study does not have the numerical resolution (157 km horizontal grid spacing) to resolve the detailed deformation occurring in these regions. More detailed studies of corner flow [Kaminski and Ribe, 2002] show that $\Omega_{flow}$ increases at plate boundaries, making $\Pi > 0.5$ likely in higher resolution studies. In fact, Behn et al. [2004] found that stations located <500 km from a mid-ocean ridge axis are best fit by the direction of spreading, which indicates that corner flow associated with spreading dominates the observed anisotropy for near-ridge stations. Stations near subduction zones will also likely be similarly affected by local flow associated with the plate boundary, which would suggest increased $\Pi$. Similarly, convective instability, or any other local flow disturbance with length scales comparable to our grid resolution, should increase $\Omega_{flow}$ and thus would lead to larger $\Pi$ than we have predicted here (Fig. 5).

Despite a few regions in which $\Pi > 0.5$, high strain-rates in the asthenosphere lead to $\Pi < 0.5$ for most of this layer (Figs. 4 and 5). In some locations, such as within a band extending ~150 km below the high-viscosity lithosphere, $\Pi$ is typically orders of magnitude smaller than 0.5. This is particularly true for density-driven flow, where $\Pi < 0.5$ for more than 90% of points located between 20% and 60% of the distance from the asthenospheric roof to its base (Fig. 6A, gray triangles). This region of particularly small $\Pi$ occurs because the vertical component of density-driven flow in the upper mantle must spread or converge laterally when it interacts with the rigid lithosphere. The
high strain rates associated with the resulting shear flow lead to diminished $\Pi$ in the upper asthenosphere. By contrast, plate-driven flow induces a shear flow that is more evenly distributed throughout the asthenosphere, leading to more uniform values of $\Pi$ through the layer (Fig. 6A, white circles). The combined flow (Fig. 6A, black squares) shows $\Pi<0.5$ for more than ~70-80% of asthenospheric points. The variation of $\Pi$ within the asthenosphere is not significantly dependent on asthenospheric viscosity (Fig. 6B), although lack of a low-viscosity asthenosphere decreases the number of points for which $\Pi<0.5$ to about 60-70% throughout the layer (Fig. 6B). Without an asthenosphere, shear flow from both density-driven and plate-driven flows occurs over a broader depth range and with smaller strain-rates. Both effects act to increase $\Pi$ in the asthenosphere.

3.4 The Infinite Strain Axis (ISA)

Because $\Pi<0.5$ throughout most of the asthenosphere, it is appropriate to use the ISA as an approximation for LPO within this layer [Kaminski and Ribe, 2002]. As discussed above, the ISA is most sensitive to shear deformation, and orients in the direction of maximum shear. For plate-driven flow, asthenospheric shear parallels plate motions, as does the ISA (Figs. 4A and 4B, where flow is dominated by convergent Nazca and South American plate motions). For density-driven flow, horizontal shear accommodates motion between the stationary lithosphere and the flowing mantle. As discussed above, this shear occurs in the upper asthenosphere, where the ISA direction is parallel to the direction of uppermost mantle flow (e.g., Fig. 4C and 4D, where flow is dominated by the downgoing Nazca slab at about 50°). For combined flows (Fig. 5), the ISA direction is generally parallel to the sublithospheric base and aligned in the direction of flow, except for a few regions near plate boundaries and upwellings where flow is nearly vertical. The fact that horizontal shear nearly parallels the ISA explains why studies that use the former as an approximation for LPO [e.g., Gaboret et al., 2003; Behn et al., 2004] do an excellent job of predicting oceanic anisotropy.

We evaluate the influence of lateral variations in lithospheric thickness on the ISA by comparing the layered and laterally-varying viscosity structures (Fig. 7). The predicted anisotropy direction changes significantly beneath the thin lithosphere of oceanic ridges, near some subduction zones, and on the edges of deeply-penetrating continental roots (eastern North America, Africa, western Australia, Fig. 7). These changes occur for two reasons. First, convergent or divergent flow at plate boundaries is sensitive to the viscosity structure because it includes a significant vertical component of flow. Second, increasing or decreasing the lithospheric thickness will increase or decrease the depth at which asthenospheric shear flow occurs. This may cause a given shear flow to move below or above the depth at which anisotropy is measured, thus changing the observed direction. On the other hand, only small changes in the ISA direction are observed beneath oceanic plates away from plate boundaries (Fig. 7), where the laterally-varying viscosity structure resembles the layered case (Fig. 3B). This explains the good fit in oceanic regions between observed anisotropy and LPO predictions made using flow models that employ layered viscosity structures [Becker et al., 2003; Gaboret et al., 2003; Behn et al., 2004].

