Deforming the Upper Mantle—Olivine Mechanical Properties and Anisotropy

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HOW DOES OLIVINE DEFORM?

Olivine is the most abundant mineral (>60 vol.%) in the Earth’s upper mantle. The slow and continuous deformation (creep) of mantle rocks is therefore controlled by the ability of olivine to deform. A reference parameter to understand how olivine deforms in the deep Earth is the homologous temperature, which is defined as the ratio of temperature T to the melting temperature (solidus) Tm of a crystalline material in Kelvin. The asthenospheric part of the upper mantle is the weakest, as the geotherm is the closest to the solidus. Similar to metals or ice, olivine can experience ductile deformation (without loss of material continuity) when the homologous temperature (T/Tm) is higher than 0.4 at mantle pressures (>1 GPa) (Wang 2016). The melting temperature of dry olivine with a typical mantle composition (Fo80) is Tm = 2053 K (1780 °C) at 0.1 MPa. Thus, T/Tm = 0.4 corresponds to a temperature of 821 K. This low temperature limits the ductile deformation of olivine and is consistent with the absence of intraplate oceanic earthquakes at depths below the ~600 °C isotherm.

At the microscopic scale, olivine deforms ductilely via three basic mechanisms: diffusion creep, dislocation creep, and grain boundary related mechanisms. All of these mechanisms involve the motion of defects. Olivine, like any crystalline solid, never exists with a perfect crystalline structure. Defects of various dimensions occur in the crystal lattice, from point, linear, and planar defects to three-dimensional inclusions or voids (cf. Jollands et al. 2023 this issue). All three deformation mechanisms are thermally activated because an increasing temperature enhances both the formation and migration of defects.

Point defects in the crystal lattice can be classified as vacancies (unoccupied lattice sites), interstitials (extra atoms between normal lattice sites), and substitutions (foreign atoms, see Jollands et al. 2023 this issue). Their interactions and equilibria are controlled by mass balance, lattice site balance, and electroneutrality (Nakamura and Schmalzried 1983; Jollands et al. 2023 this issue). Vacancies play a major role in deformation. Under differential stress, olivine may deform by (1) ionic diffusion of vacancies through the crystal lattice (Nabarro–Herring creep) and (2) ionic diffusion of vacancies along grain boundaries (Coble creep; Fig. 1A). In olivine, as in all silicates, Si⁴⁺ is the ionic species with the slowest diffusion rate. Thus, its self-diffusivity limits the diffusion creep rate, which has a positive dependence on temperature (and hydrogen concentration) but a negative dependence on pressure (e.g., Hirth and Kohlstedt 2003; Fei et al. 2013). During diffusion creep (Fig. 1A), grains tend to shorten parallel to the maximum compressive stress and lengthen parallel to the minimum compressive stress, forming a shape-preferred orientation. In nature, grain elongation is nevertheless limited by concomitant grain boundary migration aiming to decrease the grain boundary energy. Diffusive processes depend on the inverse of the grain size to a power 1 < m ≤ 3; an increase in grain size therefore reduces strain rates in the diffusion creep regime (e.g., Kohlstedt 2006).

Olivine also deforms by the motion of translational linear defects through the crystal lattice. These defects are called dislocations. A dislocation is defined by two vectors (Fig. 1B): the line vector u, also called the dislocation line, and the Burgers vector b, which defines the slip direction and has a length of one lattice parameter. Dislocations may be observed as straight lines displaying a clear edge (i.e., b perpendicular to l) or screw (i.e., b parallel to l) character...
Dislocation motion - Dislocation creep

Grain boundary motion

Vacancy motion - Diffusion creep

**Figure 1** Defects in olivine. (A) Diffusion creep by both Si diffusion within the crystalline lattice (rate controlled by vacancy self-diffusion and concentration) and along grain boundaries. (B) Dislocation creep: representation of an edge dislocation with the Burgers vector $b$ along a glide plane and a screw dislocation. Dislocation climb is not shown. (C) Grain boundary motion: sliding and migration of a grain boundary (misorientation/misalignment between the two crystalline lattices > 15°) and a subgrain boundary (misorientation < 15°).

