Modeling anisotropy and plate-driven flow in the Tonga subduction zone back arc

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Abstract. The goal of this study is to determine whether shear wave splitting observed in subduction zone back arc regions, the Tonga subduction zone in particular, can be quantitatively modeled with flow in the back arc mantle driven by the motions of the subducting slab and the upper back arc plate. We calculated two-dimensional mantle flow models using known Tonga plate motions as boundary conditions and assuming a range of uniform and variable viscosity structures. Shear wave splitting was predicted for the anisotropy due to lattice preferred orientation (LPO) of olivine and orthopyroxene in the flow model finite strain fields. The predicted shear wave splitting provides a good match to the fast directions (parallel to the azimuth of subducting plate motion) and splitting times (0.5-1.5 s) observed in Tonga, both for models where LPO anisotropy develops everywhere above 410 km and for models where LPO anisotropy is confined to regions of relatively high stress. If LPO anisotropy does develop over the entire upper 410 km of the mantle, the strength of anisotropy induced by a given amount of shear strain must be relatively weak (≈4% for shear strains of 1.5, with a maximum value of ≈6% for very large strains). The splitting observations are comparably fit by a wide range of different viscosity models. Anisotropy due to melt-filled cracks aligned by stresses in the back arc flow models predicts fast directions roughly normal to observed values and thus cannot alone explain the observed splitting.

1. Introduction

Significant shear wave splitting has been observed in numerous subduction zone back arcs [e.g., Savage, 1999]. Can these data be explained by flow in the back arc mantle that is driven by coupling to local plate motions? In addition, assuming that upper mantle anisotropy is produced by strain-induced lattice preferred orientation (LPO) of olivine and orthopyroxene, what constraints can observed splitting place on the strength of the LPO that develops for a given amount of strain or on where LPO develops in the mantle?

Because of the systematic relationship of olivine and orthopyroxene LPO to strain, observational constraints on mantle anisotropy have a unique potential to map patterns of mantle flow. Possible forms for this relationship have been revealed through deformation experiments and naturally deformed mantle samples [Nicolas and Christensen, 1987; Mainprice and Silver, 1993; Zhang and Karato, 1995; Ismail and Mainprice, 1998], numerical flow models [McKenzie, 1979; Ribe, 1989, 1992; Ribe and Yu, 1991; Wenk et al., 1991; Chastel et al., 1993; Tommasi et al., 1996; Blackman et al., 1996; Tommasi, 1998], and laboratory flow simulations [Buttles and Olson, 1998]. Blackman et al. [1996] calculated seismic travel time anomalies for olivine LPO patterns induced by upwelling flow at mid-ocean ridges, and Tommasi et al. [1996] and Tommasi [1998] investigated LPO patterns produced by a cooling oceanic plate shearing over asthenospheric mantle. In this study we focus on subduction zone regions where abundant shear wave splitting observations exist, and we predict shear wave splitting due to olivine and orthopyroxene LPOs in a range of subduction zone flow models. However, anisotropy from other types of mantle fabrics, for instance small-scale aligned heterogeneities such as pockets of partial melt [Kendall, 1994], may also exist. We therefore explored the anisotropy that might be produced by melt-filled cracks aligned by stress within the subduction zone flow models.

We modeled subduction zone anisotropy with two-dimensional flow and strain models, yielding results that are applicable to subduction zone regions where mantle flow and anisotropy appear to be roughly two dimensional. In Tonga, Izu-Bonin, and the Marianas, shear waves that predominantly sample the back arc mantle yield splitting fast directions that are approximately parallel to the azimuth of subducting plate motion [Bowman and Ando, 1987; Fischer and Wiens, 1996; Fouch and Fischer, 1996, 1998; Fischer et al., 1998]. These regions also have in common active or recent back arc extension roughly parallel to the direction of subducting plate motion. In contrast, splitting fast directions observed at stations closer to the Tonga trench, near or within the arc, are more normal to subducting plate motion [Smith et al., 1998]. However, these phases spend a larger percentage of their paths in the subducting slab and reflect structure in the shallow corner of the mantle wedge not sampled by phases recorded at the more western back arc stations; we defer their consideration to future work. Shear wave splitting has also revealed the presence of anisotropy in the mantle wedges of
the southern Kuril, Japan, Aleutian, and New Zealand subduction zones. Throughout most of these regions, observed fast directions are highly oblique to the azimuth of subducting plate motion and are roughly parallel to the arc and/or to major strike-slip zones in the upper back arc plate [Iida and Obara, 1995; Yang et al., 1995; Fouc and Fischer, 1996; Sandvol and Ni, 1997; Fischer et al., 1998; Marson-Pidgeon et al., 1999; Audone et al., 2000; Brisbourne et al., 1999]. These regions require three-dimensional plate motion boundary conditions and are thus beyond the scope of this study.