4. Comparison to Global SKS Splitting Data

We compare the above-described predictions of asthenospheric anisotropy to a global dataset of shear-wave splitting measurements. In doing so, we constrain the
absolute mantle viscosity appropriate for combining plate-driven and density-driven flows (\(\beta\) parameter) and the depth range of the asthenosphere that controls anisotropy. We also evaluate the importance of lateral viscosity variations and differences between continental and oceanic anisotropy.

4.1 Global SKS Dataset

The global shear-wave splitting dataset used in this study was derived from the Arizona State University Global Upper Mantle Anisotropy Dataset (http://geophysics.asu.edu/anisotropy/upper). This dataset contains ~1350 splitting measurements (Fig. 8) using SKS and SKKS phases compiled from 87 studies in oceanic and continental environments. In addition, we have supplemented this dataset with several recent shear-wave splitting studies in the ocean basins [Wolfe and Silver, 1998; Smith et al., 2001; Klosko et al., 2001; Fontaine et al., 2005; Hammond et al., 2005].

4.2 Inversion for the Best-Fitting Flow Model

We use the shear-wave splitting data at oceanic stations to constrain the viscosity scale factor \(\beta\), which sets the mantle viscosity and is necessary for combining the plate-driven and density-driven flows into a single flow field (section 3.1.2). Using a smaller dataset of 13 oceanic stations that surround Africa, Behn et al. [2004] found a best-fitting value of \(\beta=0.35\). Here we repeat this analysis using a larger dataset of 106 oceanic stations extracted from our global dataset. We exclude continental stations because continental lithosphere may preserve an anisotropic fabric of its own associated with past or ongoing lithosphere deformation [e.g., Silver, 1996]. Thus, using continental stations to constrain \(\beta\) may result in the flow field being fit to lithospheric anisotropy unrelated to current mantle flow. Oceanic lithosphere, on the other hand, is typically thinner and has a shorter, simpler, and less active deformation history. Thus, lithospheric anisotropy in the oceans should be less significant for oceanic stations, and single layer anisotropy models (asthenosphere only) typically fit SKS splitting data at ocean island stations better than two layer models (asthenosphere and lithosphere) [Behn et al., 2004].

We also do not wish to constrain the viscosity structure using stations where we do not expect our models to accurately predict anisotropy. Thus, we exclude stations for which \(\Pi>0.5\) in the flow model (at the depth being compared to observations) because our use of the ISA to approximate LPO may not be appropriate at these stations. Furthermore, because we expect \(\Pi>0.5\) near plate boundaries (see Section 3.3), we exclude oceanic stations located < 500 km from a plate boundary from our analysis. For the inversions for \(\beta\) below, we are left with 19 “clean” oceanic stations (stations with circles in Fig. 8) for which we know of no corrupting influences that would prevent us from comparing predictions of anisotropy to observations. These 19 stations include only 8 of the 13 stations that Behn et al. [2004] used to constrain \(\beta\) (five of their stations did not satisfy our \(\Pi<0.5\) criterion).

We first examine the variation of predicted anisotropy (ISA direction) as a function of depth within the asthenosphere, and compare it to the SKS splitting measurements made at these 19 oceanic stations. To do this, we measure the average misfit, which we define as the average angular difference between the predicted ISA direction and the SKS fast polarization direction, as a function of depth (Figs. 9A and 9B). We exclude points for which \(\Pi>0.5\), which effectively eliminates nearly all points in the lithosphere (Fig. 6). For values of \(\beta\) that bracket the \(\beta=0.35\) value obtained by Behn et al. [2004], we observe a region in the mid-asthenosphere (between about 150 and 275
km depth) where both layered (Fig. 9A) and laterally-varying (Fig. 9B) models show a significantly improved fit to the oceanic SKS splitting observations compared to the expected result for a random distribution (45°). For laterally-varying viscosity, this region is narrower (~100 km thick) and produces slightly smaller misfits, but is located in about the same region of the asthenosphere (centered at about 225 km depth) that Behn et al. [2004] used to compare predicted and observed anisotropy directions (200 km). If we examine the full range of $\beta$ between density-driven flow only (small $\beta$) and plate-driven flow only (large $\beta$), we find a best-fitting value of $\beta=0.4$ at 225 km (Fig. 9C) for layered viscosity and $\beta=0.4$ at 225 for laterally-varying viscosity (Fig. 9D). The average misfit for both layered and laterally-varying viscosity cases is 12°, which is comparable to what found by Behn et al. [2004] for a smaller SKS dataset (13°). Because its average misfit is biased by one poorly fitting station (53°), the variable viscosity model shows a distribution of misfits that is more tightly clustered toward lower values (Fig. 10A) and produces a smaller median misfit (7°) than the layered case (11°).