(Fig. 1B) or as curved segments or loops composed of edge, screw, and mixed segments. The motion of edge and screw dislocations results in shearing of the crystal perpendicular or parallel to the dislocation line, respectively. The combinations of slip directions ($uvw$) and slip planes ($hkl$) in which dislocations may glide through the crystal lattice define the possible slip systems in a crystal. Olivine only possesses three independent slip systems: [100][010], [100][001], and [001][hk0]. In a crystalline aggregate such as a rock, deformation by motion of dislocations with a limited number of slip systems induces a rotation of the grains’ crystallographic lattice. It therefore produces a crystal preferred orientation (CPO), which progressively aligns the dominant slip system with the mantle flow direction and plane.

At low temperature ($<0.6T_m$) and high stress, ductile deformation of olivine occurs essentially by dislocation glide. However, interactions between dislocations hinder their motion and produce strain hardening. At higher temperature ($>0.6T_m$), edge dislocations can climb (i.e., move to another glide plane to avoid an obstacle) with the help of diffusion of specific vacancies called jogs. Climbing dislocations have more opportunity to annihilate with dislocations of opposite sign or to reorganize into lower energy configurations (subgrains). These mechanisms, which reduce the dislocation density in a crystal, are called “recovery”. They prevent strain hardening and allow steady-state conditions to be reached. Strain hardening by accumulation of dislocations may also be avoided via recrystallization, which involves the formation and migration of grain boundaries. If this occurs syn-deformation, recrystallization is called dynamic. Deformation by dislocation creep involves a combination of dislocation glide, climb, recovery processes, and dynamic recrystallization. This combination allows for steady-state conditions or even weakening. Dynamic recrystallization accelerates the evolution of the CPO at low strain and stabilizes it at high strain (e.g., Bystricky et al. 2000; Tommasi and Vauchez 2015). It also produces noticeable grain size reduction at low temperature and/or high stress, when strain rates are faster than grain boundary migration rates.

Grain boundaries also play an important role in the deformation of mantle rocks. Grain boundaries are interfaces with misorientations > 15° between adjacent crystals (Fig. 1C, NB: 15° is an arbitrary value, commonly used for mantle rocks). They are characterized by complex structures with significantly higher defect densities than grain interiors. The high densities of defects make grain boundaries easier paths for ionic diffusion. They are also sources or sinks of dislocations. Grains may also slide parallel to their boundaries (grain boundary sliding, Fig. 1C) or be sheared near the boundary (Cordier et al. 2014). These grain-boundary processes may be coupled to diffusion or dislocation motion. Diffusion-accommodated grain boundary sliding was, for instance, proposed to allow for superplasticity in the mantle with CPO formation (Miyazaki et al. 2013). Although it is the first process that comes to mind, grain boundary sliding is not the only grain boundary mechanism assisting ductile deformation. Grain boundary migration plays an essential role in dynamic recrystallization (Fig. 1C) and, hence, in dislocation creep. In addition, processes such as coupled migration and shear due to motion of specific grain boundary defects (disclinations, see Cordier et al. 2014; and disconnections) or stress-induced grain boundary amorphization have been put forward, but require further investigation. Finally, under fluid- or melt-present conditions, mantle rocks may also deform via dissolution-precipitation creep, which requires the transport of components dissolved in a fluid percolating along grain boundaries (e.g., Hidas et al. 2016).

