

Seismological evidence for girdled olivine lattice-preferred orientation in oceanic lithosphere and implications for mantle deformation processes during seafloor spreading

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Key Points:

- We develop an *in situ* elastic tensor for oceanic lithosphere that incorporates co-located compressional and shear anisotropy constraints
- Seismic anisotropy is consistent with corner flow during spreading and shear strains of 250–400%
- Girdled D-type olivine fabric implies activation of multiple olivine easy slip systems during mid-ocean ridge spreading

Abstract

Seismic anisotropy produced by aligned olivine in oceanic lithosphere offers a window into mid-ocean ridge dynamics. Yet, interpreting anisotropy in the context of grain-scale deformation processes observed in laboratory and natural olivine samples has proven challenging due to the vast length scale differences. We bridge this observational gap by estimating an *in situ* elastic tensor for oceanic lithosphere using compressional- and shear-wave speed anisotropy observations. The elastic model for the upper 7 km of the mantle, NoMelt_SPani7, is characterized by a fast azimuth parallel to the fossil-spreading direction consistent with corner-flow deformation fabric. This model is compared with a database of 123 petrofabrics from the literature to infer olivine crystallographic orientations and shear strain accumulated within the lithosphere. Direct comparison to olivine deformation experiments indicates strain accumulation of 250–400% in the shallow mantle. We find evidence for D-type olivine lattice-preferred orientation (LPO) with fast [100] parallel to the shear direction and girdled [010] and [001] crystallographic axes. D-type LPO implies similar amounts of slip on the (010)[100] and (001)[100] easy slip systems during mid-ocean ridge spreading; we hypothesize that grain boundary sliding during dislocation creep relaxes strain compatibility, allowing D-type LPO to develop in the shallow lithosphere. Deformation dominated by dislocation-accommodated grain-boundary sliding (disGBS) has implications for *in situ* stress and grain size during mid-ocean ridge spreading and implies grain-size dependent deformation, in contrast to pure dislocation creep.

Plain Language Summary

Earth’s upper mantle is composed primarily of the mineral olivine, which responds to deformation by organizing its seismically fast axis in the flow direction. During seafloor spreading, olivine crystals align with the spreading direction and become frozen into the lithosphere preserving the near-ridge deformation history. The resulting rock fabric can be observed in place via the directional dependence of seismic wavespeeds (seismic anisotropy) as well as in hand-sample rocks collected from the field. However, interpreting seismic anisotropy observations in the context of laboratory and field data remains a challenge, due to large differences in length scale and incomplete seismic constraints. Here, we bridge this observational gap by incorporating multiple data types to solve for the complete anisotropic structure of oceanic lithosphere that formed ~ 70 Myr ago. By comparing our model to

laboratory data, we infer the magnitude of shear strain and style of olivine deformation during seafloor spreading for the first time. Our results indicate large shear strains and an olivine fabric type different to that typically assumed, implying the deformation is sensitive to the presence of grain boundaries even though most of the strain is produced by dislocations. This has new implications for the formation and evolution of oceanic plates.

1 Introduction

Observations of seismic anisotropy in the ocean basins provide unparalleled insight into Earth's mantle circulation patterns, including the plate-tectonic process of seafloor spreading (Hess, 1964; Forsyth, 1975; Nishimura & Forsyth, 1989; Montagner & Tanimoto, 1991; Gaherty et al., 2004; Toomey et al., 2007; Lin et al., 2016; Takeo et al., 2018; Mark et al., 2019). Based on olivine lattice-preferred orientation (LPO) observed in peridotites sampled from ophiolites and oceanic settings (Peselnick & Nicolas, 1978; Ben-Ismail & Mainprice, 1998; Michibayashi et al., 2006; Warren et al., 2008; Skemer et al., 2010) and in deformation experiments on olivine (Zhang & Karato, 1995; Zhang et al., 2000; Bystricky et al., 2000), upper-mantle anisotropy has historically been interpreted as evidence that deformation associated with mantle convection and plate tectonics occurs via dislocation creep in an olivine-rich upper mantle (Karato & Wu, 1993). This inference in turn has implications for key physical parameters such as grain size and stress, and more generally, the rheology of the upper mantle (Hirth & Kohlstedt, 2003).

Subsequent advances in laboratory deformation experiments on olivine have complicated this interpretation, however, illuminating variations in the activity of several olivine slip systems and associated LPO fabric types that may depend on mantle conditions such as stress, volatile content, and partial melting (Bystricky et al., 2000; Jung & Karato, 2001; Katayama et al., 2004; Jung et al., 2006; Qi et al., 2018). The three most commonly observed types of olivine LPO in experiments and natural settings are A-, E-, and D-type, each with their fast [100] crystallographic axis oriented sub-parallel to the shear direction but with variable orientations of the slow [010] and intermediate [001] axes (Figure 1). Due to incomplete constraints on *in situ* shear and compressional velocities, it has not been possible to distinguish between these three fabric types in oceanic lithosphere using traditional seismological observations, leading to renewed ambiguity of physical state and deformation processes during seafloor spreading (Karato et al., 2008).

Here, we utilize a unique set of co-located shear and compressional anisotropy observations in the central Pacific (Mark et al., 2019; Russell et al., 2019) to produce an exceptionally complete estimate of peridotite elasticity in the oceanic lithosphere. This new seismic model, NoMelt_SPani, is then combined with a database of 123 olivine petrofabrics reported in the literature from experimentally and naturally deformed rocks to constrain olivine LPO type and to quantify shear strain associated with seafloor spreading near the mid-ocean ridge (MOR). The collection of laboratory and field petrofabric observations provide a framework for quantitative interpretation of the *in situ* elasticity tensor in terms of degree of strain, composition, LPO type, and potentially deformation mechanism.

2 Data

2.1 Seismic observations at the NoMelt experiment

The NoMelt geophysical experiment in the central Pacific provided unique co-located compressional- and shear-wave constraints (V_P and V_S , respectively) on *in situ* lithosphere petrofabric over a 600×400 km² footprint with average seafloor age of ~ 70 Ma. It comprised a refraction survey that constrained V_{PH} and its azimuthal anisotropy in the upper ~ 7 km of the mantle (Mark et al., 2019) and a broadband ocean-bottom seismometer deployment that resolved the complete V_S structure via observations of both Rayleigh-wave (Lin et al., 2016) and Love-wave azimuthal anisotropy (Russell et al., 2019) in the upper 40 km of the lithosphere. Together, these independent compressional and shear observations of seismic anisotropy probe the complete elastic tensor of oceanic lithosphere.

2.2 Olivine fabric database

We have compiled a database of the elastic properties of 123 published olivine fabrics that includes 91 olivine samples deformed in the laboratory in direct shear (Zhang & Karato, 1995; Jung & Karato, 2001; Katayama et al., 2004; Jung et al., 2006) and high-strain torsion (Bystricky et al., 2000; Hansen et al., 2014, 2016) experiments, as well as 31 natural peridotite samples from diverse settings including ophiolites (Peselnick & Nicolas, 1978; Ben-Ismaïl & Mainprice, 1998; Michibayashi et al., 2006; Warren et al., 2008; Skemer et al., 2010), volcanic arcs (Mehl et al., 2003), and xenoliths and kimberlites from continental cratons (Ben-Ismaïl et al., 2001; Satsukawa et al., 2010). Of these samples,

42 (31 natural and 11 laboratory) have had LPO identified by the authors based on the orientations of their crystallographic axes: 12 have been identified as A-type, 6 E-type, and 24 D-type. The methods employed by these previous authors to calculate bulk seismic properties from individual crystallographic orientations generally follow a similar procedure and (1) nearly ubiquitously assume samples are composed of 100% olivine and (2) use the single-crystal olivine tensor of Abramson et al. (1997). For one harzburgite sample from the Oman ophiolite, the seismic properties were measured directly using ultrasonics, and therefore, contributions from other phases are inherently included (Peselnick & Nicolas, 1978). The temperature and pressure at which the elastic calculations are carried out can vary slightly among studies, but this has a negligible effect on seismic anisotropy (Ben-Ismail & Mainprice, 1998).

For all samples considered, the orientation of the elastic tensor with respect to the shear plane and shear direction were determined. Upon comparing to the seismic model, all samples in the database were oriented in the seismic reference frame: shear plane parallel to the X - Y plane defined by Earth’s surface and shear direction parallel to the X -axis defined by the fossil-spreading direction (FSD). The Z -axis is oriented perpendicular to Earth’s surface (i.e., perpendicular to the shear plane).

Estimates of shear strain associated with deformation are routinely measured for laboratory samples and range from undeformed ($\gamma \sim 0$) to $\gamma \sim 18.7$ in our dataset, but such estimates are rarely available for natural rocks. One exception is the Josephine shear zone in southwestern Oregon (Warren et al., 2008; Hansen & Warren, 2015), which has pre-existing foliations that provide passive strain markers that imply highly strained peridotites up to $\gamma \sim 20$ (Skemer et al., 2010).

3 Methods

3.1 Surface-wave inversion with P_n constraints

We invert previously measured surface-wave phase velocities for the shear and compressional velocities and anisotropy beneath the NoMelt array. Previously measured anisotropic Rayleigh- (5–150 s) and Love-wave (5–7.5 s) phase velocities from Russell et al. (2019) are inverted, while simultaneously satisfying V_P constraints in the upper ~ 7 km of the mantle from Mark et al. (2019) (“Weighted data, with gradients” in their Table 1). For consistency, the inversion is carried out to 400 km depth as in Russell et al. (2019), but

we focus here only on the upper ~ 7 km of the mantle, for which both P and S constraints exist.