SKS-related phases follow a nearly vertical path through the upper mantle, and thus represent a measure of the cumulative anisotropy in the asthenosphere. For small angular variations, these splitting measurements are approximately sensitive to the average orientation of anisotropy throughout the asthenospheric layer. For larger angular variations with depth, the net anisotropy is a more complex function. If we calculate the misfit for the average ISA direction for all asthenospheric depths where $\Pi<0.5$, we find best-fitting $\beta$ values to oceanic splitting data of 0.7 and 1.0 for the layered viscosity (Fig. 9C) and laterally-varying (Fig. 9D) viscosity structures, respectively. These $\beta$ values are slightly larger than we found using a single, best-fitting depth to infer anisotropy, which implies larger asthenospheric viscosities and a greater role for plate-driven flow in determining anisotropy. As discussed above, plate-driven flow tends to produce a thicker region of coherent anisotropy than does density-driven flow (compare Figs. 4A and 4B to Figs. 4C and 4D, also note thickness of low $\Pi$ values in Fig. 6A). Thus, larger values of $\beta$ tend to generate a thicker region within the asthenosphere that fits the observed anisotropy well (compare $\beta=0.2$ to $\beta=0.6$ curves in Figs. 9A and 9B), and are preferred if we use the entire asthenospheric thickness to infer anisotropy. However, we obtain slightly better overall fits if we consider only the best-fitting depth (Fig. 9D), in which case a more concentrated layer of well-fitting anisotropy is preferred.

In general, values of $\beta$ between about 0.3 and 1.0 yield predictions of asthenospheric anisotropy that fit oceanic SKS observations extremely well. For this study, the model that gives the best overall fit to the oceanic component of our splitting dataset is one that employs lateral viscosity variations and uses a value of $\beta=0.4$ to define the relative importance of density-driven and plate-driven flows (Fig. 9D). This “preferred” model fits the oceanic anisotropy best at 225 km (Fig. 9B), where $\Pi<0.5$ for most (~80%, Fig. 5A) locations, and more than 60% of all oceanic stations are fit to within 10°, producing an average misfit of 12°. We also note that the fit of the preferred model is remarkably good and approaches the nominal uncertainty of the splitting observations [Behn et al., 2004]. For the oceanic environment, our successful prediction of sub-lithospheric anisotropy validates the use of ISA to predict LPO, the dominance of A-type fabric [Jung and Karato, 2001] in the asthenosphere, and the flow model itself.

4.3 Oceanic vs. Continental Anisotropy

Our “preferred” model of asthenospheric flow (described above) does a good job of predicting oceanic anisotropy, with about 80% of oceanic stations matching the ISA
direction at 225 km (Fig. 8) to within 20° (Fig. 10A). This model also makes predictions of asthenospheric anisotropy beneath continental lithosphere. For continental stations with Π<0.5 at 225 km, however, our preferred flow model produces an average fit to splitting measurements (Fig. 8) that is significantly worse than it is for the oceanic stations, (39° for continents vs. 12° for oceans). When the misfit is viewed as a function of lithospheric thickness (Fig. 11), it is clear that asthenospheric flow does a much better job of predicting anisotropy in oceanic areas (lithosphere 100 km thick or less) than it does for continents (100 km thick or more). These results suggest that either lithospheric anisotropy plays a much more important role for continents than it does for oceans, or that the “preferred model” does not accurately predict asthenospheric flow beneath continents. Nevertheless, of the 895 continental splitting measurements, 34% of stations are fit to within 25° and 57% of stations are fit better than 45° (Fig. 10B). Because there are so many continental stations, we can be more than 99% confident that the distribution of continental misfits (Fig. 10B) does not occur randomly [Press et al., 1992, pp. 609-615]. (Although there are nearly 50 times fewer oceanic stations, the small misfits for these stations allows us to be equally confident that the oceanic distribution is not random.) Thus, our “preferred” model slightly improves the fit for continental stations compared to a random distribution, both overall and for lithosphere of all thicknesses (Fig. 11). This hints that asthenospheric anisotropy may contribute regionally to the net anisotropy in some continental areas.