**HOW DO WE MODEL OLIVINE DEFORMATION?**

The different deformation mechanisms presented above are likely to operate simultaneously, which can be expressed by the following constitutive equation:

$$\varepsilon_{\text{bulk}} = \varepsilon_{\text{diffusion}} + \varepsilon_{\text{dislocation}} + \varepsilon_{\text{grain boundary}} \quad (1)$$
where \( \dot{\varepsilon}_{\text{bulk}} \) is the macroscopic strain rate. One mechanism often dominates the rheology of mantle rocks at a given range of temperature, pressure, differential stress, strain rate, and chemical environment (hydration and redox states). Traditionally, the relationship between the strain rate \( \dot{\varepsilon} \) and stress \( \sigma \) for steady-state creep involving dislocation creep, diffusion creep, and grain boundary processes is approximated by a generic semi-empirical formulation (power flow law, e.g., Hirth and Kohlstedt 2003):

\[
\dot{\varepsilon}_{\text{high temperature}} = A d^{-m} f_{\text{H}_2\text{O}} \sigma^n \exp \left( \frac{V + q}{RT} \right)
\]  

where \( d \) is the grain size and \( m \) its exponent, \( f_{\text{H}_2\text{O}} \) is the water fugacity and \( r \) its exponent, \( n \) is the exponent of stress \( \sigma \), \( V \) is the activation volume, and \( R \) is the gas constant. \( A \) is a pre-exponential constant that depends on the material and hides many parameters not explicitly controlled during the experiments (e.g., grain boundary thickness, vacancy and jog concentration, Si self-diffusivity, dislocation density and velocity). Note that \( m \) and \( n \) are essentially fitting parameters to experimental data, yet their variation ranges permit discrimination between dominant deformation mechanisms in the experiments.

Most deformation experiments on olivine have been carried out at high temperature using fine-grained (2–100 \( \mu \)m) sintered olivine aggregates. These experiments typically yield the following parametrizations, which allow discrimination between different dominant deformation mechanisms (Hirth and Kohlstedt 2003; Kohlstedt 2006). Diffusion creep is sensitive to grain size \( m = 3 \) and has a linear visco-rheology \( (n = 1) \). Dislocation creep is independent of grain size \( m = 0 \) and has a non-linear dependence on stress \( (n = 3.5 \text{ or } 3 \text{ for dry and hydrated olivine, respectively}) \). Dislocation-accommodated grain boundary sliding is characterized by both a non-linear stress dependence \( (n = 3) \) and a lower influence of grain size on the strain rate \( m = 0.7 \) for dry olivine; Hansen et al. 2011, and \( m = 1.1 \) for olivine containing 20–130 ppm \( \text{H}_2\text{O} \) wt.; Ohuchi et al. 2015).

The generic semi-empirical flow law given by Eq. (2) breaks down at high stresses associated with deformation at \( T < 0.7T_m \). Under such conditions, recovery processes are less effective, and stress plays an increasing role on the dislocation interactions. Experimental data are then better approximated by an exponential flow law (e.g., Demouchy et al. 2013):

\[
\dot{\varepsilon}_{\text{low temperature}} = B \exp \left( \frac{q}{RT} (1 - \frac{\sigma}{\sigma_{\text{Peierls}}} )^q \right)
\]  

where \( B \) is a material constant, again hiding many parameters, \( \sigma_{\text{Peierls}} \) is a stress barrier for dislocation motion (see online supplemental material), and \( p \) and \( q \) are fitting parameters.

These flow laws of olivine provide a base for modeling upper mantle dynamics, but they are obtained at laboratory strain rates of \( 10^{-4} \text{–} 10^{-6} \text{ s}^{-1} \) and need to be extrapolated to mantle strain rates of \( 10^{-14} \text{–} 10^{-16} \text{ s}^{-1} \). As a first step to constrain this important extrapolation, Gouriet et al. (2019) modeled dislocation dynamics in an olivine crystal over a large range of strain rates (from laboratory rates to those prevailing on Earth) and temperatures between 800 and 1700 K. In these models, deformation is essentially produced by dislocation glide, but at a rate controlled by dislocation climb. Strain rates are thus controlled by both dislocation velocities and Si self-diffusivity, which are constrained by laboratory data. This study showed that the transition between the power and exponential laws is continuous at the microscopic scale, reconciling the low- and high-temperature ductile deformation of olivine single crystals, and proposed a unified flow law for dislocation creep (see online supplemental material for comparison between this flow law and previous formulations):