Surface-waves traveling through a weakly anisotropic medium with orthorhombic symmetry can be described by 13 independent elastic parameters, composing the elastic stiffness tensor, C_{ij} . Following Russell et al. (2019), we divide the elastic tensor into two parts and invert for each separately:

$$C_{ij} = C_{ij}^{\text{TI}} + \delta C_{ij}, \quad (1)$$

where C_{ij}^{TI} is the transversely isotropic part that consists of Love’s five elastic parameters A , C , F , L , and N (Dziewonski & Anderson, 1981) describing azimuthally averaged Rayleigh- and Love-wave phase velocities, and δC_{ij} contains eight parameters (G_c , G_s , B_c , B_s , E_c , E_s , H_c , H_s) that describe the strength and direction of azimuthal anisotropy of Rayleigh and Love waves (see Appendix A).

First, we jointly invert the azimuthally averaged Rayleigh- and Love-wave phase velocities (c^{R} , c^{L}) for the five transverse isotropy parameters $V_{\text{SV}} = \sqrt{L/\rho}$, $V_{\text{SH}} = \sqrt{N/\rho}$, $V_{\text{PV}} = \sqrt{C/\rho}$, $V_{\text{PH}} = \sqrt{A/\rho}$, and $\eta = F/(A - 2L)$ (Dziewonski & Anderson, 1981):

$$\delta c^{\text{R}}(\omega) = \frac{c^2}{U} \int (K_{\text{PV}} \delta V_{\text{PV}}(r) + K_{\text{PH}} \delta V_{\text{PH}}(r) + K_{\text{SV}} \delta V_{\text{SV}}(r) + K_{\eta} \delta \eta(r)) dr, \quad (2)$$

$$\delta c^{\text{L}}(\omega) = \frac{c^2}{U} \int (K_{\text{SV}} \delta V_{\text{SV}}(r) + K_{\text{SH}} \delta V_{\text{SH}}(r)) dr, \quad (3)$$

where $\delta c(\omega) = c_0(\omega) - c^{\text{pre}}(\omega)$ is the residual between observed and predicted isotropic phase velocity at angular frequency ω at a given model iteration, U is group velocity, and $K_m(\omega, r)$ are the Fréchet derivatives for each model parameter m calculated using MI-NEOS. Poorly resolved parameters are scaled following Russell et al. (2019). We fix $\eta = F/(A - 2L)$ to PREM values (Dziewonski & Anderson, 1981), and V_{PV} is scaled such that $\phi^{-1} = (V_{\text{PH}}/V_{\text{PV}})^2$ remains equal to $\xi = (V_{\text{SH}}/V_{\text{SV}})^2$. In the upper 7 km of the mantle, we fix V_{PH} to values from Mark et al. (2019). Radial anisotropy, ξ , provides a proxy for the fast axis alignment in the horizontal ($\xi > 1$) or vertical ($\xi < 1$) plane.

Using the transversely isotropic model, we then invert the magnitude and fast azimuth of Rayleigh- and Love-wave azimuthal anisotropy for the depth dependent magnitude and direction of azimuthal anisotropy elastic parameters G , B , H , and E (see Ap-

pendix A). Expressed in terms of sine and cosine coefficients A_s and A_c , respectively, the relationships between surface-wave azimuthal anisotropy and the eight elastic parameters are given by (Montagner & Nataf, 1986)

$$A_{c2,s2}^R(\omega) = \frac{c}{U} \int \left(L K_L \frac{G_{c,s}}{L}(r) + A K_A \frac{B_{c,s}}{A}(r) + F K_F \frac{H_{c,s}}{F}(r) \right) dr, \quad (4)$$

$$A_{c2,s2}^L(\omega) = \frac{c}{U} \int \left(-L K_L \frac{G_{c,s}}{L}(r) \right) dr, \quad (5)$$

$$A_{c4,s4}^L(\omega) = \frac{c}{U} \int \left(-N K_N \frac{E_{c,s}}{N}(r) \right) dr, \quad (6)$$

where subscripts containing ‘2’ and ‘4’ indicate 2θ and 4θ azimuthal variations and superscripts again indicate wave type.

P_n waves sample V_{PH} directly beneath the Moho. The azimuthally averaged P_n velocity and its 2θ and 4θ azimuthal dependence provides independent constraints on elastic parameters A , B , and E , respectively via

$$\rho V_{Pn}(\theta)^2 = A + B_c \cos(2\theta) + B_s \sin(2\theta) + E_c \cos(4\theta) + E_s \sin(4\theta) \quad (7)$$

and enters the surface-wave inversion simply as prior constraints on those parameters. Below ~ 7 km beneath the Moho, at which P_n constraints terminate, and for the parameter H (which is poorly constrained by surface waves), we follow the general scaling approach described in Russell et al. (2019), whereby B and H scale directly with G . For this study, the direct B and G constraints in the upper 7 km suggest an empirical B/G scaling of ~ 1.5 that is applied throughout the model. Additionally, an H/G scaling of -0.11 is applied based on ophiolite samples (Peselnick & Nicolas, 1978; Ben-Ismaïl & Mainprice, 1998). The precise scaling between anisotropic parameters is complex and likely depends on several factors including LPO symmetry, fabric orientation relative to the horizontal plane, and the LPO of secondary phases (Montagner & Anderson, 1989). Thus, this study focuses on G , B , E , and ξ in the shallowmost lithosphere for which these parameters are directly constrained by the seismic observations and do not depend on *a priori* scaling assumptions.

3.2 Constructing the orthorhombic elastic tensor

A general elastic tensor is described by 21 independent elastic parameters. This tensor is simplified to only 9 parameters (A , C , F , L , N , G , B , H , E) if orthorhombic

symmetry is assumed and the three orthogonal crystallographic axes ([100], [010], [001]) are oriented along the principal directions (i.e., in the principal coordinate system). In this configuration, any crystallographic axis may be oriented along any of the three principal directions. This requirement is relaxed for the two horizontal directions in order to allow for arbitrary orientations of azimuthal anisotropy in the horizontal plane, resulting in an elastic tensor with 13 parameters (Montagner & Nataf, 1986). As one axis is assumed to be vertical (a requirement given that surface waves are horizontally propagating), dipping fabrics are not explicitly resolvable. The symmetric elastic tensor is given by

$$C_{ij} = \begin{pmatrix} A + B_c + E_c & A - 2N - E_c & F + H_c & 0 & 0 & \frac{1}{2}B_s + E_s \\ \cdot & A - B_c + E_c & F - H_c & 0 & 0 & \frac{1}{2}B_s - E_s \\ \cdot & \cdot & C & 0 & 0 & H_s \\ \cdot & \cdot & \cdot & L - G_c & G_s & 0 \\ \cdot & \cdot & \cdot & \cdot & L + G_c & 0 \\ \cdot & \cdot & \cdot & \cdot & \cdot & N - E_c \end{pmatrix}. \quad (8)$$

Although we solve for all 13 parameters of the tensor in the upper 7 km of the mantle, only 9 are independently determined by our observations (L , N , A , $G_{c,s}$, $B_{c,s}$, $E_{c,s}$). The remaining four terms that require scaling assumptions (C , F , $H_{c,s}$) do not contribute to the quantitative comparisons between the *in situ* tensor and the natural and laboratory petrofabrics.

3.3 Accounting for pyroxene in anisotropy calculations

Secondary phases other than olivine in a given sample act to reduce the bulk strength of seismic anisotropy (Mainprice & Silver, 1993; Bernard et al., 2021), yet the laboratory and natural petrofabrics used in this study consider aggregates of pure olivine. In order to directly compare these samples against our *in situ* estimate, which inherently includes bulk chemistry, we approximate the influence of secondary phases on seismic

anisotropy following Hansen et al. (2014). Mineral physics calculations using the tool *Perple_X* (Connolly, 2009) suggest the shallow lithospheric mantle is comprised of ~ 60 vol.% olivine assuming an upper mantle composition consistent with standard depleted mid-ocean ridge basalt (MORB) mantle composition (Hacker, 2008) and a half-space cooling temperature profile for 70 Ma (Figure S1). Invoking the simplifying assumption that the remaining 40% by volume can be approximated by orthopyroxene, a composite elastic tensor is constructed for each sample by taking the Voigt average between the olivine tensor and an orthopyroxene texture from Hansen et al. (2014) (Figure S2). This orthopyroxene texture comes from samples in which the orthopyroxene is dispersed in a dominantly olivine matrix, with the orthopyroxene comprising $< 3\%$ of the total volume. We do not account for more complex structures, such as pyroxene-rich bands, which may develop in realistic polyphase aggregates with higher modal fractions of pyroxene. This treatment of the influence of orthopyroxene acts to reduce the overall strength of the fabric without having a large effect on the fast propagation azimuth (Bernard et al., 2021).

For samples in which author-reported anisotropy strengths were used instead of being calculated from an elastic tensor, an empirical scaling was applied to account for the effect of pyroxene. Considering only well-developed fabrics with $\gamma > 2$ from the laboratory data of Hansen et al. (2014, 2016), fabric strength was calculated with orthopyroxene content ranging from 0% to 100% by volume for each sample and was fit with a linear function (Figure S3). The relationship between pyroxene content and reduction of anisotropy magnitude is nearly -1:1 and provides a straightforward method for scaling anisotropy magnitude.

4 Results

4.1 A comprehensive elastic model of oceanic lithosphere

Previously measured high-frequency ambient-noise Rayleigh- and Love-wave phase velocities were inverted for the complete elastic structure at NoMelt, incorporating co-located P_n constraints. The resulting model, NoMelt_SPani, is shown in Figure 2, and here we focus on the upper ~ 7 km beneath the Moho for which P- and S-constraints coincide. Azimuthal anisotropy increases with depth beneath the Moho for both G (V_{SV} anisotropy) and B (V_{PH} anisotropy), but remains relatively constant for E (V_{SH} anisotropy), largely due to the lack of Love-wave depth sensitivity (Russell et al., 2019). Anisotropy

fast azimuths Ψ_G and Ψ_B are sub-parallel to the fossil-spreading direction (FSD), and Ψ_E is rotated by 45° , as predicted for orthorhombic olivine (Montagner & Nataf, 1986). In detail, while Ψ_G and Ψ_B are each consistent with FSD within error, they differ from one another by $5\text{--}10^\circ$. This subtle mismatch is likely attributed to the different depth sensitivities of P_n and surface waves: the refraction imaging is primarily sensitive to the shallowest ~ 7 km of the mantle, while the surface waves integrate across the upper ~ 20 km.