The average fit for the layered viscosity case is comparable to that of the laterally-varying case (Fig. 10B), although regionally the average misfit may be significantly improved (North and South America, Figs. 10C and 10D) or degraded (Eurasia, Fig 10E) by the introduction of laterally-varying viscosity. This shows that asthenospheric anistropy is expected to be sensitive to the viscosity structure of the continental lithosphere, and leaves open the possibility that refinement of the asthenospheric flow model beneath continents could improve the prediction of continental splitting observations. On the other hand, the fact that lateral viscosity variations do not significantly improve the global average fit to continental anisotropy may suggest that asthenospheric anisotropy does not contribute significantly to observations of continental anisotropy. Instead, splitting observations made on continents may be dominated by fossil lithospheric fabric, which cannot be predicted by mantle flow models.

5. Discussion

The observation that Π<0.5 throughout most of the asthenosphere implies that the ISA can be used as an approximation for LPO in most locations [Kaminski and Ribe, 2002]. This greatly simplifies the prediction of asthenospheric anisotropy because the full determination of LPO is computationally expensive and complicated [Kaminski et al., 2004], and requires an accurate method to trace finite strain accumulation along flow lines [e.g., Becker et al., 2003]. Such a method cannot be implemented from “instantaneous” flow models that predict the present-day flow field because such models do not take into account the time-dependence of mantle flow. Furthermore, uncertainty in the rates of grain boundary migration and dynamic recrystallization may introduce uncertainty into the calculation of LPO [Kaminski and Ribe, 2001]. By contrast, the ISA can be determined from an instantaneous “snapshot” of the mantle flow field without consideration for the time-dependent accumulation of finite strain. The few areas of the asthenosphere where long-wavelength upwellings (African Superplume, Fig. 5D) or
downwellings (Farallon slab, Fig. 5C) may cause development of LPO away from the ISA can easily be screened using a Π<0.5 cutoff criterion. This criterion is also useful for differentiating regions with significant anisotropy-producing shear flow (low Π) from the slowly-deforming lithospheric regions (large Π) where anisotropy may be associated with past deformation (Figs. 4 and 5).

Short-wavelength complexities to the flow field are more problematic because our global flow model (with 157 km horizontal resolution) cannot resolve details of flow associated with plate boundaries, localized convective downwelling, or upwelling plumes. Thus, we may predict Π<0.5 even though more detailed models might predict Π>0.5, which would indicate that the ISA may not be parallel to the LPO. Because this is likely true at certain plate boundaries, we have ignored stations near plate boundaries when constraining our model using observed splitting data. It is not known, however, which parts of the lithosphere might be currently experiencing localized convective downwelling because such downwellings likely remove only the lower part of the lithospheric layer and are difficult to detect from the surface [Conrad and Molnar, 1997]. If downwelling instability is present beneath a station, however, flow from this downwelling will influence LPO in an unpredictable way, and could generate LPO that is unrelated to background mantle flow. However, localized convective downwelling is likely to be short-lived and infrequent in most environments [e.g., Conrad and Molnar, 1999], and thus unlikely to affect large numbers of shear-wave splitting measurements. The fact that we fit the oceanic shear-wave splitting data to ~12° indicates that anisotropy beneath most oceanic stations is not strongly influenced by localized convective downwelling. Similarly, although localized upwelling from mantle plumes may influence many of the oceanic island stations used for our inversion, our successful prediction of oceanic anisotropy suggests that these local upwellings do not strongly perturb shear-wave splitting measurements at these oceanic stations. Furthermore, Kaminski and Ribe [2002] showed that the ISA direction does not differ significantly from the background shear direction directly above an upwelling plume. Thus, although we have ignored convective instability and upwelling plumes in our analysis of oceanic anisotropy, their influence is probably not significant for oceanic stations.