\[
\dot{\varepsilon}_{\text{unified dislocation creep}} = B \left( \frac{q}{p} \right)^q \exp \left( \frac{q}{RT} (1 - \frac{\sigma}{\sigma_{\text{Peierls}}} )^q \right)
\]  

where \( \mu \) is the shear modulus and the other parameters are similar to those in Eqs. (2) and (3). Best fit parameters for an olivine crystal well-oriented to deform with major activation of the [001](100) system are \( B = 1.7 \times 10^{-16} \text{ s}^{-1} \), \( n = 2.95 \), \( \sigma_{\text{Peierls}} = 2 \text{ GPa} \), \( p = 1.52 \), and \( q = 2 \). Equation (4) yields a nonlinear rheology with a stronger dependence of the strain rate on stress at low temperature. This equation cannot be inverted analytically, but alternative mathematical formalisms (hyperbolic tangent parameterizations) have been proposed, allowing it to be implemented in geodynamical models (e.g., Garel et al. 2020).

**WATER-WEAKENING PHENOMENON IN OLIVINE**

The water-weakening phenomenon was first observed in quartz, which showed a drastic drop in stress under hydrous conditions by one order of magnitude relative to compression experiments at constant strain rate on dry samples (Griggs 1967). Nine years later, a first interpretation was proposed based on observations in glasses: the hydroxylation of Si-O bonds accelerates the average velocity of mobile dislocations. Deformation experiments later demonstrated that hydrogen-doped olivine crystals and aggregates also deform at lower stresses than dry samples subjected to similar (constant) strain rates, but the stress drop is only by a factor of two. The current theoretical framework to explain this phenomenon is provided in Kohlstedt (2006). The fundamental equation behind this effect states that the self-diffusivity of an ionic species \( D_i \) is proportional to the diffusivity of its vacancies \( D_v \) times their concentration \( [V] \) (see Nakamura and Schmalzlried 1983).

\[
D_i \propto D_v \times [V]
\]

Due to mass balance and electroneutrality constraints, the addition of hydrogen, even at trace-element concentration levels, increases the concentration of Si vacancies, enhances Si self-diffusivity, and induces weakening during diffusion creep. Water-weakening also occurs in dislocation creep, where the strain rate depends on both dislocation glide and climb velocities. The latter are controlled by jog mobility, which depends on the Si vacancy concentration and Si self-diffusivity (Kohlstedt 2006). A similar reasoning can be postulated for grain boundaries, with incorporation of hydrogen facilitating recrystallization, sliding, and migration. However, experimental data quantifying these effects in a robust manner are still missing.

The quantification of the water-weakening effect at deformation rates \( (10^{-14} \text{–} 10^{-15} \text{ s}^{-1}) \) and hydrogen concentrations \( (10^{-20} \text{ wt. ppm H}_2\text{O}) \) relevant for the Earth’s mantle remains an open question. A compilation of hydrogen concentrations in olivine (Fig. 2) implies that hydrogen concentrations are usually not high enough to produce significant water-weakening (cf. online supplemental material).