2. Shear Wave Splitting Observed in Tonga

The Tonga subduction zone provides an excellent setting in which to test two-dimensional models of back arc flow against observed shear wave splitting. Not only do abundant shear wave splitting measurements exist for stations in the northern Tonga back arc (Figure 1), but these phases have long paths through the back arc wedge with only short paths within the subducting slab, minimizing the influence of slab anisotropy on the observed splitting [Fischer and Wiens, 1996].

For local S phases observed at back arc stations in Fiji (Labasa (LB), Lakeba (LKB), and Lauotoka (LTU)) from the Southwest Pacific Seismic Experiment [Wiens et al., 1995], and Monasavu (MSVF) from the Incorporated Research Institutions for Seismology (IRIS) Global Seismic Network, shear wave splitting fast directions have a distribution that is strongly peaked at an azimuth of -60°, parallel to the absolute motion of the Pacific plate (Figure 1). Splitting times lie in the range of 1 ± 0.5 s. These observations hold true regardless of the depth of the source or the azimuth of the shear wave path (Figure 2). If these data are modeled using a layer of mantle anisotropy with uniform orientation and strength, the bottom of this layer cannot exceed 435 km, indicating that the anisotropy cannot extend significantly into the mantle transition zone [Fischer and Wiens, 1996]. However, strain, LPO, and anisotropy in the back arc are not likely to be uniform, particularly given the probable entrainment of mantle by the subducting slab and upper plate. We therefore turned to numerical flow models to determine whether the shear wave splitting data are consistent with back arc flow that is driven by coupling to the subducting and upper plates.

3. Viscous Flow Models

Models of two-dimensional viscous, incompressible mantle flow driven by prescribed plate motions were formulated using finite element approximations. A standard penalty function formulation to enforce incompressibility with linear quadrilateral elements and a uniform viscosity within each element [e.g., Reddy, 1993] was employed. The model geometry and boundary conditions in a vertical plane parallel to the direction of subducting plate motion are defined in Figure 3. These boundary conditions approximate the components of Tonga region plate motions that lie in the plane of the model (parallel to the azimuth of Pacific plate motion). Smaller plate motion components normal to the model plane also exist [Gripp and Gordon, 1990; Bevis et al., 1995] but were ignored in the simple two-dimensional models of this study. Use of more realistic three-dimensional plate motion boundary conditions is the subject of ongoing work [Hall et al., 1999].

The right boundary represents the top side of a slab subducting at a prescribed rate (10 cm/yr) and with a dip of 60°. The real dip of the Tonga slab at shallow depths (< 100 km) is overestimated in these models (Figure 3), resulting in underestimation of the true distance between the back arc spreading center and the trench (Figure 1). However, this simplification has a substantial effect only on flow in the shallowest corner of the mantle wedge, a region not sampled by the phases that we are attempting to model.

The top boundary of the grid represents the bottom of the overriding lithosphere with a back arc spreading center located 190 km from the trench. In the model reference frame, the overriding plate above the slab is fixed, and back arc spreading is represented by the motion of the plate to the left of the spreading center (also at 10 cm/yr). Although these boundary conditions reflect absolute plate motion components in the plane of the model, they do not completely capture the motion of the back arc spreading center relative to the trench over time. If spreading were symmetric, the spreading center would migrate away from the trench at half the spreading rate. The earliest stage of plate growth in the Lau Basin involved rifting of the back arc plate, followed by the onset of true seafloor spreading at the Eastern Lau Spreading Center (19.25°S-23°S) at ~5.5 Ma and the later development of the Central Lau Spreading Center (17.5°S-19.25°S) [Parson and Wright, 1996] (Figure 1). Although total plate growth has been greater to the west of the Eastern Lau Spreading Center, particularly in the southeastern Lau Basin, ridge to trench distance has still increased by as much as 100 km since seafloor spreading began. However, increases in ridge to trench distance would significantly affect flow velocity. The 240 km distance to the spreading center and the trench is underestimated in these models (Figure 3), resulting in underestimation of the true distance between the back arc spreading center and the trench (Figure 1). However, this simplification has a substantial effect only on flow in the shallowest corner of the mantle wedge, a region not sampled by the phases that we are attempting to model.

