

Backstress governs transient postseismic creep in the continental lithospheric mantle

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Abstract

Understanding the physical processes that control how the continental lithospheric mantle responds to sudden stress changes during the earthquake cycle is essential for interpreting postseismic deformation. For example, recent experiments have revealed that elastic interactions among dislocations generate backstresses that control transient creep of olivine, and the relevance of this mechanism can be confirmed by analysing the stress fields of dislocations in various tectonic settings. Therefore, we examine peridotites from the Finero Complex in the European Alps, which preserve microstructures formed during transient, earthquake-related deformation in exhumed continental lithospheric mantle. High-angular resolution electron backscatter diffraction reveals pronounced intragranular stress heterogeneity in olivine, with variations of several hundred megapascals over length scales on the order of 1–10 μm and correlated with elevated densities of geometrically necessary dislocations. These observations demonstrate that long-range dislocation interactions occur in naturally deformed olivine of the continental lithospheric mantle. Deformation-mechanism maps indicate that both dislocation-mediated deformation and diffusion creep contribute at steady state. We construct a model of the transient rheological behaviour of olivine including contributions from dislocation-mediated deformation, as well as diffusion creep, and elastically accommodated grain-boundary sliding (EAGBS) to explore the characteristics of such transients in contexts similar to those represented by the Finero peridotite. This analysis indicates that an earthquake-induced stress increase produces a complex transient evolution in mantle strength. Following rapid initial relaxation by EAGBS, dislocation-mediated deformation dominates subsequent stress relaxation; its strain rate initially increases by approximately three orders of magnitude, followed by progressive restrengthening as long-range elastic interactions build up. These results demonstrate that time-dependent microphysical models, rather than flow laws constructed for steady state alone, are required to describe the postseismic evolution of mantle viscosity.

Plain Language Summary

Earth's mantle does not only respond instantly to the stress changes caused by large earthquakes. Instead, it deforms over time, and understanding this behaviour is essential for interpreting how the mantle relaxes after earthquakes. In this study, we examined peridotite rocks that once experienced earthquake-related deformation and were brought to Earth's surface from the mantle in the Southern Alps. Using electron-microscopy techniques, we mapped distortions of the crystal structure of olivine, the dominant mineral in

the upper mantle. These measurements reveal that the olivine crystals contain abundant dislocations, which are defects in the crystal structure that distort the crystal and thereby generate stresses. We also estimated the stresses applied to the bulk rock from the grain sizes of different rock types and compared them with theoretical models of how olivine deforms. The results indicate that following an initial stage of grain-boundary sliding, most of the subsequent stress relaxation is accommodated by the motion of dislocations. We then used a microphysical model to predict how the mantle would respond to an earthquake-like stress change. The model predicts that the mantle becomes briefly weaker and flows faster just after an earthquake, before gradually strengthening again as internal stresses rebuild. Our findings show that the upper mantle can experience short-lived weakening after seismic events, and that this behaviour is controlled by the way dislocations interact inside olivine grains.

Key points

1. Microscopic stress heterogeneity in mantle rocks evidence long-range dislocation stresses.
2. After seismic loading, coarse-grained olivine weakens, then gradually restrengthens as dislocations elastically interact.
3. Models incorporating evolving internal stresses predict complex postseismic mantle flow that cannot be captured by steady-state laws.

1. Introduction

Earthquake cycles impose rapid changes in stress state that potentially extend across the full thickness of the lithosphere (Avouac, 2015; Campbell & Menegon, 2019; Ellis & Stöckhert, 2004; Henschel et al., 2022; Hetland & Simons, 2010; Muto et al., 2019; Nüchter & Ellis, 2010; Wang et al., 2012). Experiments performed on a wide variety of materials demonstrate that stress changes induce transient creep, which manifests as evolution in viscosity as a function of strain towards a steady-state value (e.g., Andrade, 1910; Duval et al., 1983; Fan et al., 2020, 2021a; Glen, 1952; Hansen et al., 2019, 2021; Trepmann et al., 2013; Trepmann & Seybold, 2019; Weertman, 1978; Weertman & Weertman, 1975). Consequently, transient creep is a fundamental process in the evolution of the rheological behaviour of the lithosphere following major earthquakes (Ellis & Stöckhert, 2004; Masuti et al., 2016; Muto et al., 2019).