Extrapolation of the most recent dislocation creep flow laws for hydrogen-doped olivine single crystals including the recent low-temperature data (Tielke et al. 2017) to upper mantle strain rates imply that significant water-weakening (effectively viscosity drop of 10; Fig. 3) is probably restricted to high strain rates \( (10^{-12} \text{ s}^{-1}) \) under a limited low temperature range (950–1100 °C; cf. online supplemental material). This extrapolation suggests that water-weakening might

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**Notes:**

- **ElEmEnts June 2023**
- **Equations (2), (3), and (4)**
- **WATER-WEAKENING PHENOMENON IN OLIVINE**
- **Experimental Data**
- **Supplemental Material**
- **Figure 2**
- **Table 1**
- **References**

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**Figure 2:**

- **Dislocation dynamics in an olivine crystal:**
  - **Experimental Data**
  - **Supplemental Material**
  - **Figure 1**
  - **Table 1**
  - **References**

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**References:**

be relevant for localized deformation in the mantle lithosphere, at plate boundary frontiers, for example, but not for the convective mantle. However, the impact of hydrogen stored at grain boundaries on the rheological properties of olivine-rich rocks still needs to be quantified.

PROBING MANTLE FLOW USING SEISMIC ANISOTROPY

Exposures of upper-mantle rocks (peridotites) at the Earth’s surface by tectonic or magmatic processes are rare but valuable. The study of these rocks constrains: (1) the deformation mechanisms involved in the lithospheric and shallow asthenospheric mantle, (2) possible interactions between deformation and fluids or magma transport, and (3) deformation patterns up to the kilometer scale. However, even the largest peridotite massifs, like the mantle section of the Oman ophiolite, with outcrops over 500 km², are only fossil testimonies of past deformation, extracted from their original context.

In situ observation of deformation and first-order mapping of the flow patterns in the upper mantle are possible using seismological data due to: (1) the strong elastic anisotropy of olivine (Abramson et al. 1997) and (2) mantle deformation producing coherent crystal preferred orientation (CPO, see above) of olivine at the scales sampled by seismic waves (tens of kilometers). This approach was pioneered by Hess (1964), who interpreted variations in sub-Moho P-wave velocities in different directions along refraction profiles as recording past ridge dynamics related to flow patterns frozen in the shallow mantle of the Pacific plate. Since then, measurement of seismic anisotropy, that is, velocity variations of seismic waves as a function of their propagation or polarization direction (Fig. 4), has been recognized as the most effective tool to probe past and present-day upper-mantle flow patterns.

Yet at face value, deduction of mantle flow patterns from seismic anisotropy data is not straightforward because it depends on: (1) the relationship between seismic anisotropy and olivine CPO or other structures that may produce large-scale seismic anisotropy, such as oriented melt pockets, and (2) the relationship between the source of anisotropy and deformation. As multiple processes or deformation geometries may produce similar seismic anisotropy signatures, the latter cannot be inverted to obtain unique flow patterns. Moreover, in most cases, seismological data only sample part of the elastic stiffness tensor and integrate data along the seismic wave path.

Fortunately, analysis of > 700 naturally deformed peridotites extracted from the lithospheric and shallow asthenospheric mantle in varied geodynamical settings revealed that 95% of the samples display olivine CPO characterized by dominant orientation of the [100] axes within the flow plane, with a maximum parallel to the flow direction (Fig. 5). Similar olivine CPOs develop during simple shear or compression experiments at high temperature, moderate pressure, and low water content (e.g., Bystricky et al. 2000).
This dominant olivine CPO in nature establishes a simple relationship between flow geometry and seismic anisotropy orientation: fast polarization directions of teleseismic S waves and fast propagation of P and Rayleigh waves indicate the flow direction (Fig. 4). We recall that when a S wave enters an anisotropic medium, it is split in two quasi-S waves polarized in perpendicular planes, which have different velocities ($V_{S1} > V_{S2}$). This phenomenon is called S-wave splitting. The delay time between the arrival of the fast $S_1$ and slow $S_2$ waves is a measure of the intensity of the anisotropy. Because it relates to the CPO concentration, it is tempting to interpret the intensity of seismic anisotropy as recording deformation intensity. However, this is not possible because CPO evolution saturates at low strain. CPO evolution also depends on the initial state and one should not forget that probably none of the present-day mantle is pristine or undeformed. Moreover, the anisotropy intensity also depends on how the seismic waves sample the anisotropic rock layer (pathlength and direction).