The bottom boundary of the model, at a depth corresponding to the base of the mantle, is shear stress free with vanishing vertical velocity. On the left boundary the vertical velocity and horizontal normal deviatoric stress vanish. These latter conditions are not physically determined since this would require calculating flow outside the region shown in Figure 3. The conditions imposed minimize the effect of this boundary on flow above the subducting slab by allowing mantle flow into and out of the region of interest without viscous resistance from the rest of the mantle. The results shown here are computed with a grid of 49 quadrilateral elements in the vertical and horizontal directions. As indicated in Figure 3, element sizes are smallest in the mantle wedge above the subducting slab but provide good resolution throughout the model domain.

Finite strains were calculated by accumulating the strain along flow lines. Flow lines were calculated using a fourth-order Runge-Kutta algorithm. The finite strain analysis follows the treatment of Malvern [1969] with the numerical approach suggested by McKenzie [1979]. The accuracy of our implementation of the method was checked by applying it to simple flow fields resulting from pure shear and simple shear.

We considered models with several different viscosity distributions to evaluate the possible effect of viscosity variations on the seismic observations. Two models examined viscosity variations adjacent to the slab by
Figure 1. Shear wave splitting in the Tonga subduction zone observed at back arc stations in Fiji (triangles) from the Southwest Pacific Seismic Experiment [Wiens et al., 1995] (LBSA, Labasa; LKBA, Lakeba; and LTKA, Lautoka) and from the Incorporated Research Institutions for Seismology (IRIS) Global Seismic Network (MSVF, Monasavu). (a) Regional bathymetry and plate motions. Bold arrows denote the direction and relative rates of absolute Pacific plate motion, absolute Australian plate motion, and absolute Tonga arc motion [Gripp and Gordon, 1990; Bevis et al., 1995]. Double lines show centers of active and recent (< 6 Ma) back arc spreading [Hawkins, 1995]. Bathymetry is contoured in 1000-m intervals. (b-d) Local S splitting measured at stations LBSA (Figure 1b), LKBA (Figure 1c), and LTKA and MSVF (Figure 1d). Figures 1b-1d show magnification of area outlined by the box in Figure 1a. Splitting parameters are plotted at the earthquake source and are shown as vectors parallel to $\phi$ (fast direction) with lengths scaled to $\Delta t$ (splitting time); corresponding 95% confidence limits are represented by thick and thin line segments. A 1 s $\Delta t$ reference vector is shown at bottom right of Figure 1a. Fast directions for paths in Tonga are roughly parallel to Pacific plate motion.
considering a roughly 160-km-thick layer in which the viscosity is increased or decreased by a factor of 10. Water from dehydration within the slab that migrates upward and hydrates the overlying mantle could produce such a reduction in viscosity [Chopra and Paterson, 1984; Hirth and Kohlstedt, 1995]. The effect of this viscosity reduction on the flow and strain patterns would be greatest if it were confined near the slab, thus partially decoupling the overlying mantle from the motion of the slab. A viscosity increase adjacent to the slab should be created by the cooling that results from heat transfer to the cold slab. It is not clear whether the effect of water or temperature would be larger, so we chose to consider only simple extreme models. We also considered the effect of a low-viscosity zone at the base of the overriding plate. This might be present in the Earth because of the pressure dependence of viscosity or because of mantle upwelling and melting due to back arc spreading. The role of viscosity variations in influencing the observed shear wave splitting was assessed by comparing models with these extreme cases of variable viscosity to a uniform viscosity reference case. Note that only relative, not absolute, values of the viscosity affect the velocity and strain rate fields that are of primary interest here. Absolute values of the viscosity determine the magnitude of stresses.

An example of the flow field in the complete model domain is shown in Figure 3, with flow lines superimposed on the viscosity field (shading in Figure 3). In this example a low-viscosity zone above the subducting slab is shown by the darker shaded region. Because of the low-viscosity zone, flow is driven primarily by plate motion at the top boundary except in the immediate vicinity of the slab.

Flow lines and strain ellipses for fluid particles moving along them are shown in Figure 4 for the four viscosity models defined above. The strain ellipses represent finite strain that begins to accumulate as material rises through a depth of 410 km, taken as the depth above which strain due to dislocation creep dominates that due to diffusion creep. Only strain accumulated in the dislocation creep regime should create lattice preferred orientation resulting in seismic anisotropy. Flow lines in the three different variable viscosity cases compared to uniform viscosity illustrate the large effect of the modest viscosity variations considered on the flow structure. With a low-viscosity zone beneath the upper plate (Figure 4b) and with a high-viscosity zone adjacent to the slab (Figure 4d), the motion of the slab entrains significantly more material than in the uniform viscosity case. With a low-viscosity zone next to the slab (Figure 4c) the slab entrains less material, and flow, except in the region immediately adjacent to the slab, is dominated by the motion of the upper plate.