Global mantle flow models do a good job of explaining observed shear-wave splitting measurements made in oceanic regions. The asthenospheric flow predicted by these models is characterized by shear flow that accommodates the relative motion between surface plate motions and subasthenospheric upper mantle flow. Using globally-distributed oceanic shear-wave splitting measurements as a constraint, we have confirmed the result of Behn et al. [2004], who showed both plate-driven and mantle-driven flows are necessary to explain splitting observations at oceanic stations surrounding Africa. We find that an asthenospheric viscosity of ~4x10^19 Pa s generates rates of asthenospheric flow driven by mantle density heterogeneity that, when combined with shear flow associated with plate motions, best explains oceanic anisotropy. This result is nearly identical to that obtained by Behn et al. [2004], despite the fact that Behn et al. [2004] used a spectral code that could not handle lateral viscosity variations and used the direction of maximum shear, rather than the ISA, to predict anisotropy. Most of the stations that we have added to the Behn et al. [2004] analysis, however, are on the fast-moving Pacific or Australian plates, where plate-driven flow dominates observed anisotropy. Thus, additional shear-wave splitting measurements made on slowly-moving plates in the Atlantic or Indian basins will be crucial to further constraining the relative importance of plate-driven and density-driven asthenospheric flows. Oceanic anisotropy
inferred from surface wave models [e.g., Becker et al., 2003; Debayle et al., 2005] should also provide useful constraints because coverage is more globally uniform (although less horizontally-resolved) than shear-wave splitting observations.

The fact that our models for asthenospheric flow predict oceanic anisotropy significantly better than continental anisotropy can be explained in several ways. First, while our flow models seem to accurately predict flow beneath oceanic lithosphere, they may do so more poorly beneath continents. This would be the case if the viscosity or density structures beneath continents are not accurately characterized due to poor constraints on these parameters. Alternatively, it is possible that our shear flow approximation is less appropriate for sub-continental asthenosphere than it is for oceans. The incorporation of horizontal velocity gradients, which are not included in our analysis, may be enhanced by lithospheric thickness variations (Fig. 3) and should be balanced by larger vertical velocity gradients caused by narrowing the asthenospheric channel beneath thicker lithosphere. Future calculations with greater horizontal resolution may be able to address these effects. A poorer fit to the continental splitting data could also result if small-scale convection is more prevalent beneath continents (as we might expect because continents are typically thicker and more deforming than oceans [e.g., Conrad, 2000]), or if sub-continental shear flow produces a different anisotropic fabric than does shear flow beneath oceans (for which we have successfully predicted anisotropy using the A-type fabric). However, the non-A-type fabrics are expected primarily at high water contents [Jung and Karato, 2001], and the presence of water is probably only a factor near plate boundaries and not preferentially beneath continents. Thus, we cannot rule out the possibility that continental anisotropy may be dominated by an asthenospheric component that is not correctly predicted by our current models.

Another, perhaps more likely, explanation for why asthenospheric shear flow successfully predicts anisotropy in oceanic areas but fails to do so for continents is the possibility that lithospheric anisotropy, which cannot be predicted by our models, is more significant for continents than it is for oceans. In fact, lithospheric anisotropy has been invoked to explain anisotropy in several continental regions. First, lithospheric deformation associated with orogenic deformation may induce significant anisotropy into the lithospheric fabric [Silver, 1996]. In fact, our predictions of asthenospheric anisotropy (Fig. 8) poorly fit observed anisotropy in the Alpine, Tibetan, and parts of the Andean orogenic zones (although the predictions of Andean anisotropy are likely also influenced by unmodeled flow associated with subduction). Second, because cratonic lithosphere is old, thick, and cold, lithospheric anisotropy created by any past deformation may be preserved for long periods of geologic time. For example, Silver et al. [2004; 2006] argue that patterns of anisotropy observed in southern Africa (Fig. 8) are lithospheric in origin. Although the generally NE-SW orientation of anisotropy in southern Africa is consistent with an asthenospheric contribution (Figs. 8, 10F), local deviations over short spatial scales argue for a strong lithospheric component in these data [Fouch et al., 2004]. Furthermore, other cratonic areas (Canadian, Siberian) are poorly fit by asthenospheric shear flow (Fig. 8). Thus, the presence of lithospheric anisotropy in many continental areas may explain why our predictions of asthenospheric anisotropy fit the continental splitting data more poorly than they do the oceanic data.