The simple relationship between flow geometry and seismic anisotropy presented above may break down in the deep upper mantle, forearc sections of the mantle wedge, and shear zones in the shallow lithospheric mantle because in these settings, high pressure, hydration status, or high stress may induce changes in olivine’s dominant slip systems, favoring glide along [001] relative to [100] (e.g., Jung and Karato 2001; Couvy et al. 2004). Such a change in the dominant slip system results in a concentration of [001] parallel to the flow direction and fast propagation of P waves and polarization of S waves normal to the flow direction. This may explain the change in orientation and decrease of seismic anisotropy in the upper mantle below 250 km (Mainprice et al. 2005). Moreover, it may also reconcile trench-parallel, fast S-wave polarization directions with trench-normal corner flow in the mantle wedge. Such olivine CPO is rare in natural mantle specimens, as illustrated in Fig. 5, being only observed in high-pressure peridotite massifs associated with eclogites, which record deformation under high-pressure conditions (e.g., Frese et al. 2003). Layering and CPO of hydrous minerals, like serpentine or chlorite, may alternatively explain trench-parallel anisotropy in forearc regions (e.g., Boudier et al. 2010). High-stress deformation could produce fast shear-wave polarization normal to the flow direction in the shallow sections of shear zones crosscutting the lithospheric mantle. However, samples from this type of shear zone show evidence for extensive recrystallization, grain size reduction, and a significant contribution of fluid-assisted grain boundary processes. These samples usually display weak olivine CPO and seismic anisotropy (e.g., Hidas et al. 2016).

Simple shear experiments with 3%–30% anhydrous melt fractions also produced alignment of olivine [001] axes subparallel to the shear direction and [010] axes subnormal to the shear plane, which was attributed to strain partitioning between melt-rich layers and melt-depleted lenses (Holtzman et al. 2003). However, natural peridotites deformed in the presence of anhydrous basaltic melts show dispersion of [100] axes in the flow plane, but no preferred alignment of [001] parallel to the flow direction (Higgie and Tommasi 2012). A weak preferred orientation of olivine [001] axes parallel to the lineation has nevertheless been described in cumulative dunites deformed by compaction of a crystal–magma mush (Yao et al. 2019). Yet, the seismic anisotropy associated with these CPOs are only observed seismically if the melts are extracted or crystallized. When melts are present, their effect overcomes that of the olivine CPO. Deformation-induced preferred orientation of melts in layers at low angle or parallel to the shear plane promotes strong seismic anisotropy with fast S-wave polarization and P-wave propagation directions parallel to the shear plane. Furthermore, S-wave velocities strongly decrease as a function of the melt concentration along the path.

Diffusion creep probably predominates in slowly deforming sections of the convective upper mantle. How does this affect the seismic anisotropy signal? Traditionally, diffusion creep was considered to not produce CPO. Moreover, recent deformation experiments in the diffusion creep regime produced weak olivine CPO with similar relationships to the flow pattern as those developed during dislocation creep under high-temperature and low-stress conditions (Miyazaki et al. 2013). Thus, the presence or absence of seismic anisotropy cannot be used to map transitions between dislocation and diffusion creep in the upper mantle.
SYNTHESIS: OLIVINE AND UPPER MANTLE DEFORMATION