From the 13 elastic parameters, we construct the equivalent orthorhombic elastic tensor at each depth and average the upper 7 km to produce a single representative elastic tensor, NoMelt_SPani7, given in Table 1. The tensor has been rotated into the seismic reference frame such that X is parallel to the FSD, Y is perpendicular to the FSD and parallel to Earth’s surface, and Z is perpendicular to Earth’s surface. The V_P pole figure shown in Figure 2c indirectly expresses the relative orientations of the three crystallographic axes. The well-defined maximum parallel to the FSD indicates a sub-horizontal, clustered [100] fast axis. The girdled slow and intermediate directions indicate dispersed [010] and [001] axes perpendicular to the inferred shear direction, characteristic of D-type olivine fabric (Figure 1b).

4.2 Quantification of *in situ* strain

Olivine LPO strength, and in turn the magnitude of seismic anisotropy, increases with shear strain (Hansen et al., 2014, 2016). The NoMelt model provides four independent estimates of the magnitude of seismic anisotropy and three independent estimates of anisotropic directions that can be directly compared to olivine LPO formed as a function of shear strain (Figure 3). We compare NoMelt_SPani7 to samples deformed in laboratory torsion experiments (Hansen et al., 2014, 2016) as well as to samples deformed naturally in a shear zone in the Josephine Peridotite, for which strain was determined measurable in outcrop (Warren et al., 2008; Hansen & Warren, 2015).

When 100% olivine is assumed, the high strains required to match the orientation of fast directions in Figure 3f–h correspond to magnitudes of anisotropy in Figure 3a–d that are approximately twice that which we observe. As the NoMelt model represents harzburgitic oceanic lithosphere, it is not surprising that a direct comparison to pure olivine fabrics fails to produce a range of shear strains compatible with the seismic model. Incorporating the effect of composition on the sample fabrics by accounting for the pres-

ence of pyroxene helps resolve this discrepancy. For each sample, we approximate the contributions from pyroxene by combining each pure olivine elastic tensor with the tensor for an appropriately oriented orthopyroxene texture following Hansen et al. (2014) as described in Section 3.3. We assume a nominal harzburgite composition of 60% olivine and 40% orthopyroxene by volume, which represents the lower bound of olivine content in abyssal peridotites observed globally (Warren, 2016). For the natural samples from Hansen & Warren (2015), dunite samples are used for the 100% olivine case, while harzburgite samples with 40% orthopyroxene added are used for the 60% olivine case. This composition is also consistent with the average lithospheric V_P and V_S in the NoMelt model, as compared to Perple_X calculations (Connolly, 2009) for a typical depleted mid-ocean ridge basalt (Figure S1). Compared to the pure olivine estimates, the mixture systematically decreases the strength of anisotropy due to the weaker single-crystal anisotropy and LPO of orthopyroxene compared to olivine (Mainprice & Silver, 1993), while fast azimuths are largely unaffected. This observation is consistent with the recent findings of Bernard et al. (2021), who investigated orthopyroxene effects on seismic anisotropy using a large suite of upper mantle xenoliths and found that the fast azimuth of anisotropy is not affected by orthopyroxene LPO for samples with >60% olivine.

The overall agreement between NoMelt anisotropy and the laboratory torsion data for 60% olivine is remarkable given the vast difference in length scale of the measurements (~ 9 orders of magnitude assuming seismic wavelengths on the order of ~ 100 km and laboratory LPO texture measurements from areas of $\sim 100 \times 100 \mu\text{m}^2$). The magnitude of anisotropy at NoMelt is consistent with laboratory samples for shear strains ranging from approximately 2 to 4 (minimum average misfit at $\gamma = 2.7$), and the orientation of fast directions is consistent with $\gamma > 3$ (Figure 4). We therefore infer that this portion of the upper mantle was subject to shear strains ranging from 2.5 to 4, which is on the upper end of that expected at shallow depths during seafloor spreading (Tommasi, 1998). The angle between the fast V_P axis and inferred shear plane is small ($\lesssim 10^\circ$) for $\gamma > 1.5$ (Figure 3e), implying a sub-horizontal concentration of [100] crystallographic axes (Skemer et al., 2012). Based on this observation we interpret a subhorizontal fabric at NoMelt, although we are unable to directly constrain fabric dip with our dataset that consists only of horizontally propagating waves.

The strain evolution of natural samples is more scattered and exhibits clear differences from the experimental samples. For a given strain and olivine content, the mag-

nitude of anisotropy of the natural fabrics is consistently weaker. This is perhaps un-
 surprising given the differences in scale between natural shear zones, such as at the Josephine
 ophiolite, and laboratory torsion experiments. In addition, the anisotropy fast directions
 of the natural samples fail to consistently rotate into the shear direction with increas-
 ing strain, remaining misaligned by $\sim 25^\circ$ at $\gamma = 5.25$ (Figure 3f–h). The relatively weak
 and misaligned anisotropy of the highly strained Josephine samples is likely attributed
 to the presence of a pre-existing LPO prior to shear zone development, which prolongs
 fabric development and misalignment of [100] with respect to the shear direction (War-
 ren et al., 2008; Skemer et al., 2012; Kumamoto et al., 2019). The natural samples also
 include contributions of secondary phases to LPO development, while the laboratory ex-
 periments on pure olivine do not. Uncertainties associated with measuring orientations
 in the field and preparing oriented samples for electron backscatter diffraction (EBSD)
 measurements in the lab may also contribute to scatter. NoMelt anisotropy is stronger
 than even the most highly strained natural sample ($\gamma = 5.25$) with 60% olivine, high-
 lighting the remarkably coherent LPO in the lithosphere across the $600 \times 400 \text{ km}^2$ NoMelt
 footprint (Russell et al., 2019). Any pre-existing vertical LPO associated with upwelling
 at the ridge was likely weak in comparison to that imposed after corner flow and there-
 fore overprinted.

Strain evolution of LPO from D-type at intermediate strains to A-type at high strains
 was previously identified by Hansen et al. (2014). This LPO evolution is reflected in the
 average V_P surfaces shown above Figure 3a (see supplementary Figure S6). The inferred
 shear strain at NoMelt of 2.5–4 based on strength and direction of anisotropy in com-
 parison to laboratory samples corresponds to fabrics with girdled slow and intermedi-
 ate V_P directions on average, consistent with a D-type LPO.

4.3 Distinguishing LPO fabric type

We further investigate LPO fabric type by comparing the relative magnitudes of
 azimuthal and radial anisotropy, which provides a discriminant of LPO type if the ori-
 entation of the sample with respect to the shear plane is known (Karato et al., 2008; Karato,
 2008) (Figure 1). The idea is similar to the more general Flinn diagram method of Michibayashi
 et al. (2016) based on ratios of V_P along different crystallographic directions and was re-
 cently proposed as a way to distinguish intrinsic LPO from extrinsic anisotropy mech-
 anisms such as compositional layering or oriented melt pockets (Hansen et al., 2021). In

Figure 5, we compare the strength of NoMelt anisotropy with that calculated for A-, D-, and E-type olivine fabrics from both natural settings (Peselnick & Nicolas, 1978; Ben-Ismaïl & Mainprice, 1998; Ben-Ismaïl et al., 2001; Mehl et al., 2003; Michibayashi et al., 2006; Warren et al., 2008; Karato, 2008; Skemer et al., 2010; Satsukawa et al., 2010) and laboratory deformation experiments (Zhang et al., 2000; Bystricky et al., 2000; Jung & Karato, 2001; Katayama et al., 2004; Jung et al., 2006). For most samples, the LPO type was characterized by the original authors.

To first order, samples with A-type LPO tend to exhibit strong radial anisotropy relative to azimuthal anisotropy, whereas the opposite is true for E-type. NoMelt displays moderate radial and azimuthal anisotropy most similar to the samples with D-type LPO. This result holds regardless of the olivine content assumed (Figure S4), though we find that a composition of 75% olivine and 25% orthopyroxene provides the best overall fit to the seismic observations. As the D-type LPO data in Figure 5 represent mostly natural samples, the higher olivine content required to match the seismic model is consistent with Figure 3, which showed that natural samples underestimate the magnitude of anisotropy for 60% olivine and 40% orthopyroxene. The true pyroxene content of the Josephine harzburgites ranges between 12.8–35.1% (Hansen & Warren, 2015), and accounting for this fact would bring the natural samples closer to the NoMelt observations of B/A , G/L , and E/N in Figure 3. A harzburgite composition with 75% olivine is also consistent with abyssal peridotites from fast-spreading MORs, which tend to have higher olivine content (70–95%) compared to slow-spreading MORs (Dick & Natland, 1996; Niu & Hékinian, 1997; Warren, 2016). If lower pyroxene content is assumed, the relationship in Figure 3a–d would indicate a moderately lower magnitude of strain at NoMelt that is inconsistent with the higher strains implied by the anisotropy fast orientations parallel to the shear direction Figure 3f–h.

Scatter among samples of the same fabric type in Figure 5 is likely attributed to differences in shear strain, with low strain samples exhibiting weaker anisotropy, as demonstrated in Figure 3a–d. At lower shear strains (and in the presence of a pre-existing LPO) dip of [100] with respect to the shear plane is common (Skemer et al., 2012) (Figure 3e), resulting in weaker radial and azimuthal anisotropy. Therefore, dipping fabrics in Figure 5 will tend to skew towards the origin, which does not change the overall inference of D-type LPO (see supplementary Figure S5). In other cases, scatter may be attributed to samples that straddle the boundary between two LPO types, such as the peridotite

averages of Ben-Ismail & Mainprice (1998), which included both A-type and D-type samples.