Despite the likely presence of an anisotropic continental lithosphere, the fit to continental anisotropy is improved in some areas by the introduction of lateral variations in lithospheric viscosity (Fig. 10), which suggests an asthenospheric component to some continental anisotropy. For example, the introduction of lateral viscosity variations
improves the average fit for North American splitting observations (Fig. 10C), particularly in areas away from cratons (eastern and western North America, Fig. 8). Lateral viscosity variations also improve the fit for Africa (Fig. 10F) and South America (Fig. 10D), although the station coverage for these continents is poor (Fig. 8) and overlaps areas where we expect significant lithospheric anisotropy (e.g., southern Africa). On the other hand, the introduction of lateral viscosity variations in Eurasia degrades the fit to observed anisotropy (Fig. 10E). This may indicate problems with the Eurasian lithospheric thickness model (Fig. 3A), although many European and Asian data are located in orogenic or cratonic areas where we expect significant lithospheric anisotropy (Fig. 8). Alternatively, the oceanic flow model may not be appropriate for this very large, slowly moving continental area. Clearly a more detailed study of individual continental regions is required to more completely assess regional variations in the relative importance of lithospheric and asthenospheric contributions to continental anisotropy.

We hypothesize that the most likely reason that asthenospheric flow models fail to predict continental anisotropy is that anisotropic fabric in both the lithospheric and asthenospheric layers probably contribute to the net anisotropy measured at continental stations. There are several reasons why we might expect continental lithosphere to be more anisotropic than oceanic lithosphere. First, because continental lithosphere is older and colder than oceanic lithosphere [e.g., Rudnick et al., 1998], it may better preserve tectonic fabric imparted to it due to previous deformation over geologic time [e.g., Savage, 1999]. Continental lithosphere also experiences more deformation than does oceanic lithosphere due to the fact that it is older, but also possibly because it is thicker [e.g., Conrad and Lithgow-Bertelloni, 2006]. The greater thickness of continental lithosphere provides a thicker layer in which lithospheric anisotropy can be preserved. Finally, dislocation creep rheology, which generates anisotropic LPO, is only dominant over diffusion creep down to 250-350 km depth [e.g., Hirth and Kohlstedt, 2003]. Thick continental roots may push the sub-lithospheric deforming region below this depth, thus preventing the development of anisotropic fabric by asthenospheric flow.

The spatial distribution of shear wave splitting observations (Fig. 8) is highly variable. Oceanic coverage is particularly sparse, and continental coverage includes some areas with excellent (North America, Europe, and Southern Africa) and some with very poor (Australia, Amazonia, Antarctica, the rest of Africa) spatial coverage. Thus, a complete characterization of continental anisotropy (lithospheric and asthenospheric components) requires significantly better spatial coverage of shear-wave splitting observations. Surface wave anisotropy models [e.g., Becker et al., 2003; Nettles and Dziewonski, 2004; Debayle et al., 2005] could be especially useful because they provide some depth resolution that could be used to separate the lithospheric and asthenospheric components of the net anisotropy measured by shear-wave splitting. Initial analyses of surface wave studies do suggest differences between continental and oceanic anisotropy with patterns consistent with the model predictions made here. For example, the azimuthal anisotropy model of Debayle et al. [2005] shows a slightly deeper peak in the magnitude of anisotropy (~100 km) for oceans than for continents (~50 km). This suggests a shallow lithospheric source of anisotropy dominates for continents, and a deeper asthenospheric origin for oceans. Similar patterns are seen in regional studies of North America [Nettles and Dziewonski, 2004], where azimuthal anisotropy is large 100-200 km beneath the eastern Pacific Ocean and the Basin and Range province, but from 50-100 km beneath the craton. The presence of significant anisotropy beneath the Basin and Range is consistent with our predictions for western North America (Fig. 8) and with
other anisotropy studies [Silver and Holt, 2002; Becker et al., 2006b], which suggests that asthenospheric flow may contribute to continental anisotropy in some areas.