Flow laws, which are key to modelling the rheological behaviours of mantle flow, reflect our understanding of the physical processes involved in deformation (Breithaupt et al., 2023; Hansen et al., 2011; Hirth & Kohlstedt, 2003; Yabe et al., 2020). Many large-scale geodynamic models developed for predicting the postseismic rheological behaviour of the mantle are based on modified versions of steady-state flow laws (e.g., Masuti et al., 2016; Muto et al., 2019). Although frequently used, the relevance of steady-state flow laws for describing transient postseismic creep has been questioned by both the experimental and geodetic modelling communities as the rate-limiting microphysical processes likely differ between transient and steady-state creep (e.g., Freed et al., 2010; Hansen et al., 2021; Nüchter & Ellis, 2010; Wallis et al., 2021, 2026).

The mechanical behaviours and microstructures of rocks provide information that is key to deciphering the processes associated with transient creep (Hansen et al., 2021; Hanson & Spetzler, 1994; Karato, 2008; Matysiak & Trepmann, 2015; Wallis et al., 2021, 2022; Wiesman et al., 2024). Laboratory experiments

often reveal a decrease in the apparent viscosity of olivine-rich rock following a change in applied stress (Chopra, 1997; Hansen et al., 2021; Hanson & Spetzler, 1994; Misra & Murrell, 1965; Post, 1977; Smith & Carpenter, 1987). Moreover, Hansen et al. (2021) observed transient, time-dependent reverse strain following stress reductions during uniaxial compression of single crystals of olivine, revealing the influence of an internal backstress that acts in opposition to the stress driving deformation. Microstructural analyses of the deformed olivine from these experiments, along with a range of other deformation experiments on polycrystalline olivine samples, revealed highly heterogeneous distributions of intragranular residual stress generated by dislocations (Hansen et al., 2021; Wallis et al., 2017, 2020, 2021; Wiesman et al., 2024). This intragranular stress heterogeneity causes long-range elastic interactions among dislocations that generate backstress and thereby result in a specific type of transient creep termed kinematic hardening (Hansen et al., 2019, 2021; Wallis et al., 2017, 2020, 2021). Motivated by these observations, Breithaupt et al. (2023) developed a quantitative framework for the deformation of olivine that couples a dislocation-glide flow law to the evolution of dislocation density and associated backstress. We use the term *backstress-based constitutive framework* to refer broadly to models in which the strain rate arising from dislocation-mediated deformation depends on an effective stress that accounts for an internal backstress generated by long-range interactions among dislocations, and in which this backstress evolves with the dislocation microstructure.

A key next step is to determine whether naturally deformed olivine preserves the long-range internal stresses required for this framework to describe transient creep under geological conditions. Wallis et al. (2022) demonstrated that olivine in peridotite from the oceanic lithosphere in the paleosubduction interface of the Oman-UAE ophiolite exhibits intragranular stress heterogeneity with characteristics similar to that developed in experimental samples. These observations motivated application of a backstress-based constitutive framework to predict instantaneous changes in viscosity across step changes in stress.

However, there has been limited testing for the presence or characteristics of intragranular stress heterogeneity in samples with microstructural evidence for stress pulses caused by nearby earthquakes. Likewise, the characteristics of intragranular stress heterogeneity in olivine in peridotites from the continental lithosphere remain poorly constrained. Furthermore, although Breithaupt et al. (2023) used their model to predict the strain- and time-dependent evolution of viscosity for generalised conditions in the lower lithosphere and asthenosphere, it has not yet been applied to explore transient behaviour using parameters guided by a specific geological example from the shallow continental lithosphere.

In this contribution, we address these points by providing microstructural observations of olivine from the Finero peridotite, Italy, to constrain the microphysical processes that contribute to transient creep of the continental lithospheric mantle during postseismic deformation. In particular, we use high-angular resolution electron backscatter diffraction (HR-EBSD, Wallis et al., 2019; Wilkinson et al., 2006) to map densities of geometrically necessary dislocations (GND) and intragranular stress heterogeneity. These data test whether naturally deformed olivine contains long-range internal stresses generated by dislocations and therefore whether evolving dislocation backstress is a physically relevant component of the transient rheological behaviour. We subsequently use the models of transient creep, including that of Breithaupt et al. (2023) for dislocation-mediated deformation, as an available implementation of a backstress-based constitutive framework to explore the time-dependent evolution of viscosity during postseismic transient creep under conditions guided by the Finero peridotite. We pair their model of transient dislocation-mediated deformation with a recently published exploration of the transient rheology of diffusion creep by Rudge (2026), anchored to the steady-state diffusion creep flow law from Hansen et al. (2011) and the theory of elastically accommodated grain-boundary sliding (Ashby, 1972).