4. Mapping Strain to LPO and Anisotropy

To infer anisotropic velocities from a given flow/strain model, one must make assumptions regarding (1) where in the mantle olivine and orthopyroxene LPOs develop owing to
dislocation creep, (2) the orientation of olivine and orthopyroxene LPO with respect to strain geometry, (3) the strength of olivine and orthopyroxene LPO with respect to strain magnitude, and (4) specific elastic coefficients that represent olivine and orthopyroxene anisotropy.

In these calculations, strain-induced LPO of olivine and orthopyroxene was assumed to develop in the back arc mantle from a depth of 410 km to the base of the upper plate, although in section 7 we also discuss the effects of confining anisotropy to high shear stress regions of the upper mantle. The former assumption is consistent with the results of earlier modeling in which observed splitting times were fit with layers of uniform anisotropy. Models in which anisotropy extends to depths of 410 km provided good fits to the shear wave splitting observed in Tonga, but the lack of sources at depths shallower than 350 km made it difficult to assess whether anisotropy is required to this depth [Fischer and Wiens, 1996]. However, in the Japan and southern Kuril subduction zones, splitting times increase systematically over source depth ranges of 200 - 400 km and 300 - 600 km, respectively, and indicate the existence of anisotropy over a broad range of upper mantle depths [Fouch and Fischer, 1996]. On the basis of olivine rheology, Karato and Wu [1993] suggested that a transition from dislocation creep at shallower depths to diffusion creep at greater depths occurs at depths of 200-300 km within relatively warm regions of the upper mantle. However, given uncertainties in dislocation and diffusion creep activation volumes, grain size, and stress, it is not implausible to assume that dislocation creep in olivine occurs throughout the back arc upper mantle.

In regard to the geometry of peridotite LPO, the orientation of the olivine $a$ axis at any point in the model was defined by the dip and azimuth of the long axis of the local finite strain ellipse, and the olivine $c$ axis was set normal to the plane of the model which is equivalent to the flow plane. This assumption agrees with olivine $a$ axis orientations observed in an extensive database of olivine aggregates from ridge, subduction zone, and cratonic environments [Ismail and Mainprice, 1998]. It is only partially consistent with the results of experimental deformation of olivine under simple shear [Zhang and Karato, 1995; Zhang et al., 2000]. These experiments indicate that at lower temperatures (1473'K) the average olivine $a$ axis alignment is within 15' of the long axis of the strain ellipse for strains up to 1.1. However, at higher temperatures (1573'K), where recrystallization processes appear to play a greater role in LPO development, the average olivine $a$ axis for relict grains starts out subparallel to the long axis of the strain ellipse but rotates parallel to the flow direction at strains greater than 1.0. For recrystallized grains, $a$ axes are bimodally oriented parallel to the shear direction and normal to the maximum principal compressive stress [Zhang et al., 2000]. Several numerical studies of LPO development in polycrystalline olivine or peridotite aggregates found that simple shear generated girdles of olivine $a$ axes in the foliation plane, normal to the short axis of the finite strain ellipse [Ribe and Yu, 1991; Wenk et al., 1991; Ribe, 1992; Tommasi, 1998]. While the anisotropy and shear wave splitting times predicted by such $a$ axis girdles would be somewhat weaker than results obtained with the assumptions of this study, the predicted fast directions would be similar. In more complex two-dimensional flow fields [Chastel et al., 1993; Blackman et al., 1996] olivine $a$ axes are not consistently aligned with either finite strain or flow direction, although $a$ axes do lie within roughly 20' of the maximum finite extension direction over broad regions of the models.

Numerical, experimental, and natural peridotite investigations concur that the strength of olivine and
orthopyroxene LPO increases rapidly with strain for small strains and then levels off, but the absolute amount of LPO and anisotropy at a given strain varies from study to study [Ribe, 1992; Mainprice and Silver, 1993; Zhang and Karato, 1995; Ismail and Mainprice, 1998; Tommasi, 1998; Zhang et al., 2000]. To reflect this variability, we calculated splitting for four sets of elastic coefficients, each with its own implications for the evolution of anisotropy with strain. Since a particular set of elastic coefficients corresponds to only one shear strain magnitude, we scaled the strength of anisotropy up or down using the anisotropy versus shear strain relationship of Tommasi [1998]. This curve shows anisotropy progressively approaching its maximum value with increasing strain, reaching ~90% of its maximum value at shear strains of ~5.0.