5 Discussion

This study presents, to our knowledge, the first *in situ* quantification of the orthorhombic elastic tensor of oceanic lithosphere that incorporates both compressional- and shear-anisotropy observations at a single location. The model, NoMelt_SPani7, represents typical oceanic lithosphere produced by corner flow at a fast-spreading MOR, providing a benchmark for LPO formation during seafloor spreading. The relatively simple corner-flow geometry followed by strain accumulation with horizontal flow lines allows for straightforward comparison with deformed olivine petrofabrics from both natural and laboratory settings. Under the assumption of horizontal shear, we are able to infer the accumulation of shear strain in the upper 7 km of the mantle to be of order 250–400% (Figure 3), with [010] and [001] girdled perpendicular to the shear direction, characteristic of D-type LPO (Figure 5).

5.1 Limitations of the seismic model

While this study provides an exceptionally complete *in situ* elastic model of oceanic lithosphere, several limitations exist. First, we are unable to explicitly solve for fabric dip because only horizontally propagating waves (surface waves and P_n) are used. Theoretically, it is possible to utilize P and S body waves with a range of incidence angles to constrain fabric dip via the zero-valued off-diagonal elements of C_{ij} (eq 8). In practice however, this is extremely challenging, requiring earthquake observations from a range of epicentral distances and back-azimuths that have been effectively corrected for near-source and deep-earth 3-D structure outside the region of interest. Our use of laboratory data to indirectly infer fabric dip based on the observation of fast azimuth alignment with the FSD suggests fabric is horizontal to within $\sim 10^\circ$ in our study region, providing confidence in our current approach.

A consequence of D-type LPO is the inability to uniquely determine the orientation of the shear plane using seismologically determined crystallographic orientations alone. Fortunately, the assumption of a horizontal shear plane is valid for many plate-tectonic processes including MOR spreading once far enough from the ridge. Our ability to com-

pare the seismological model with natural and laboratory data in a relatively straight-forward way is largely a result of the simple geometry. For more complex tectonic environments such as subduction zones or directly beneath MORs, the orientation of the shear plane may be significantly rotated from horizontal, and this simplified approach breaks down.

As our 1-D model represents an average over the 600×400 km² NoMelt footprint, a variation in LPO type across the region cannot be completely ruled out. For instance, a spatial variation from A-type to E-type could appear as D-type LPO when averaged over the array. Alternatively, averaging over domains with A-type LPO that have a common shear direction but orthogonal orientations of the shear planes could produce an apparent D-type LPO. This could be the case, for example, if averaging over oceanic lithosphere that includes transform faults and fracture zones, which could incorporate transitions between horizontal and vertical shear planes. However, NoMelt is situated between the Clarion and Clipperton Fracture Zones on lithosphere formed at a single MOR segment, and therefore the model captures only normal seafloor-spreading fabric. In addition, lateral variations in surface-wave phase velocities imaged at NoMelt are small ($<1\%$) (Russell et al., 2019), consistent with a homogeneous LPO type within the footprint.

5.2 Strain accumulation in the shallow lithosphere

Our estimate of 250–400% strain accumulated in the upper 7 km of the mantle based on observed seismic anisotropy is, as far as we are aware, the first of its type. The primary assumption underlying this estimate is that the strain evolution of olivine LPO – and in turn seismic anisotropy – observed in the laboratory is the same as that which occurs in the mantle. This necessary assumption could be an oversimplification. The influence of secondary phases, such as pyroxene, on the development of olivine LPO has not been well quantified in laboratory experiments and may depend on the proportion of the secondary phase relative to olivine. In addition, if a pre-existing fabric forms during upwelling at the MOR, LPO development associated with corner flow would be delayed, and strain estimates based solely on seismic anisotropy would be underestimated (Skemer et al., 2010). Therefore, we consider the range 250–400% to be a lower bound on strain accumulated in the shallow lithosphere. Because the strength of LPO and seismic anisotropy saturate at higher strains (Figure 3) (Tommasi, 1998; Tommasi & Vauchez, 2015), seismic anisotropy is limited in its ability to quantify strains greater than $\sim 400\%$.

5.3 Interpretation of D-type fabric

The relative strengths of radial and azimuthal anisotropy in the shallow lithosphere at NoMelt are consistent with that of D-type peridotites (Figure 5). As NoMelt is thought to represent typical oceanic plate, this observation suggests that D-type LPO is widespread in oceanic lithosphere. This hypothesis is supported by numerous studies that document D-type LPO (in addition to A-type) in oceanic peridotites (e.g. Ben-Ismaïl & Mainprice, 1998; Tommasi et al., 2004; Michibayashi et al., 2006; Vonlanthen et al., 2006; Warren et al., 2008; Michibayashi et al., 2016; Liu et al., 2019; Tommasi et al., 2020). Given our result, we explore utilizing this constraint on fabric type to improve our understanding of mantle deformation during seafloor spreading.

It has been well established from laboratory experiments (e.g. Karato et al., 1980; Karato & Wu, 1993; Zhang & Karato, 1995), direct sampling (e.g. Peselnick & Nicolas, 1978; Ben-Ismaïl & Mainprice, 1998; Tommasi et al., 2004), and seismic observations (e.g. Hess, 1964; Forsyth, 1975; Nishimura & Forsyth, 1989) that the olivine-rich upper mantle deforms via dislocation creep. This led to the common assumption that A-type LPO pervades the upper mantle, as it results from dislocation creep with slip dominantly occurring on the easiest olivine slip system, (010)[100]. In contrast, D-type LPO results from slip on both easy slip systems (010)[100] and (001)[100] (e.g. Bystricky et al., 2000; Hansen et al., 2014). Single-crystal experiments demonstrate that critical resolved shear stresses for (001)[100] are similar to (010)[100] for comparable strain rates and temperatures $<1300^{\circ}\text{C}$ (Bai et al., 1991).

Similar amounts of slip on both (001)[100] and (010)[100] are favored if the von Mises strain compatibility criterion is relaxed, minimizing activity of the hardest slip system, (010)[001] (Tommasi et al., 2000; Castelnau et al., 2008; Signorelli & Tommasi, 2015). Several grain-scale mechanisms have been proposed for relaxing strain compatibility in the oceanic lithosphere. On the one hand, numerical modeling suggests that subgrain rotation recrystallization can relax the requirement for strain compatibility by increasing intragranular strain heterogeneity (Signorelli & Tommasi, 2015). On the other hand, laboratory studies indicate that von Mises criterion can be relaxed by grain boundary sliding during dislocation creep (Hirth & Kohlstedt, 2003; Hansen et al., 2011). Based on the observed grain-size dependence, dislocation-accommodated grain-boundary sliding (disGBS) has been observed to be the dominant dislocation creep mechanism dur-

ing experiments on olivine aggregates over a wide range of stresses and grain sizes (Hansen et al., 2011; Bollinger et al., 2019), including in higher temperature experiments that produced D-type fabric (e.g. Bystricky et al., 2000; Hansen et al., 2011). We emphasize that a significant amount of strain — and accompanying grain rotations that lead to LPO — is accommodated by dislocation glide on the easy slip systems in this dislocation creep regime, even when processes at the grain boundaries are rate limiting (e.g., see Discussion in Ferreira et al., 2021).

DisGBS has been previously invoked to explain the occurrence of D-type LPO in oceanic peridotites (Braun, 2004; Warren et al., 2008). In ophiolites, A- and D-type LPOs have been observed in coarse-grained dunites and finer-grained harzburgites, respectively, suggesting pyroxene may play a role in D-type LPO formation by limiting olivine grain size and promoting deformation by disGBS (Warren et al., 2008). Indeed, contributions from pyroxene are important for reconciling the NoMelt observation with sample data more broadly (Figures 3,5).

5.3.1 *Implications of the disGBS hypothesis*

We examine the hypothesis that deformation via disGBS during MOR spreading produced the D-type LPO that we observe in the oceanic lithosphere. Olivine flow laws derived from recent laboratory experiments suggest that the disGBS process dominates at upper-mantle conditions (Figure 6a). Similarly, MOR geodynamic modeling that incorporates grain-size evolution and multiple deformation processes (diffusion creep, dislocation creep, and disGBS) indicates that shallow, near-ridge deformation is dominated by disGBS in some cases (Turner et al., 2015). These results have potentially important rheological implications, as disGBS deformation is sensitive to grain size, while dislocation creep without grain-boundary sliding is insensitive to grain size (Hirth & Kohlstedt, 2003; Hansen et al., 2011). Here, we explore the implications for upper mantle grain size and stress.

Olivine flow laws at a given temperature and pressure describe the strain-rate contributions of grain-size insensitive dislocation creep, disGBS, and diffusion creep processes as a function of stress and grain size. A deformation mechanism map is presented in Figure 6a for reasonable sub-Moho conditions at the MOR ($T = 1250^\circ\text{C}$, $P = 360\text{ MPa}$), using olivine flow law parameters from Hansen et al. (2011) (Table S3). For a given stress,

grain-size insensitive dislocation creep dominates at large grain sizes, and diffusion creep occurs at small grain sizes. The disGBS regime occupies the transition between the two and is distinguished by its modest grain-size dependence.

Assuming tectonic strain rates of $10^{-14.5}$ – $10^{-12.5}$ s⁻¹, deformation via disGBS implies *in situ* grain sizes of 0.3–15 mm and relatively low stress (0.2–1.4 MPa), broadly consistent with recent geodynamic modeling at a MOR (Turner et al., 2015, 2017) as well as olivine grain sizes observed in peridotites (e.g. Vonlanthen et al., 2006; Tommasi et al., 2020). However, these grain sizes are on the smaller end of the steady-state predictions of ~5–50 mm from the laboratory-calibrated grain-size piezometers (Karato et al., 1980; Hirth & Kohlstedt, 2015) and wattmeter (Austin & Evans, 2007) shown in Figure 6a (see Appendix Appendix B for details of wattmeter calculation).