2. Geological context

We selected samples from well-preserved, unaltered peridotitic shear zones distributed across the Finero Complex within the Ivrea-Verbano Zone (IVZ), Southern Alps. Figure 1 summarises the geological context of the IVZ and the internal structure of the southwestern Finero Complex. The IVZ has traditionally been interpreted to include material assembled in a subduction-related accretionary setting, in which mantle peridotite was tectonically mixed with metasedimentary rocks within the continental lower to middle crust (Quick et al., 1995).

Recent geochronological work suggests that the internal Finero Complex may include an inherited Ediacaran–Cambrian lower-crustal and upper-mantle assemblage (Daczko et al., 2026). Zircon U–Pb ages from syn-tectonic gabbroic sheets in the Layered Internal Zone, considered together with isotopic ages from the adjacent Phlogopite Peridotite and Amphibole Peridotite units, have been interpreted to indicate that these units were assembled during Ediacaran–Cambrian arc-related magmatism and deformation (Daczko et al., 2026). This older assemblage was subsequently reworked during late Palaeozoic–Mesozoic extension.

Post-Hercynian crustal extension, which started by the Late Carboniferous–Early Permian, was associated with the widespread development of localised high-temperature shear zones (Brodie & Rutter, 1987; Langone et al., 2017; Rutter et al., 1993). The crustal extension and thinning continued through the Late Triassic–Early Jurassic, during which the microstructure of the peridotite was transformed through deformation in long-lived, seismically active shear zones during exhumation (Brodie & Rutter, 1987; Hodges & Fountain, 1984; Langone et al., 2018; Rutter et al., 1993; Schmid et al., 1987). Previous microstructural, petrological, and geochronological studies have identified three stages of metamorphism under peak to retrograde conditions coeval with ductile and brittle deformation of these shear zones, as follows.

(1) Granulite-facies conditions, at temperatures of 670–830°C, are indicated by clinopyroxene–orthopyroxene thermometry of recrystallized protomylonitic clinopyroxene and orthopyroxene (Altenberger, 1995; Davidson & Lindsley, 1985; Gasparik, 1984; Lindsley, 1983; Wells, 1977). This temperature range is supported by Ti-in-zircon thermometry on magmatic and recrystallized zircons from the Mafic Granulite and Gabbro Unit (Langone et al., 2018).

(2) Amphibolite-facies conditions, at minimum temperatures of 630–650°C, are indicated by single-pyroxene thermometry of recrystallized ultramylonitic clinopyroxene (Altenberger, 1995; Davidson & Lindsley, 1985). These conditions are consistent with temperatures of 500–650°C and pressures of 400–600 MPa estimated by Kenkmann (2000) based on amphibole thermobarometry (Laird & Albee, 1981) and plagioclase–hornblende thermometry (Spear, 1980).

(3) Greenschist-facies conditions are constrained to temperatures less than 520°C by the upper stability limit of antigorite coexisting with tremolite (Altenberger, 1995; Evans & Trommsdorff, 1970; Trommsdorff & Connolly, 1990).

Consequently, the Finero peridotites preserve an older tectonomagmatic history that was overprinted during late Palaeozoic–Mesozoic extension, exhumation and cooling, together with microstructural records of shear-zone activity under progressively retrograde metamorphic conditions.