Ismail and Mainprice [1998] found that the strength of anisotropy (defined as $2(V_f - V_s) / (V_f + V_s)$, where $V_f$ and $V_s$ are the propagation velocities of the fast and slow shear waves in a given propagation direction) reaches a maximum of 10 - 15%. We employed the elastic coefficients for the average of

Figure 4. Streamlines and finite strain ellipses for flow above the subducting slab for different viscosity models: (a) uniform viscosity, (b) viscosity reduced by a factor of 10 in a 100-km-thick layer (lightly shaded region) adjacent to the upper plate, (c) viscosity reduced by a factor of 10 in a 160-km-thick layer (lightly shaded region) adjacent to the subducting slab, and (d) viscosity increased by a factor of 10 in a 160-km-thick layer (darker shaded region) adjacent to the subducting slab. Strain ellipses represent the accumulated strain for fluid elements moving along flow lines. Strain begins to accumulate as material rises through a prescribed depth (410 km; see text for discussion). Physical motivation for these models is discussed in the text. Ray paths for which splitting was predicted are shown by thinner lines. Triangles at top of model show the locations of the Tonga back arc stations projected onto the model plane. For very large finite strains, the length of the long axis of the strain ellipse is held at a constant maximum value to improve figure clarity.
their olivine aggregate data set, and in order to achieve a maximum olivine anisotropy of 12%, we assigned these coefficients to a shear strain of 1.5. We also tested the elastic coefficients of Zhang et al. [2000] from their olivine simple shear experiments at 1473K and 1.1 strain and 1573K and 1.5 strain. These two sets of elastic coefficients were assigned to their experimentally determined shear strains.

The Zhang et al. [2000] and Ismail and Mainprice [1998] coefficients were diluted to a roughly peridotitic composition (70% olivine and 30% enstatite) using single-crystal orthopyroxene coefficients [Frisillo and Barsch, 1972]. The orthopyroxene c axis was assumed to be parallel to the olivine a axis, and the orthopyroxene a axis was parallel to the olivine b axis. This orientation of orthopyroxene was observed in the kimberlite nodules of Mainprice and Silver [1993]. When orthopyroxene anisotropy with this orientation is added to a predominantly olivine matrix, its effect is to weaken the net strength of the anisotropy by a few percent without altering the fast axis of anisotropy defined by the olivine. Use of single-crystal orthopyroxene coefficients, as opposed to the more weakly anisotropic coefficients expected for a realistic orthopyroxene LPO, will slightly reduce the overall strength of the anisotropy for these three models. Including the effects of orthopyroxene, the maximum anisotropies implied by the Ismail and Mainprice and Zhang et al. 1473K coefficients were 13.1 and 13.8%, respectively. The Zhang et al. 1573K coefficients implied a maximum anisotropy of 19.9%. This value is unrealistically high given that it exceeds the anisotropy possible with single crystal olivine and orthopyroxene. It suggests that in the 1573K Zhang et al. experiments LPO and anisotropy reach their maximum values at smaller shear strains than in the anisotropy versus shear strain relationship assumed here. Therefore, these coefficients were not used to calculate shear wave splitting.

Finally, in order to provide a model with weaker anisotropy, we assumed that the elastic coefficients of Mainprice and Silver [1993] correspond to a shear strain of 1.5, producing a maximum anisotropy of ~6%. The modal composition for the Mainprice and Silver coefficients is 65% olivine, 31% enstatite, and 4% garnet.

5. Calculating Shear Wave Splitting

Shear wave splitting was predicted from sources at 400- and 600-km depth at the slab-mantle interface on ray paths contained in the plane of the two-dimensional models (Figure 5). These paths were computed relative to the AK135 radial earth model [Kennett et al., 1995] and are comparable to observed shear wave paths with back azimuths roughly parallel to Pacific plate motion, although the real paths originate from a more continuous range of source depths and from sources that may be located up to 50 km to the east or west of the planar slab interface assumed in the model. Because the dip of the Tonga slab shallows in the transition zone [Van der Hilst, 1995] and the deep Tonga seismic zone exhibits along-arc curvature, some sources for the observed shear wave splitting from the complete Tonga data set are located up to 300 km from the slab interface in the model and may not be adequately modeled by the shear wave splitting calculated here.