Relatively small experimental uncertainties result in large uncertainties when extrapolated to mantle conditions. Considering the uncertainty in the flow law parameters (see Appendix Appendix C for details), scenarios are possible in which the disGBS field expands to larger grain sizes coinciding with the piezometer and wattmeter predictions at mantle conditions while still satisfying the laboratory data (Figure 6b). Figure C1 shows the full range of possible flow law parameter values (within experimental uncertainties) for which disGBS is predicted to be the dominant deformation mechanism at the stress/grain size values estimated by extrapolation of the grain size piezometer/wattmeter. If disGBS is responsible for producing the D-type LPO we observe and the extrapolation of the piezometer/wattmeter relationships are applicable, this analysis suggests that the nominal olivine flow law of Hansen et al. (2011) underpredicts the contribution of disGBS at mantle conditions, with a preference for smaller disGBS stress and grain-size exponents n_{GBS} and m_{GBS} , respectively, and/or a larger dislocation creep stress exponent n_{DISL} relative to their nominal reference values (Table S3).

Alternatively, it is possible that the flow law parameters are skewed in the opposite sense, such that disGBS does not occur at mantle conditions (Figure 6c) and deformation occurs purely in the dislocation creep regime without a significant contribution from grain-boundary sliding. This scenario would require an alternative mechanism for producing the evidence for D-type LPO we observe, such as dynamic recrystallization (Signorelli & Tommasi, 2015). Another possibility is that the nominal flow law parameters in Figure 6a are valid, but mantle grain sizes are somewhat smaller than predicted

by the wattmeter and piezometer. Indeed, olivine grain sizes in the range of 0.5–5 mm are commonly observed in mantle peridotites (e.g. Vonlanthen et al., 2006; Michibayashi et al., 2006; Liu et al., 2019; Tommasi et al., 2020). These grain sizes observed in nature sit squarely in our predicted disGBS field and may be due to grain-boundary pinning in the presence of secondary phases such as pyroxene, which is not captured by predictions based on results from olivine experiments. New experiments that place tighter bounds on olivine flow law parameters and quantify contributions of pyroxene are required to resolve these ambiguities.

This exercise is not meant to invalidate the olivine flow laws; rather, it 1) highlights the uncertainties associated with extrapolating to mantle conditions across nearly two orders of magnitude in stress and grain size and 2) demonstrates how *in situ* observations of LPO type may be utilized to better constrain such extrapolations.

5.4 Mid-ocean ridge deformation

The strong correspondence between observed anisotropic fast directions and the direction of fossil seafloor spreading at NoMelt implies that the *in situ* elastic tensor represents mantle deformed by corner flow during spreading. Our inferences of girdled D-type LPO, high strain accumulation ($\gamma = 2.5\text{--}4$), and sub-horizontal [100] orientations will inform new geodynamic modeling efforts of MOR dynamics and associated fabric development. Recent models often contain relatively weak anisotropy in the shallow lithosphere with a dipping [100] orientation and A-type LPO (e.g. Blackman et al., 1996, 2017), in contrast to what we infer at NoMelt. Our observation of D-type LPO, implying equal amounts of slip on both (010)[100] and (001)[100] during MOR spreading, provides a new target for future geodynamic models. More girdled fabrics have been previously produced in viscoplastic self-consistent models by relaxing strain compatibility constraints (Tommasi et al., 2000), but generally such alternative LPO types are not explored in geodynamic models.

Although the precise grain-scale mechanism(s) involved is still uncertain, D-type LPO is consistent with shallow deformation via grain-size sensitive disGBS. This would imply a complex non-Newtonian and grain-size dependent rheology that impacts solid-state convection streamlines, melt extraction, and thermal and rheological evolution of the lithosphere and asthenosphere (Turner et al., 2017). Grain size, in particular, has

been shown to play an important role in melt focusing near MORs through alteration of the viscosity structure resulting in gradients in compaction pressure (Turner et al., 2015, 2017). Smaller grain sizes in the shallow mantle required for disGBS ($\sim 1\text{--}10$ mm) would decrease porosity and permeability, acting as a permeability barrier along which melt may be focused towards the ridge axis. Although our results are only sensitive to the shallow lithospheric mantle, significantly shallower than where melt is stable off-axis, geodynamic models that include non-Newtonian viscosities predict that disGBS also dominates in the shallow asthenospheric mantle (Turner et al., 2015).

5.5 Importance of pyroxene

The influence of pyroxene on bulk anisotropy, LPO development, and mantle rheology is still under investigation. A recent study of natural peridotites containing various olivine and orthopyroxene LPO types shows that orthopyroxene has no significant effect on the fast azimuth of seismic anisotropy when olivine composes at least 60% of the sample (Bernard et al., 2021). However, the strength of anisotropy does depend on both the proportion of pyroxene and the magnitude of strain (Figure 3a–d). In practice, it is not possible to separate these two effects when composition is unknown, and one must assume a realistic mantle composition based on prior knowledge as we have done. Additionally, shear strain is not readily determined in natural samples except in special cases for which strain markers are present, such as at the Josephine shear zone (Warren et al., 2008; Hansen & Warren, 2015), and those samples by definition contain a pre-existing LPO, which also impacts strain development. Untangling these competing effects on anisotropy will require peridotite elastic tensors that account for the true modal compositions and LPO for all main phases within the sample (e.g. Tommasi et al., 2004; Vonlanthen et al., 2006; Liu et al., 2019; Tommasi et al., 2020). In addition, high-strain experiments are needed that examine the effect of olivine and pyroxene together and that quantify the influence of pre-existing fabric on LPO development using realistic aggregate compositions.

6 Conclusions

We develop an *in situ* elastic tensor for ~ 70 Ma oceanic lithosphere constrained by co-located shear and compressional seismic observations sensitive to the upper ~ 7 km of the mantle. By comparing our model to a suite of previously published tensors from

591 olivine aggregates deformed in nature and in the laboratory, we provide new constraints
 592 on LPO type and magnitude of shear strain accumulated in oceanic lithosphere. The rel-
 593 atively strong azimuthal anisotropy with fast azimuth parallel to the spreading direc-
 594 tion is consistent with shear strains of 250–400% and sub-horizontal fabric produced by
 595 corner flow near the MOR. The relative strengths of azimuthal and radial anisotropy cor-
 596 respond to D-type LPO ([100] parallel to the shear direction and girdled [001] and [010]
 597 perpendicular to shear) irrespective of assumed olivine content or fabric dip. Assuming
 598 NoMelt represents the average oceanic plate, this result implies that D-type LPO is more
 599 ubiquitous in the oceanic lithosphere than typically thought. Our observations can be
 600 explained by a mantle composition of 60–75% olivine, consistent with observed mantle
 601 peridotites, but the competing effects of composition, strain, and pre-existing LPO on
 602 anisotropy strength cannot be completely isolated.

603 We hypothesize that D-type LPO in the shallow lithosphere results from near-ridge
 604 deformation dominated by disGBS that relaxes strain compatibility requirements and
 605 produces equal amounts of slip on both easy slip systems. Under this assumption, our
 606 observation implies *in situ* grain sizes of 0.3–15 mm, consistent with observed peridotites,
 607 and stresses on the order of 0.2–1.4 MPa. The grain-size sensitivity of disGBS deforma-
 608 tion impacts rheology of the spreading mantle and subsequent LPO development. This
 609 study demonstrates the ability to integrate seismological- and laboratory-scale observa-
 610 tions of seismic anisotropy to improve understanding of MOR dynamics and mantle de-
 611 formation during seafloor spreading.

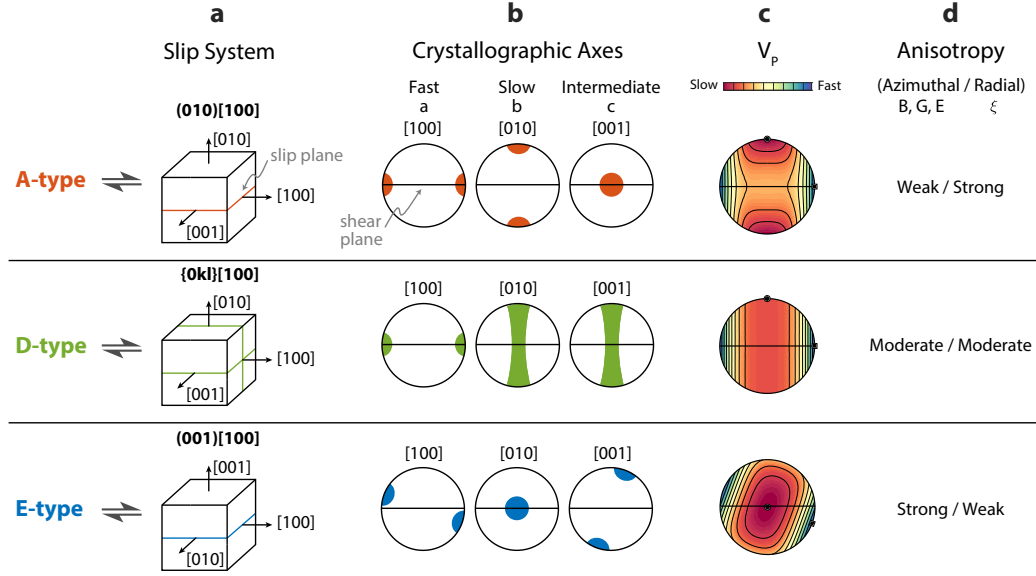


Figure 1. Schematic summary of the most commonly observed LPO types. (a) slip systems, (b) crystallographic orientations, and (c) compressional velocities, V_p , expected for right lateral shear (top to the right shear sense). Note the qualitative differences in V_p between the three fabric types arising from the relative orientations of $[100]$, $[010]$, and $[001]$. (d) Relative strengths of radial and azimuthal anisotropy for A-, D-, and E-type fabrics under the assumption that the shear plane is parallel to Earth's surface. Differences in the orientation of slow $[010]$ and intermediate $[001]$ axes result in measurable differences in seismic anisotropy. Azimuthal anisotropy is sensitive to the two crystallographic axes in the horizontal plane, and radial anisotropy is sensitive to the difference between the azimuthally averaged horizontal axes and the vertically oriented axis. The depiction of a slight rotation of $[100]$ from the shear plane for E-type LPO is based on laboratory observations (Katayama et al., 2004) and natural samples (Skemer et al., 2010).