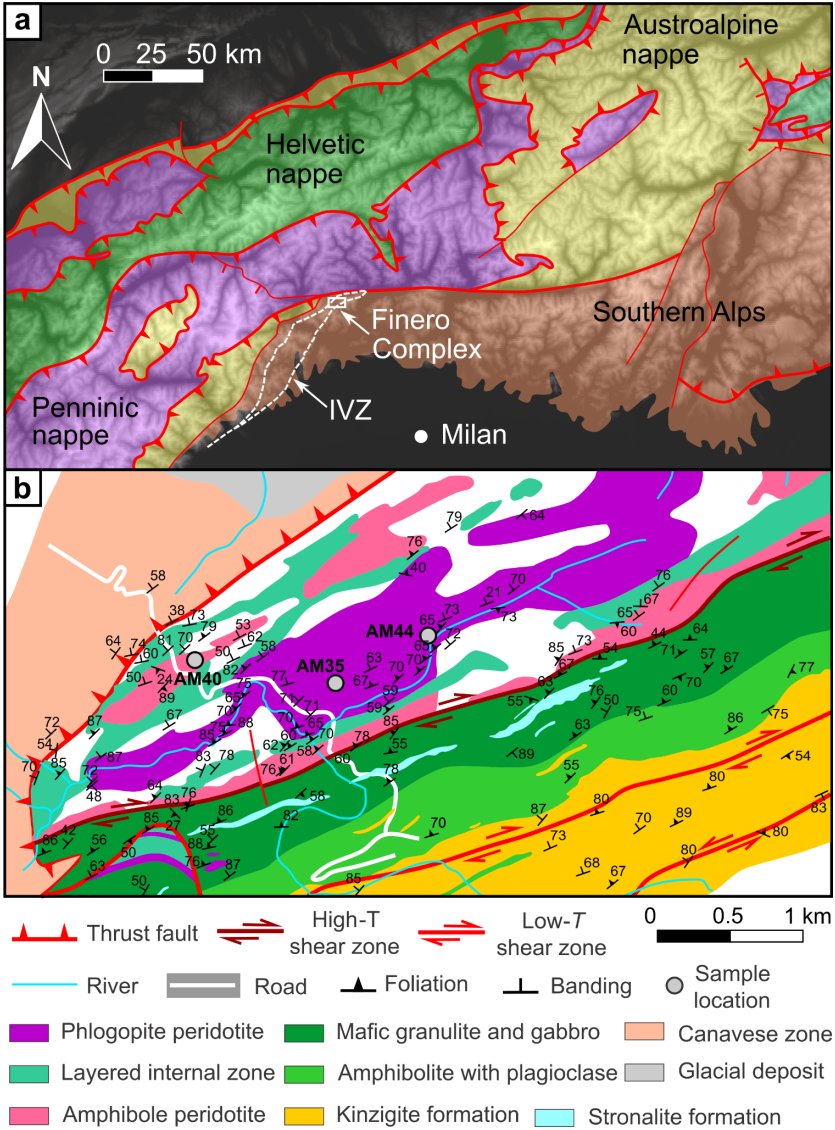


Figure 1. (a) Geological map of part of the western Southern Alps (modified after Neubauer, 2014) overlaid on a terrain elevation map with a horizontal resolution of 30 arc-seconds (Danielson & Gesch, 2011). White dashed lines indicate the extent of the Ivrea-Verbano Zone (IVZ); the white rectangle marks the location of the Finero Complex. **(b)** Geological map of the Finero Complex modified after Biedermann et al. (2020), Boriani (1995), Corvò et al. (2020), Giovanardi et al. (2014), Matysiak & Trepmann (2015), and Tommasi et al. (2017). Sample locations are from Matysiak & Trepmann (2015).

We analyse three olivine-rich samples collected from separate locations within two principal lithological units in the Finero Complex (Fig. 1b; Table 1). These samples, including relatively undeformed and deformed peridotites, were collected from or adjacent to largely unaltered peridotitic shear zones distributed over an area of approximately 2 km² within the peridotite massif at the core of the Finero Complex (Fig. 1b; Matysiak & Trepmann, 2015). This core mainly comprises three lithological units; the Phlogopite Peridotite Unit (purple), the Layered Internal Zone (mint green), and the Amphibole Peridotite Unit (pink) (Fig. 1b; Cawthorn, 1975; Vogt, 1962). These central units are bounded by the Insubric Fault Zone to the N-NW and the Mafic Granulite and Gabbro Unit (dark green) to the S-SE (Fig. 1b) (Giovanardi et al., 2014;

Matysiak & Trepmann, 2015; Rutter et al., 1993; Tommasi et al., 2017). Each lithological unit has subvertical contacts and contains widespread internal structures, such as layering/banding and/or foliation (Fig. 1b; Biedermann et al., 2020; Boriani, 1995; Corvò et al., 2020; Matysiak & Trepmann, 2015; Tommasi et al., 2017).