Splitting parameters were obtained for each path by 1) calculating incremental splitting due to the local mantle anisotropy sampled by a shear phase at 10-km depth intervals, 2) integrating the effects of the incremental splitting on the shear phase particle motions from the source to the surface, and 3) calculating shear wave splitting from the predicted surface particle motions using the same method (that of Silver and Chan [1991]) that was applied to the observations. Horizontal shear wave particle motion at the source, \( u_0 \), was assigned equal radial and transverse amplitudes and a period of 5 s, roughly comparable to most phases analyzed for shear wave splitting in Tonga. In each 10-km depth increment the splitting due to local anisotropy was calculated using the Christoffel equation, \( \det (\rho^{-1} \epsilon_{ijkl} \epsilon_i \epsilon_j - \lambda \delta_{ik}) = 0 \), where \( \epsilon_{ijkl} \) denotes the elastic coefficients in the model at the given point along the path (appropriately rotated so that the olivine a axis is aligned with the long axis of the local strain ellipse), the density is given by \( \rho \), and \( \epsilon_i \) and \( \epsilon_j \) are the directional cosines of path. For the three possible positive eigenvalues given by \( \lambda_i \), if \( \lambda_1 > \lambda_2 > \lambda_3 \), then \( V_f = (\lambda_2)^{1/2} \) and \( V_r = (\lambda_3)^{1/2} \) are the velocities of the fast and slow shear wave particle motions. The splitting time for the nth depth increment, \( \delta_{tn} \), is equal to \( L (V_r^{-1} - V_f^{-1}) \), where \( L \) is the path length of the phase over the 10-km depth range. The fast direction, \( \phi_{tn} \), is equal to the horizontal projection of the eigenvector of \( \rho^{-1} \epsilon_{ijkl} \epsilon_i \epsilon_j \) that corresponds to \( \lambda_2 \). For a path with m depth increments, \( u \), the horizontal shear wave particle motion at the surface, is:

\[
u(\omega) = \left[ \prod_{n=1}^{m} R^1(\phi_n) D(\delta_{tn}) R(\phi_n) \right] u_0(\omega)
\]

where

\[
R(\phi_n) = \begin{bmatrix} \cos \phi_n & \sin \phi_n \\ -\sin \phi_n & \cos \phi_n \end{bmatrix}
\]

and

\[
D(\delta_{tn}) = \begin{bmatrix} e^{i\omega \delta_{tn} / 2} & 0 \\ 0 & e^{-i\omega \delta_{tn} / 2} \end{bmatrix}.
\]

The splitting parameters for the complete path were defined to be the values of \( \phi \) and \( \delta t \) which, when removed from \( u \), yielded the most linear particle motion. In practice, the splitting parameters were taken to be the \( \phi \) and \( \delta t \) values for which the covariance matrix between the components of \( u \) (corrected for \( \phi \) and \( \delta t \)) was the most singular [Silver and Chan, 1991].

This approach to calculating shear wave splitting oversimplifies the frequency-dependent interaction of the shear wavefield with anisotropic structure [e.g., Rümpker and Silver, 1998], and care must be taken to keep \( \delta t \) values small with respect to the period of \( u_0 \). However, shear wave splitting determined with this method was tested against propagator matrix synthetic seismograms [Keith and Crampton, 1977] for stacks of anisotropic layers containing fast direction and anisotropy strength variations comparable to that found in the flow models. At periods corresponding to the waveforms from which the shear wave splitting observations were made, and for the relatively steep incidence angles used in this study, the splitting predicted by the two approaches agrees to within ±3° for fast directions and ±0.2 s for splitting times.
6. Shear Wave Splitting for Ubiquitous Upper Mantle Anisotropy

Predicted splitting times and fast directions (Figure 5) were obtained for the four different viscosity models, assuming the ray paths shown in Figure 4, and the three sets of elastic coefficients based on Mainprice and Silver [1993], Isma'il and Mainprice [1998], and the Zhang et al. [2000] experiments at 1473 K. For all models and elastic coefficients, predicted fast directions are parallel to the azimuth of Pacific plate motion, in agreement with the fast directions observed in the Tonga back arc. This result is not surprising given that in the two-dimensional flow models the long axes of the strain ellipses (and hence the olivine a axes) have azimuths that are in the plane of the model and, over much of the region sampled by the shear wave paths, plunges that are 45°-90° away from the local direction of propagation.