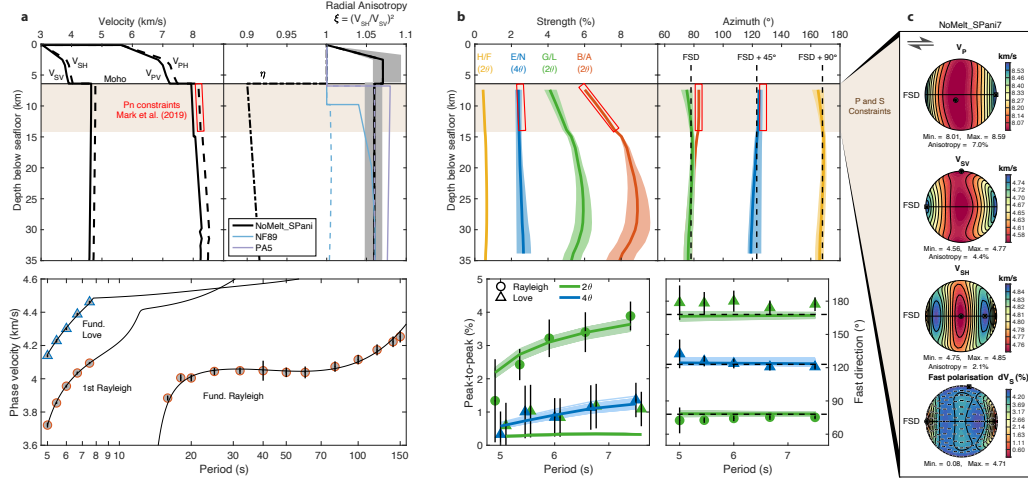


Figure 2. Comprehensive elastic model and data fit for the NoMelt region. Summary of the NoMelt.SPani model. (a), V_P , V_S , and Radial anisotropy model and data fit after (Russell et al., 2019). Grey shading for NoMelt radial anisotropy captures the range of models that fit the data, with the best fit model shown in black. Brown shading indicates where both V_P (Mark et al., 2019) and V_S constraints exist in the upper ~7 km of the mantle. NF89, (Nishimura & Forsyth, 1989) 52–110 Ma region; PA5, (Gaherty et al., 1996). (b), Azimuthal anisotropy strength and azimuth. Model uncertainties of two standard deviations from bootstrapping are shaded. Data fit and range of bootstrap model fits are shown in the lower panels. (c), Velocity calculations for the average elastic tensor for the upper ~7 km of the lithospheric mantle (NoMelt.SPani7), plotted using the MATLAB Seismic Anisotropy Toolkit (Walker & Wookey, 2012). The horizontal black line denotes the horizontal plane with the fossil-spreading direction (FSD) to the east and west. Maxima and minima are denoted by a square and circle, respectively. $dV_s = 200(V_{SH} - V_{SV})/(V_{SH} + V_{SV})$

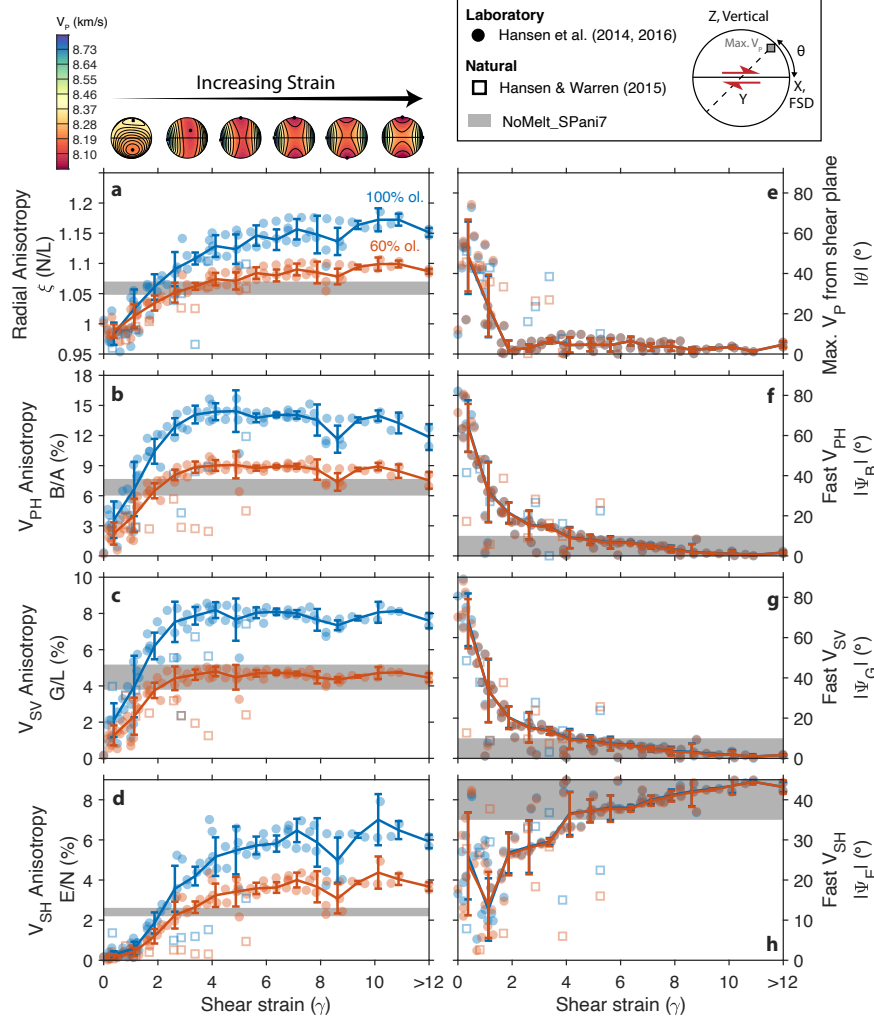


Figure 3. NoMelt anisotropy compared with olivine petrofabric data as a function of shear strain and pyroxene content. Anisotropy strength (a–d) and fast azimuth with respect to the shear direction or FSD (f–h) for laboratory (solid circles) (Hansen et al., 2014, 2016) and natural samples (open squares) (Hansen & Warren, 2015). The NoMelt seismic model for the upper ~ 7 km of the lithospheric mantle, NoMelt_SPani7, is shown in grey with a width that represents the full range of values in that depth range. NoMelt FSD uncertainty is $\sim 10^\circ$. **e** Magnitude of the angle between maximum V_P and the shear plane; seismic model does not constrain this parameter. Calculations for (blue) pure olivine and (red) 60% olivine, 40% orthopyroxene are shown (see Methods for details on the inclusion of orthopyroxene). Error bars show median and standard deviation for laboratory data binned by strain with bin width $\gamma = 0.75$. V_P surfaces are shown above **a** for laboratory samples averaged by increments of $\gamma = 2$ (See also Figure S6). All samples oriented consistent with the seismic reference frame: shear plane parallel to Earth’s surface and shear direction parallel to the FSD.

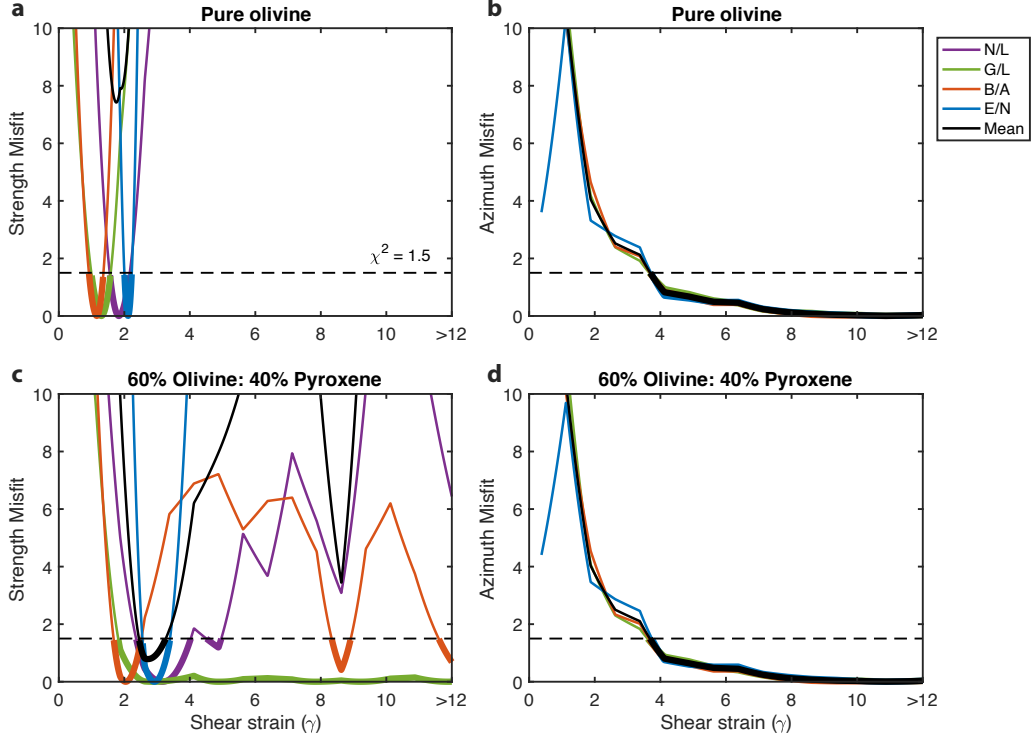
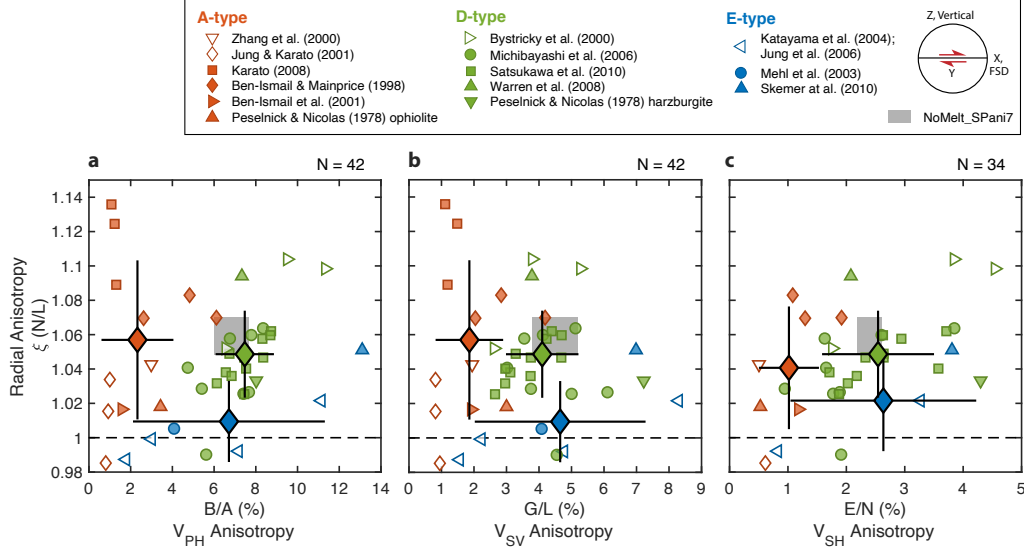