Table 1. Locations and geological context of the samples analysed in this study. Coordinates are from Matysiak and Trepmann (2015).

Sample No.	UTM zone	Easting (m)	Northing (m)	Lithological unit	Domain analysed
AM35	32T	464968	5106116	Phlogopite Peridotite Unit	Relatively undeformed wall rock
AM40	32T	464022	5106272	Amphibole Peridotite Unit and/or near its contact with the Layered Internal Zone	Protomylonite
AM44	32T	465595	5106439	Phlogopite Peridotite Unit	Protomylonite, mylonite and ultramylonite

Samples AM35 and AM44 were collected from the >500 m-thick Phlogopite Peridotite Unit (Fig. 1b), which exhibits decimetre-scale zones of ultramylonite (Giovanardi et al., 2014). This unit is predominantly composed of olivine (>85% by volume) along with diopside, enstatite, phlogopite, spinel, magnetite, and amphibole (Cawthorn, 1975; Vogt, 1962). AM35 contains relatively undeformed, olivine-rich wall rock. We use “wall rock” here in the structural sense of relatively less-deformed material adjacent to a shear zone.

Sample AM40 was collected from the Amphibole Peridotite Unit (Fig. 1b; Matysiak & Trepmann, 2015). Its amphibole-rich assemblage and geological setting differ from those of AM35 and AM44, which were collected from the mantle-derived Phlogopite Peridotite Unit. The Amphibole Peridotite has instead been interpreted as an ultramafic lower-crustal cumulate, and AM40 may therefore have a distinct petrogenetic history (Daczko et al., 2026; Siena & Coltorti, 1989). The ~400 m-thick Amphibole Peridotite Unit is characterised by the absence of phlogopite and a greater abundance of amphibole than the Phlogopite Peridotite Unit (Cawthorn, 1975; Vogt, 1962). The Layered Internal Zone is approximately 70–120 m thick in the study area, although its thickness varies considerably along strike, and comprises a compositionally layered sequence of peridotite, pyroxenite, anorthosite, hornblendite and amphibole-bearing gabbro, locally containing garnet (Daczko et al., 2026; Giovanardi et al., 2014).

Matysiak & Trepmann (2015) conducted detailed microstructural analyses on seven samples, collected from the core of the Finero Complex (Fig. 1b), that consist of wall-rock peridotite, along with protomylonitic, mylonitic, and ultramylonitic peridotite. Petrographic and electron backscatter diffraction (EBSD) data revealed contrasts in grain size and/or grain shape among different domains of the Finero peridotite (Matysiak & Trepmann, 2015), as follows.

(1) The wall rock probably represents the original mantle peridotite and was annealed at a temperature of ~1000°C and a pressure of ~1 GPa (Lu et al., 1997; Siena & Coltorti, 1989). The wall rock

exhibits a homogeneous, foam-like structure with grain sizes (area-equivalent diameters) generally larger than 1 mm.

(2) Protomylonite is characterised by spatially localised aggregates of small grains (50–500 μm) that are present along sutured grain boundaries of large porphyroclasts or kink-band boundaries. Small grains and porphyroclasts exhibit intragranular distortion as deformation bands, kink bands, and subgrain boundaries.

(3) Mylonite has an average grain size of $\sim 100 \mu\text{m}$ and isolated millimetric porphyroclasts. The grain-size refinement is inferred to have occurred by dynamic recrystallization similar to, but more complete than, that affecting the protomylonite. The mylonitic matrix has a foam-like structure suggestive of possible static annealing after dynamic recrystallization.

(4) Ultramylonite is characterised by bands of aggregates of finer grains, with grain sizes of approximately 5–30 μm , that are distributed along the sutured grain boundaries of coarser grains ($>50 \mu\text{m}$).

The microstructural contrasts amongst annealed mylonite and strongly deformed protomylonite (and possibly ultramylonite) suggests spatially inhomogeneous, localised deformation, which likely reflects transient deformation related to multiple seismic cycles (Matysiak & Trepman, 2015). Therefore, peridotite samples from Matysiak & Trepman (2015) have the potential to preserve key microstructural signals related to the underlying mechanisms of transient creep of olivine, including in lithologies typical of the continental lithospheric mantle.