The two most striking features of the predicted splitting times are the large variation in the magnitudes of the splitting times predicted for the different sets of elastic coefficients and the lack of splitting time dependence on viscosity structure. For both the 400- and 600-km source depths (Figure 4) the Mainprice and Silver [1993] coefficients (scaled to a maximum anisotropy of ~6%) gave splitting times of the order of 1-1.5 s, and the Isma'il and Mainprice [1998] and Zhang et al. [2000] coefficients for 1473 K gave splitting times of 2-5 s (Figure 5). These differences in overall splitting time magnitude are a direct result of variations in the strength of the olivine LPO for similar amounts of strain. The boxes in each plot of Figure 5 indicate the range of splitting times observed in Tonga for phases with back azimuths within 20° of absolute Pacific plate motion. The splitting times for the Mainprice and Silver elastic coefficients (scaled to a maximum anisotropy of ~6%) provide a good fit to the range of observed splitting times, while the splitting times for the Isma'il and Mainprice and Zhang et al. elastic coefficients overestimate observed values. Differences in viscosity structure, flow, and strain between the four models have relatively little impact on the predicted splitting times, although a few subtle variations do occur. Because the variations between splitting times for the four viscosity models are small relative to the scatter in the observed splitting times, the observations cannot be used to discriminate between the viscosity structures. In other words, all four viscosity models and flow regimes are equally consistent with the shear wave splitting observed in Tonga.

7. Confining Dislocation Creep to High Stress Regions of the Back Arc Mantle

Taken at face value, the above results argue for relatively weak LPO at a given strain, for instance, ~4% at shear strains of 1.5 with a maximum anisotropy of ~6%. However, the assumption made in these models that strain-induced LPO develops over the entire upper mantle may not be valid. For instance, shear stress, temperature, water content, and grain size all affect the relative strength of dislocation creep versus diffusion creep [Karato and Wu, 1993]. Shear stress and temperature certainly vary within the back arc mantle, while variations in water content and grain size are likely.

Here we investigate the possibility that LPO of mantle minerals and anisotropy preferentially develop in higher shear stress regions of the back arc mantle. The physical basis for this scenario is that strain rate due to dislocation creep varies as shear stress to a power of 3.0-3.5, whereas strain rate due to diffusion creep varies only as shear stress to a power of 1.0 [Karato and Wu, 1993]. As shear stress rises, the relative contribution to strain rate by dislocation creep is increased for mantle material of a given grain size. An important caveat is that dynamic recrystallization in the dislocation creep regime will decrease grain size with increasing shear stress [Karato and Wu, 1993], thus driving mantle conditions back toward those favorable to diffusion creep. However, given uncertainties in the processes of grain growth and recrystallization, we focus on the effects of shear stress at a
Figure 6. The direction and magnitude of deviatoric stresses within the four flow models: (a) uniform viscosity, (b) viscosity reduced by a factor of 10 in a 100-km-thick layer (lightly shaded region) adjacent to the upper plate, (c) viscosity reduced by a factor of 10 in a 160-km-thick layer (lightly shaded region) adjacent to the subducting slab, and (d) viscosity increased by a factor of 10 in a 160-km-thick layer (darker shaded region) adjacent to the subducting slab. Line segments are aligned parallel to the direction of the maximum compressive stress, and their length scales to the maximum stress difference. For very large maximum stress difference values within the high-viscosity zone, line segment length is held at a constant maximum value to improve figure clarity. Ray paths for which splitting was predicted are shown by thinner lines. Triangles at top of models show the locations of the Tonga back arc stations projected onto the model plane.

Shear stress patterns in the four viscosity and flow models are simple and intuitive. In the uniform viscosity model, shear stresses are high in the mantle wedge near the subducting and upper plates (Figure 6a) but decrease rapidly away at distances of more than 200 km from either plate. The effect of low-viscosity zones is to locally reduce stress. In the model with a low-viscosity layer at the base of the upper plate (Figure 6b), a high-stress region occurs only near the subducting slab, and in the model with a low-viscosity layer at the slab-wedge interface (Figure 6c), a high-stress region occurs only at the base of the upper plate. In the model with a high-viscosity layer at the slab-wedge interface (Figure 6d), high shear stresses occur near the subducting and upper plates, and the width of the high-stress region near the slab is increased relative to the uniform viscosity model.