6. Conclusions

We have shown that viscous deformation in the asthenosphere tends to align the LPO of olivine aggregates into the infinite strain axis (ISA) corresponding to their orientation after infinite deformation. This is because asthenospheric shear flow rotates the LPO of olivine aggregates toward the ISA faster than the complexities of the flow can change the sense of deformation that these olivine aggregates experience. Kaminski and Ribe [2002] expressed this comparison of rates within their definition of the Grain Orientation Lag parameter $\Pi$; here we have shown that $\Pi$ is small throughout the asthenosphere away from plate boundaries. This means that the strain-history of flow need only be considered in a few locations (such as plate boundaries) where the flow field is extremely time-dependent or changes rapidly along flow lines. This is an important simplification because computing strain deformation is expensive and may involve complicated (and often poorly constrained) modeling of the time-dependence of the flow field [e.g., Becker et al., 2003]. This is particularly true if the effects of dynamic recrystallization [Kaminski and Ribe, 2001; Kaminski et al., 2004] are taken into account along with finite strain [Becker et al., 2006a]. By contrast, the ISA and $\Pi$ can be calculated easily from any given flow field and its time derivative.

Using the assumption that the ISA orientation approximates asthenospheric LPO, we have shown that shear-wave splitting observations in oceanic regions are well fit asthenospheric shear flow driven by a combination of plate motions and mantle density heterogeneity inferred from seismic tomography [e.g., Behn et al., 2004]. Furthermore, the fact that $\Pi < 0.5$ throughout most of the asthenosphere implies that the direction of asthenospheric anisotropy beneath the ocean basins is a reasonable approximation for the direction of asthenospheric flow in most locations away from plate boundaries [Kaminski and Ribe, 2002]. By contrast, continental anisotropy is more poorly fit by the ISA predicted by models that fit oceanic anisotropy well. The average fit to continental shear-wave splitting measurements, however, is improved compared to a random distribution, which suggests that sub-lithospheric viscous flow in the asthenosphere may contribute to splitting observations in some continental areas.

Thus, it seems that viscous shear flow in the asthenosphere controls LPO formation and produces an asthenospheric layer of flow-induced anisotropy worldwide. In oceanic regions, which feature a thin lithosphere with little history of tectonic deformation, this asthenospheric anisotropy largely explains shear-wave splitting measurements made at oceanic stations. This anisotropic layer is probably present beneath continental lithosphere as well, and contributes to shear-wave splitting observations at continental stations. However, continental lithosphere contains an anisotropy of its own that is not present in oceanic lithosphere to the same degree. This lithospheric anisotropy contributes, and may in fact dominate, the net anisotropy observed by shear-wave splitting measurements at continental stations. This inference that asthenospheric anisotropy dominates beneath oceans, but lithospheric anisotropy may dominate in continental areas has also been suggested by surface wave tomography studies [e.g., Nettles and Dziewonski, 2004; Debayle et al., 2005].

Continental lithosphere is colder, thicker, and older, and more deformed, than oceanic lithosphere. All of these factors should contribute to the preferential generation
of long-lived anisotropic fabric in continental lithosphere. In addition, increased 
continental thickness may decrease the thickness of the deforming zone beneath the 
lithosphere where asthenospheric anisotropy is formed. Given the dramatic differences 
between oceanic and continental lithosphere, it is not surprising that oceanic anisotropy 
can be largely explained by asthenospheric flow while a fossil lithosphere fabric must be 
invoked to explain continental anisotropy.

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Figure 1. Comparison of (A) simple shear flow, in which the lattice preferred orientation (LPO, gray line) rotates toward the infinite strain axis (ISA, black line) as it is exposed to finite strain, with (B) a more complicated flow in which crystal rotation by the flow occurs faster than ISA development. This causes the LPO to lag behind the ISA.