3 Microstructural analytical method

We analyse thin sections from three olivine-rich samples from the Finero Complex: (1) AM35, containing relatively undeformed wall rock; (2) AM40, containing protomylonite; and (3) AM44, containing protomylonitic, mylonitic and ultramylonitic domains (Table 1).

3.1 Conventional EBSD

3.1.1 EBSD data collection

We collected EBSD data from the thin sections using a Zeiss Gemini 300 scanning electron microscope with an Oxford Instruments Symmetry EBSD detector and AZtec 4 acquisition software at the University of Cambridge. We coated the thin sections with $\sim 3 \text{ nm}$ of carbon prior to EBSD data collection to reduce charging and acquired data under high vacuum at an accelerating voltage of 30 kV. We collected EBSD datasets from the wall rock (AM35), protomylonite (AM40), mylonite (AM44), and ultramylonite (AM44) with step sizes of 30 μm , 11 μm , 2.5 μm , and 1 μm , respectively. All datasets were acquired using a camera gain of 1 and frame averaging of 2. For AM35, patterns were collected in sensitivity mode with 622×512 pixels; for AM40 and AM44, resolution mode was used to collect patterns with 1244×1024 pixels. The indexing rate of olivine across all datasets was greater than 93%.

3.1.2 EBSD data processing

We processed conventional EBSD data using the MTEX toolbox in MATLAB (Bachmann et al., 2010) with full datasets presented in Supporting Information S1. We identified olivine grains from raw EBSD data using a Voronoi decomposition algorithm (Bachmann et al., 2011) based on a threshold misorientation angle of 10° for grain boundaries. We applied noise reduction by removing clusters of fewer than four pixels with the same orientation that are likely to result from misindexing and by removing poorly constrained grains with $<50\%$ indexed pixel coverage.

We quantified the grain size, aspect ratio, and shape preferred orientation (SPO) of olivine. We measured grain size using the area-equivalent diameter, i.e., the diameter of a perfect circle that has the same area as the grain. We measured the aspect ratio from the quotient of the lengths of the long axis and short axis of an ellipse fit to each grain. We removed grains with aspect ratios less than 1.5 for SPO analyses as the angle between a given vector and the long axis of each grain estimated from the fitted ellipse has a low signal-to-noise ratio for grains with a convex hull close to a circle. Using the median grain size for a given domain, we estimated the bulk differential stress with a grain-size piezometer calibrated on experimentally deformed Fo90 using EBSD (Qi et al., 2021).

We identified subgrain boundaries as boundaries with misorientation angles less than 10° . We calculated misorientation axes of subgrain boundaries and displayed them in pole figures in the sample reference frame and in inverse pole figures in the crystal reference frame. We removed subgrain boundaries with misorientation angles less than 2° for this analysis, as misorientation axes calculated from conventional EBSD data are typically imprecise when the associated misorientation angle is small (Fan et al., 2022; Prior, 1999). To visualise the crystallographic preferred orientation (CPO), we plotted one point per olivine grain. To better display the CPO pattern, we contoured the CPO data with a half-width of 7.5° based on multiples of uniform distribution (MUD). We quantified CPO intensity using the M-index (Skemer et al., 2005).

3.1.3 Using grain-boundary sphericity to identify recrystallized grains

We segregated recrystallized grains from remnant grains using grain-boundary sphericity, a method first developed for identifying recrystallized grains in deformed ice samples by Fan et al. (2021b). In brief, we first calculated the grain-boundary sphericity parameter, Ψ , from grain area, A , grain-boundary perimeter, P , and area-equivalent radius, R , following $\Psi = A/PR$, for each grain of olivine. Sphericity, Ψ , ranges between 0.5 for grains that are perfectly circular to 0 for grains with infinitely irregular boundaries. In general, two distinct linear trends are observed in sphericity-grain size space for deformed samples: (1) small grains follow a steep trend where Ψ decreases rapidly with increasing grain size, and (2) grains above a threshold grain size follow a trend where Ψ decreases more gradually with grain size. Decreasing sphericity up to the threshold grain size is interpreted as a signal of newly formed, small, near spherical, recrystallized grains that grow by strain-induced grain-boundary migration (GBM) and thereby develop increasingly irregular grain boundaries (Fan et al., 2021b). The gradual decrease in sphericity at larger grain sizes represents a population of original grains (i.e., remnant grains) that becomes increasingly irregular (at similar rates) due to the operation of dynamic recrystallization (Fan et al., 2021b). Thus, the intersection between these trends provides an approximate practical threshold for discriminating between recrystallized and remnant grains. Linear regression was used to fit the data from each domain, and the intersection of these trends is taken as the grain-size threshold. We report the best-fit threshold and its 95% confidence interval based on the uncertainty of the regression fits.