Figure 4. Misfit of anisotropy strength and fast azimuth with shear strain. (a) Misfit between the strength of anisotropy of NoMelt_SPani7 observations and experimental data from Figure 3, assuming 100% olivine. Line color indicates anisotropy type, where black is the average misfit value. Thick lines indicate values of misfit that are below the misfit threshold of $\chi^2 = 1.5$ (dashed line), meaning that the seismic observation and laboratory data agree. (b) same as (a) but for anisotropy fast azimuth. (c,d) same as (a,b) but for 60% olivine and 40% pyroxene. Misfit is calculated as $\chi^2 = (\bar{d}_{\text{obs}} - \bar{d}_{\text{exp}})^2 / \sigma_{\text{obs}}^2$, where \bar{d}_{obs} is the median observed value at NoMelt, \bar{d}_{exp} is the median experimental value, and σ_{obs} is the uncertainty on the observed value.



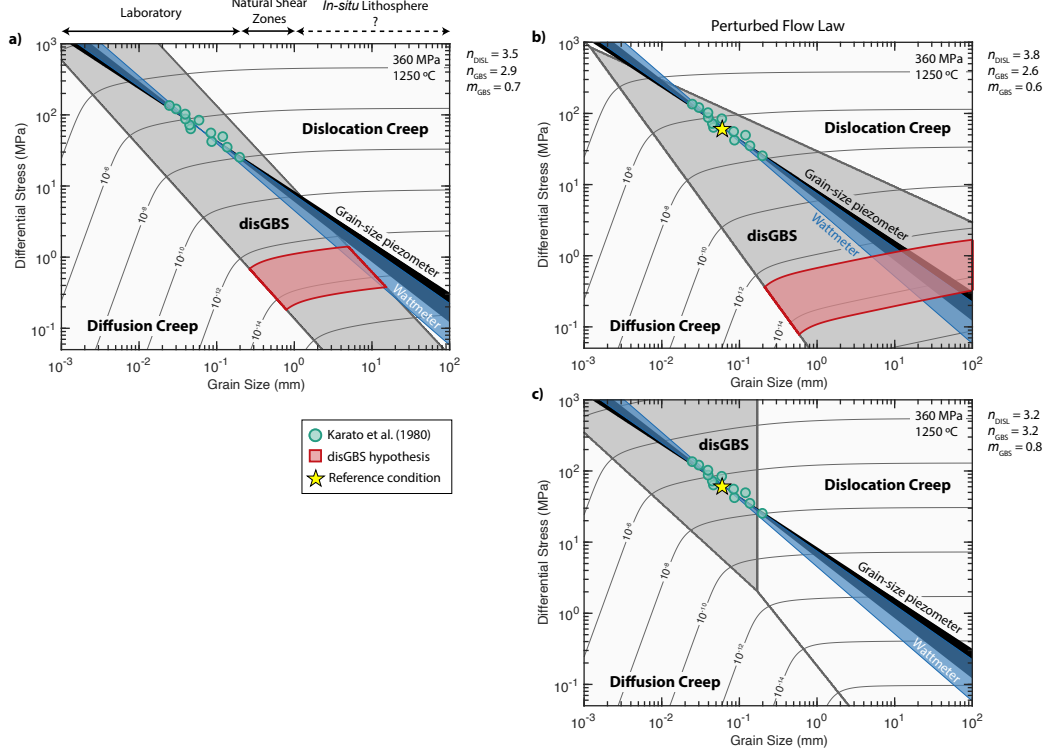


Figure 6. Evaluation of the disGBS hypothesis for producing D-type LPO at mid-ocean ridge conditions. (a) Strain rate contours are calculated using the nominal olivine flow laws of Hansen et al. (2011) at a temperature of 1250°C and pressure of 360 MPa (see Table S3), approximating shallow mantle conditions at the MOR. The red region bounded by strain rates of $10^{-14.5}$ – $10^{-12.5}$ s $^{-1}$ and within the grey dislocation-accomodated grain boundary sliding (disGBS) regime represents the range of stress and grain sizes predicted by the disGBS hypothesis to produce D-type fabric (e.g. Braun, 2004; Warren et al., 2008). The grain-size wattmeter of Austin & Evans (2007) is shown in blue and includes uncertainties in laboratory derived stress and grain-size exponents (see Methods for details). A commonly used grain-size piezometer (Karato et al., 1980; Hirth & Kohlstedt, 2015) is shown in black. Laboratory data of Karato et al. (1980) are shown in teal. (b–c) same as a) but for perturbed flow law parameters n_{DISL} , n_{GBS} , and m_{GBS} within their uncertainty bounds, representing endmember cases. The star marks the reference state ($d = 60$ μm , $\sigma = 60$ MPa) used to ensure agreement with laboratory data (See methods for details). Conditions in (b) support the disGBS hypothesis, while (c) does not.

Table 1. NoMelt_SPani7 elastic tensor, C_{ij} , averaged over the upper ~ 7 km of the mantle. The tensor is oriented in the seismological reference frame with X parallel to the fossil-spreading direction, Y parallel to Earth’s surface and perpendicular to fossil spreading, and Z perpendicular to Earth’s surface.

i	j	1	2	3	4	5	6
1		243.5427	71.5726	73.6646	0	0	1.5311
2		–	212.0492	74.4322	0	0	1.2569
3		–	–	213.2303	0	0	0.0068
4		–	–	–	68.5948	-0.0441	0
5		–	–	–	–	75.3675	0
6		–	–	–	–	–	74.4751

Appendix A Elastic parameter definitions

The five Love parameters A , C , L , N , and F (Dziewonski & Anderson, 1981; Anderson & Dziewonski, 1982) that describe a transversely isotropic medium can be written in terms of the elastic tensor C_{ij} as follows (Montagner & Nataf, 1986):

$$A = \rho V_{\text{PH}}^2 = \frac{3}{8}(C_{11} + C_{22}) + \frac{1}{4}C_{12} + \frac{1}{2}C_{66}, \quad (\text{A1})$$

$$C = \rho V_{\text{PV}}^2 = C_{33}, \quad (\text{A2})$$

$$L = \rho V_{\text{SV}}^2 = \frac{1}{2}(C_{55} + C_{44}), \quad (\text{A3})$$

$$N = \rho V_{\text{SH}}^2 = \frac{1}{8}(C_{11} + C_{22}) - \frac{1}{4}C_{12} + \frac{1}{2}C_{66}, \quad (\text{A4})$$

$$F = \rho\eta(V_{\text{PH}}^2 - 2V_{\text{SV}}^2) = \frac{1}{2}(C_{13} + C_{23}), \quad (\text{A5})$$

$$\xi = \left(\frac{V_{\text{SH}}}{V_{\text{SV}}}\right)^2 = \frac{N}{L}, \quad (\text{A6})$$

where radial anisotropy, ξ , is a proxy for vertical ($\xi < 1$) or horizontal ($\xi > 1$) flow in the mantle.

In the case of a medium with orthorhombic symmetry, each elastic moduli varies about its average value according to a sinusoid with magnitude (G , B , H , and E) and corresponding azimuth (Ψ_G , Ψ_B , Ψ_H , and Ψ_E) of anisotropy:

2 θ :

$$G = \delta L = \sqrt{G_c^2 + G_s^2}, \quad \Psi_G = \frac{1}{2} \arctan\left(\frac{G_s}{G_c}\right); \quad (\text{A7})$$

$$B = \delta A = \sqrt{B_c^2 + B_s^2}, \quad \Psi_B = \frac{1}{2} \arctan\left(\frac{B_s}{B_c}\right); \quad (\text{A8})$$

$$H = \delta F = \sqrt{H_c^2 + H_s^2}, \quad \Psi_H = \frac{1}{2} \arctan\left(\frac{H_s}{H_c}\right); \quad (\text{A9})$$

4 θ :

$$E = \delta N = \sqrt{E_c^2 + E_s^2}, \quad \Psi_E = \frac{1}{4} \arctan\left(\frac{-E_s}{-E_c}\right); \quad (\text{A10})$$

621

$$G_c = \frac{1}{2}(C_{55} - C_{44}), \quad (\text{A11})$$

$$G_s = C_{54}, \quad (\text{A12})$$

$$B_c = \frac{1}{2}(C_{11} - C_{22}), \quad (\text{A13})$$

$$B_s = C_{16} + C_{26}, \quad (\text{A14})$$

$$H_c = \frac{1}{2}(C_{13} - C_{23}), \quad (\text{A15})$$

$$H_s = C_{36}, \quad (\text{A16})$$

$$E_c = \frac{1}{8}(C_{11} + C_{22}) - \frac{1}{4}C_{12} - \frac{1}{2}C_{66}, \quad (\text{A17})$$

$$E_s = \frac{1}{2}(C_{16} - C_{26}), \quad (\text{A18})$$

622 The magnitude and direction of azimuthal anisotropy depend on the bulk orientation
623 and degree of organization of olivine [100] within the horizontal plane.