Because of uncertainties in the value of shear stress that might mark the transition from diffusion to dislocation creep and in the rate with which diffusion creep might erase LPO previously developed under dislocation creep, the extent and strength of LPO in the high-stress regions is difficult to assess. We therefore assumed that LPO develops in high-stress regions that are 150-200 km in thickness, and we used the incremental splitting calculated in the models of section 6 to estimate the net splitting for these high-stress zones. For the models with uniform viscosity and high- and low-viscosity zones above the slab, splitting accrued in their high-stress regions provides a good fit to the observations but only if the maximum strength of anisotropy is more than 6%. Because the viscosity model with a low-viscosity zone at the base of the upper plate contains high stresses only near the slab, little to no splitting is predicted for sources at 600 km or deeper if anisotropy is assumed to occur only shallower...
than 410 km, leading to an obvious misfit with the observations (Figure 2). However, if significant anisotropy with a fast direction in the plane containing Pacific plate motion is allowed to occur in the transition zone, perhaps caused by LPO development in modified spinel within the transition zone or because of LPO created in the upper mantle that survived the olivine to modified spinel phase transition [Mainprice et al., 1990], splitting predicted for this high-stress region fits the observed distribution of splitting times with depth reasonably well. As was obtained with the models in which LPO was assumed to develop throughout the upper mantle (section 6), the splitting estimated for the high-stress regions implied by the four viscosity models does not clearly rule out or favor any of these viscosity structures. In addition, allowing LPO development and anisotropy to be localized in limited high-stress regions of the upper mantle increases the strength of anisotropy permissible for a given amount of strain and indicates that the anisotropy-strain relationships implied by Ismail and Mainprice [1998] and the Zhang et al. [2000] coefficients for 1473 K could provide acceptable fits to the splitting observations.

If anisotropy is confined to such high-stress regions, phase path lengths through the anisotropic mantle wedge would still be longer than most phase path lengths in the slab, but if anisotropy had similar strengths in the slab and wedge, slab contributions to observed splitting times could be significant for some phases. However, because path lengths in the slab vary by an order of magnitude [Fischer and Wiens, 1996], whereas observed splitting times vary by a factor of 3 or less (Figure 2), it is unlikely that slab anisotropy alone could explain the shear wave splitting observations.

8. Melt Pocket Anisotropy

The presence of small-scale oriented isotropic heterogeneity, such as melt-filled cracks or pockets, can also produce significant mantle anisotropy [Kendall, 1994]. We therefore examined the alternative hypothesis that the splitting observed in the Tonga back arc mantle, as well as in other subduction zones, may be attributed to melt-filled pockets or cracks oriented by the local direction of the stress field. Following the experiments of Zimmerman et al. [1999], we assumed that melt-filled cracks would align at 20°-30° from the maximum deviatoric compressive stress. The crack faces would strike normal to the plane of the two-dimensional flow models, and their projection on to the model plane would lie 20°-30° away from the maximum compressive stress difference, whose local orientation is shown in Figure 6. In general, the fast shear polarization would be normal to the model plane, except where rays are normal to the crack faces and particle motions lie within the crack face plane, in which case no splitting would occur. Apparent splitting fast directions would be aligned normal to the plane of the model and would thus be inconsistent with the observed fast directions. We therefore reject melt-filled cracks as the sole explanation of the observed shear wave splitting. However, relatively weak anisotropy due to melt-filled cracks could exist in addition to stronger LPO-induced anisotropy, in which case the melt anisotropy would reduce the net splitting time without altering the trend of the fast directions from an azimuth parallel to Pacific plate motion.

9. Implications for Flow and Anisotropy in the Back Arc Mantle

The primary conclusion of this study is that it is possible to quantitatively model shear wave splitting observed on back arc paths in the Tonga subduction zone with LPO anisotropy that is induced by two-dimensional back arc flow, where back arc flow is driven by the local motions of the subducting and upper back arc plates. The lack of splitting time sensitivity to viscosity distribution suggests that this conclusion is valid regardless of viscosity structure. In contrast, melt-filled cracks aligned by stresses due to back arc flow can be rejected as the sole explanation for the observed splitting. Although the presence of weak anisotropy due to melt-filled cracks cannot be ruled out, its strength must be substantially less than the LPO anisotropy created by back arc flow.

This study also places constraints on the relationship between LPO anisotropy strength and strain magnitude in the upper mantle. If the entire upper mantle deforms by dislocation creep, this study indicates that the strength of the LPO that develops for a given amount of strain must be relatively weak. For instance, a mantle with a maximum anisotropy of roughly 6% for shear strains of 5 or more would provide an acceptable fit to the observed splitting. However, if LPO anisotropy occurs only in limited high-stress regions of the mantle, or if the presence of melt anisotropy reduces the net amount of shear wave splitting produced in the upper mantle, then LPO and anisotropy for a given amount of back arc strain may be substantially stronger.

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