Figure 2. Ratio of the magnitude of vertical strain rates to the magnitude of horizontal strain rates, averaged throughout the layered asthenosphere (100-300 km). Shown here are strain-rates calculated using the semi-analytical spectral model of Behn et al. [2004], updated for comparison to the finite element models presented here (S20RTSb tomography model and $\beta=0.4$). For most locations, vertical gradients in velocity dominate the strain-rate tensor.
Figure 3. (A) Map of assigned characteristic lithospheric thickness and (B) the depth dependence of viscosity through the lithosphere and asthenosphere for the layered viscosity structure (black line) and for the laterally-varying viscosity structure for different assigned characteristic lithospheric thicknesses (colored lines). Tracks in (A) show the great circle paths of cross sections shown in Figures 4 and 5.
Figure 4. Cross sections showing $\Pi$ (colors) and the ISA (bars) through the upper mantle along cross section AB, which traverses the Nazca and South American plates (Fig. 3A). Shown are results for flow driven by imposed plate motions (A and B) and for flow driven by mantle density heterogeneity (C and D). Both a layered viscosity structure (A and C) and a laterally-varying viscosity structure (B and D) are imposed for the lithosphere and asthenosphere. The upper and lower boundaries of the low-viscosity asthenosphere, defined as the depths where viscosity crosses $10^{21}$ Pa s (Fig. 2B) are highlighted by black lines. The vertical length axis is exaggerated by a factor of 9 so that vertical variations in $\Pi$ and ISA can be viewed more easily. To maintain the ISA direction relative to the flow field, the vertical component of the ISA is exaggerated by the same factor. The magnitude of the ISA, however, is unity, so variations in the length of ISA bars represent changes in the component of ISA that is perpendicular to the cross section.
Figure 5. Planform at the best-fitting depth of 225 km (A) and cross sections (B-D) showing Π (colors) and ISA (bars) for a “preferred” flow model that combines platedriven and density-driven flow (using $\beta = 0.4$), as described in the text. The surface paths (AB, CD, EF, and GH) of the cross sections (parts B-D) are shown in (A) and in Fig. 3A. The asthenosphere is bounded above and below by black lines as in Fig. 4.
Figure 6. The fraction of points globally for which $\Pi < 0.5$, as a function of depth within the lithosphere and asthenosphere. Here depth is given as a fractional component of the thickness of each layer because the layer thicknesses vary laterally (Fig. 3). The 0.5 cutoff is the value of $\Pi$ below which the ISA approximates LPO given the effects of finite strain accumulation [Kaminski and Ribe, 2002]. Shown in (A) are tabulations for the case of laterally-varying viscosity and plate-driven flow (white circles, Fig. 4B), density-driven flow (gray triangles, Fig. 4D), and the combination of the two flows (black squares, Fig. 5). Shown in (B) are tabulations for combined plate- and density-driven flows ($\beta = 0.4$) assuming the asthenospheric viscosity of the base model of Fig. 3 (black squares), or using asthenospheric viscosities that are 10 (which imparts no asthenospheric viscosity drop, gray circles) or 0.3 (low viscosity case, white inverted triangles) times this value.
Figure 7. Comparison of the ISA direction at 225 km for laterally-varying (Figure 5A) and layered viscosity structures. Plate-driven and density-driven flows are combined using $\beta = 0.4$.

Figure 8. Comparison of the ISA direction for the “preferred” model (combined density-driven and plate-driven flows ($\beta=0.4$) and lateral viscosity variations) at 225 km (black bars, as shown in Fig. 5A) with the global compilation of SKS splitting directions. The directional misfit (in degrees) for each station is given by its color. Oceanic stations used in the global misfit calculation are shown with a central circle over the station location. Stations colored gray are not used in the analysis because $\Pi>0.5$ at 225 km, which indicates that our approximation of the ISA direction for the LPO may not be appropriate at these stations.
Figure 9. The average misfit for oceanic stations. Shown in both A (layered viscosity) and B (laterally-varying viscosity) is the variation of misfit with depth for different values of the viscosity scale factor $\beta$ (as defined by Behn et al. [2004], see text). Shown as a function of $\beta$ in both C (layered viscosity) and D (laterally-varying viscosity) is the variation of misfit at 225 km (black triangles), which is the depth that provides the smallest misfit in parts A and B, and the average misfit through the asthenospheric layer (white circles).
Figure 10. Histogram showing the misfit (in 10° depth bins) for (A) oceanic, (B) continental, (C) North American, (D) South American, (E) Eurasian, and (F) African stations between SKS observations and predictions of ISA at 225 km for the layered (dashed boundaries) and laterally-varying (gray, solid boundaries) viscosity structures. We combine plate-driven and density-driven flow using a value of $\beta=0.4$ consistent with the “preferred” model in both cases.
Figure 11. Station misfit (Fig. 8) as a function of characteristic lithospheric thickness (Fig. 3). Oceanic stations used to constrain the viscosity scale factor $\beta$ are shown with stars. Gray bars show the average misfit within 50 km depth intervals. The large number of points at 100 km thickness corresponds to our assignment of thin continental or thick oceanic lithosphere to this characteristic thickness. The 10 km depth interval used to infer lithosphere thickness from tomography [e.g., Conrad and Lithgow-Bertelloni, 2006] is evident for large lithosphere thickness, which is present beneath many stations in southern Africa (especially at 210 km) and North America (Fig. 8).