624 Appendix B Wattmeter steady-state grain size calculation

625 Grain-size evolution as a function of stress (σ) is estimated following the wattmeter
626 approach of Austin & Evans (2007):

$$\dot{d} = \frac{K_g}{p} d^{1-p} \exp\left(-\frac{E_g + \bar{P}V_g}{RT}\right) - \frac{2\chi}{c\gamma} d^2 \sigma : \dot{\epsilon}, \quad (\text{B1})$$

627 where K_g and E_g are the grain growth prefactor and activation energy, respectively.
628 The first term on the right-hand side represents grain growth by material diffusion be-
629 tween grains, and the second term represents grain-size reduction through dynamic re-
630 crystallization, which depends on χ the fraction of work done by dislocation mechanisms
631 in order to change grain boundary area, average specific grain boundary energy, γ , and
632 a geometric constant, c . We follow (Behn et al., 2009) and assume a grain growth ex-
633 ponent of $p = 3$. Assuming deformation occurs via disGBS, $\dot{\epsilon} = A_{\text{GBS}}\sigma^n d^{-m} \exp(-(E_{\text{GBS}} + PV_{\text{GBS}})/RT)$
634 and setting eq (B1) equal to zero, the steady-state grain size is given by:

$$d_{\text{ss}} = \left[\frac{K_g c \gamma}{2\chi p A_{\text{GBS}} \sigma^{n+1}} \exp\left(\frac{(E_{\text{GBS}} - E_g) + (PV_{\text{GBS}} - PV_g)}{RT}\right) \right]^{\frac{1}{1+p-m}} \quad (\text{B2})$$

635 All parameter values can be found in Table S3. The resulting steady-state grain-
636 size variation with stress, or wattmeter, at $T = 1250^\circ\text{C}$ is shown in Figure 6.

Appendix C Flow-law uncertainties

Uncertainty in grain-size and stress exponents in olivine flow laws affect both the wattmeter calculation as well as deformation mechanism maps when extrapolated from laboratory to mantle conditions. We evaluate the effect of experimentally determined uncertainties in m_{GBS} , n_{GBS} , and n_{DISL} on the wattmeter and deformation mechanism maps by perturbing them within their uncertainty bounds ($n \pm 0.3$ and $m \pm 0.1$ relative to the reference values listed in Table S3). In order to remain consistent with the laboratory data, we correspondingly solve for prefactors A_{GBS} and A_{DISL} using the strain rate evaluated at a reference state of $d = 60 \mu\text{m}$ and $\sigma = 60 \text{ MPa}$, representative of average laboratory conditions. The deformation mechanism map and wattmeter are then calculated using the new value of A and perturbed values of n and m . The resulting effects on the deformation mechanism map, and the disGBS region in particular, are shown for two endmember cases in Figure 6b–c. Because we calculate the deformation mechanism maps and the grain size versus stress relationships at a temperature of 1250°C (which is within the temperature range that the grain-growth and deformation experiments were conducted), uncertainties in the activation energies for the flow laws and grain-growth law (Speciale et al., 2020) do not significantly impact our analysis. This flow-law exploration was implemented using the Very Broadband Rheology calculator (Havlin et al., 2021).

We perform a grid search over values of m_{GBS} , n_{GBS} , and n_{DISL} within their uncertainty bounds to determine the range of parameters that satisfy our *in situ* inference of disGBS as well as the laboratory derived grain-size piezometer and wattmeter (Figure C1). A set of parameters is considered acceptable if disGBS is predicted to occur at grain sizes at least as large as those predicted from the piezometer and wattmeter in Figure 6 (i.e., when the piezometer and wattmeter fall completely within the disGBS regime) at mantle strain rates of $10^{-14.5}$ – $10^{-12.5} \text{ s}^{-1}$. Figure C1 shows that this criterion is met when m_{GBS} and n_{GBS} are smaller than their reference values and/or n_{DISL} is larger than its reference value. Furthermore, there is a clear tradeoff between the disGBS parameters (m_{GBS} , n_{GBS}) and the dislocation creep parameter n_{DISL} .

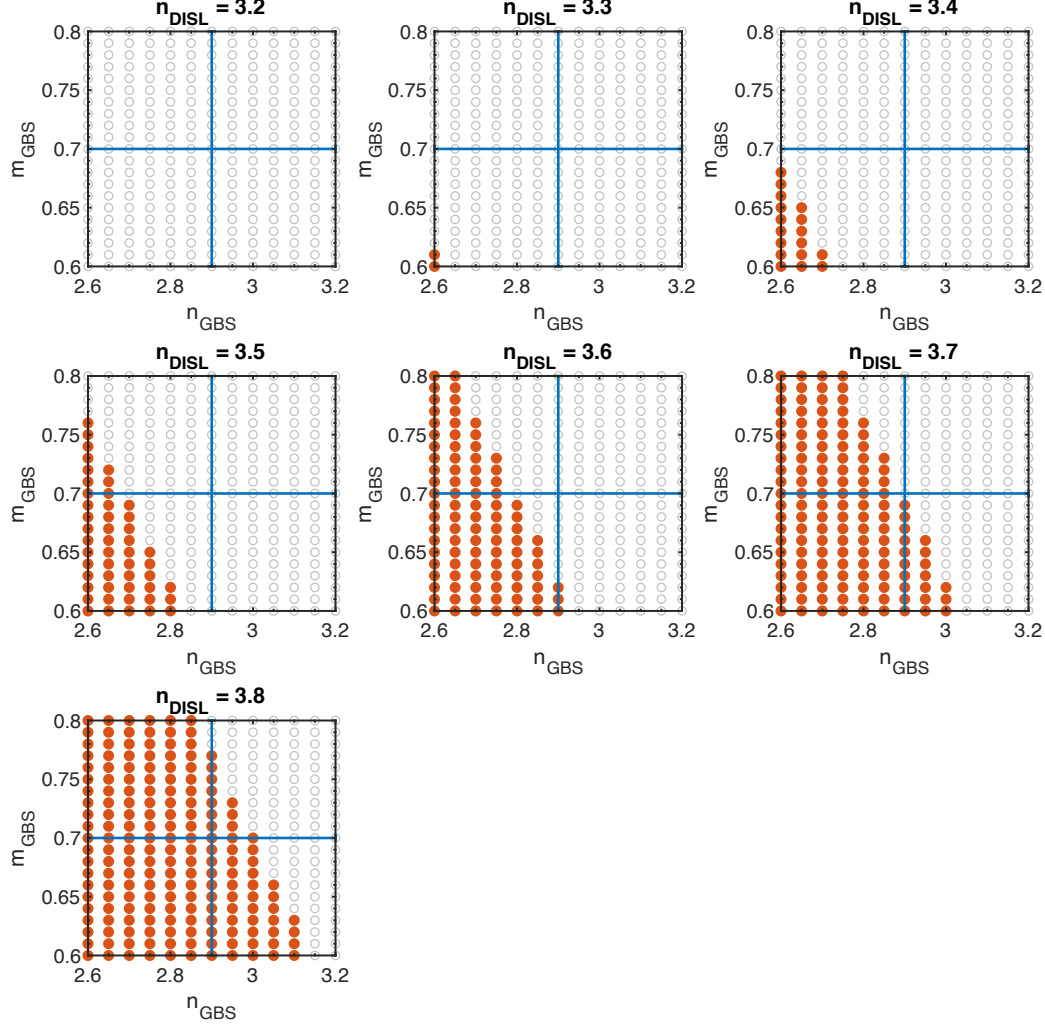


Figure C1. Flow law parameter combinations which predict disGBS at mantle strain rates. A grid search is performed over values of n_{DISL} , n_{GBS} , and m_{GBS} within their uncertainty bounds in order to determine the combinations for which disGBS is predicted to occur at grain sizes corresponding to the piezometer and wattmeter shown in Figure 6 at mantle strain rates. A combination of parameters is considered acceptable when the grain-size piezometer and wattmeter are completely within the disGBS region of the deformation mechanism map for strain rates of $10^{-14.5}$ – $10^{-12.5} \text{ s}^{-1}$. Filled red circles indicate parameter combinations for which this criterion is met, while open circles failed to meet this criterion. The vertical and horizontal blue lines indicate the reference values of n_{GBS} and m_{GBS} , respectively from Table S3. The reference value of n_{DISL} is 3.5. We find that smaller values of n_{GBS} and m_{GBS} and/or larger values of n_{DISL} relative to their reference values are compatible with the disGBS hypothesis.

Open Research

Seismic data was accessed through the IRIS Data Management Center under network code ZA using the open-source Python package ObsPy (<https://docs.obspy.org/>). The complete elastic model NoMelt_SPani is included in the Supplementary Material. Elastic tensors were manipulated and visualized using the MATLAB Seismic Anisotropy Toolkit (<https://geophysics.gly.bris.ac.uk/MSAT/>). Olivine flow laws were calculated using the Very Broadband Rheology calculator (<https://vbr-calc.github.io/vbr/>).

Acknowledgments

The authors thank Haemyeong Jung, Katsuyoshi Michibayashi, and Misha Bystricky who shared previously published elastic tensors used in this study. We thank Andr ea Tommasi and one anonymous reviewer whose thoughtful comments helped greatly improve this article. This work also benefited greatly from conversations with Benjamin Holtzman. This research was supported by the NSF under Grant No. OCE-0928270 and OCE-1538229 (J.B. Gaherty).

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