The current evidence from deformation experiments does not always explain the observations on olivine-rich rocks. Nevertheless, some points can be asserted. (1) Analyses of natural peridotites imply that in the lithospheric mantle and shallow asthenosphere, the dominant deformation mechanism is dislocation creep of olivine. (2) By combining experimental results and numerical models, we now have a unified flow law to describe dislocation creep over the entire lithospheric and asthenospheric mantle. (3) Incorporation of hydrogen in the olivine structure could lower the strength of mantle rocks and change the relationship between seismic anisotropy and flow direction. However, evidence for a change in the dominant slip system due to hydrogen incorporation remains rare among naturally deformed peridotites (Fig. 5). In addition, extrapolation of the most recent experimental data to natural strain rates predicts no major water-weakening in olivine, except possibly in localized shear zones in the lithospheric mantle (Fig. 3). (4) The ever-growing database of geophysical measurements of seismic anisotropy is a unique tool to probe in situ upper mantle flow patterns (Fig. 4). The predominance of dislocation creep by [100] glide results in a simple relationship between olivine CPO and the seismic anisotropy of peridotites, which usually allows mantle flow directions to be determined from seismic anisotropy data. This simple relationship may nevertheless be questioned in domains that deform at high stresses, in the deep upper mantle (high pressure), and when olivine may hold a high hydrogen concentration. However, because multiple processes can produce similar seismic anisotropy signatures, seismic anisotropy data cannot be unequivocally inverted to infer the active or ancient deformation processes in the mantle or physio-chemical conditions of deformation. (5) The fundamental mechanisms of olivine deformation are not yet fully understood, especially grain boundary–based mechanisms and the impact of microstructure evolution on the mantle rheology with time.

Figure 5
Pie chart showing the repartition of olivine CPO and seismic anisotropy patterns from the database of the Geosciences Montpellier EBSD facility, composed of > 600 peridotites including both peridotite massifs and peridotite xenoliths (Tommasi and Vauchez 2015). Olivine CPO and seismic anisotropy are displayed as pole figures, which are a stereographic representation of the orientation of lines and planes in 3D space. In all pole figures, the line marks the flow plane (Z is normal to the flow plane) and the flow direction (X) is E-W. Olivine CPO pole figures present the orientation of the three main crystallographic axes ([100], [010], and [001]) relative to the structural reference frame (here the flow plane and direction). Seismic anisotropy pole figures present the variation of P-wave velocity and the difference in velocity between the fast and slow quasi-S waves as greyscale colors. The polarization orientation of the fast quasi-S waves is indicated by white or black line segments for different propagation directions relative to the structural reference frame. (A) Typical seismic anisotropy pattern for olivine CPO formed by dominant [100] glide (green sections). (B) Typical seismic anisotropy pattern for olivine CPO formed by dominant [001] glide (orange section). (C) Maximum P- and S-wave anisotropies resulting from these olivine CPOs; variations in seismic anisotropy intensity result from variations in CPO intensity (dominant effect) and symmetry. The 85% pie piece includes three CPO types (A, D, and AB) defined by Jung and Karato (2001) which produce similar seismic anisotropy patterns.
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ONLINE SUPPLEMENTS

1. Methodology for defining the unified dislocation creep flow law for Eq. (4)
2. Data source and methodology for Fig. 2
3. Methodology for Fig. 3
4. Data source for Fig. 4

REFERENCES


Cordier P and 5 coauthors (2014) Micro-mechanical tests to provide information on defect mechanisms, water-weakening, and presence of fluids/melts in the mantle. Olivine has yet to reveal all its secrets.


Seismic anisotropy of Earth’s upper mantle. Scientific Reports 6: 22772-22778, doi: 10.1038/srep22772


FUTURE OUTLOOKS

The multi-method approach presented here has not reached its limits. Further laboratory experiments should be conducted to explore how the rheology of upper-mantle rocks may evolve with time or strain. Novel approaches to unravel the olivine deformation mechanisms and anisotropy of physical properties include the use of in situ micro-mechanical tests to provide information on defect interactions in olivine, experiments on analog materials, and numerical modeling techniques, such as artificial intelligence, to quantitatively predict ductile deformation and its anisotropy. Numerical modeling at a large scale still has a role to play by exploring the types of feedback that are impossible to track in laboratory experiments, including those associated with changes in the dominant deformation mechanisms, water-weakening, and presence of fluids/melts in the mantle. Olivine has yet to reveal all its secrets.