3.2 High-angular resolution EBSD (HR-EBSD)

We collected 56 additional EBSD datasets from the same thin sections for HR-EBSD analysis following the procedures described by Wallis et al. (2016, 2019). We acquired diffraction patterns using a camera gain of 1 and frame averaging of 30. Testing by Wiesman & Wallis (2026), using the same acquisition settings, indicates that averaging ≥ 10 frames is enough that the addition of more frames makes negligible difference to the results and therefore that the 30 frames used here are more than sufficient. The patterns were saved as 8-bit TIFF images with 1244×1024 pixels, at an accelerating voltage of 30 kV and a working distance of 25 mm using an Oxford Instruments Symmetry EBSD camera and AZtec 4 data acquisition software. The microscope-detector system was calibrated for HR-EBSD using a Si-wafer standard

following Wilkinson et al. (2006). The position of the pattern centre was calibrated using diffraction patterns collected over a range of detector insertion distances (Maurice et al., 2011). Reference-frame conventions were validated using an approach similar to that of Britton et al. (2016).

We collected maps for HR-ESBD from within individual olivine grains in areas with a near absence of micrometre- to nanometre-scale scratches. For domains with different grain sizes, we utilized different combinations of map sizes and step sizes, with at least 50×50 points in each map, to optimise the characterisation of substructures and micromechanical states. For the wall rock (AM35), we collected 4 datasets with map sizes of $100 \times 100 \mu\text{m}$ or $50 \times 50 \mu\text{m}$ at step sizes of $1 \mu\text{m}$ or $0.5 \mu\text{m}$, respectively. For the protomylonite (AM40), we collected 28 datasets with map sizes of $25 \times 25 \mu\text{m}$ or $12.5 \times 12.5 \mu\text{m}$ at step sizes of $0.5 \mu\text{m}$ or $0.25 \mu\text{m}$, respectively. For the mylonite (AM44), we collected 14 datasets with a map size of $10 \times 10 \mu\text{m}$ at a step size of $0.2 \mu\text{m}$. For the ultramylonite (AM44), we collected 10 datasets with a map size of $5 \times 5 \mu\text{m}$ at a step size of $0.1 \mu\text{m}$. Three maps collected from the wall rock (AM35) and three maps collected from the protomylonite (AM40) contain a few scratches, so data points affected by these scratches were removed.

We processed the raw diffraction patterns following the HR-EBSD procedures described by Britton & Wilkinson (2012a) and Wallis et al. (2016, 2019a). The conceptual workflow is summarised in Fig. 2. In detail, for each map, we selected a high-quality diffraction pattern as the reference pattern (e.g., Fig. 2a) and compared the pattern at every other measurement point with this reference (e.g., Fig. 2b). Each comparison used 100 corresponding regions of interest (ROI), each containing 256×256 pixels (e.g., Figs. 2c, d). Fourier-based cross-correlation of each pair of ROIs determined a two-dimensional shift vector, $\mathbf{Q}_i = (Q_{x,i}, Q_{y,i})$ (e.g., Figs. 2e, f).

The set of ROI shifts, $\{\mathbf{Q}_i\}$, was fitted using the projective geometry of the diffraction pattern to determine the relative deformation-gradient tensor, β , which maps the lattice distortion at each measurement point relative to that at the reference point (Fig. 2g; Wallis et al., 2019). Under the small-strain and small-rotation approximation, the relative deformation-gradient tensor can be additively decomposed as

$$\beta = \epsilon_e + \omega + \mathbf{I} \quad (1)$$

where ϵ_e is the elastic strain tensor, ω is the rotation tensor, and $\mathbf{I}$ is the identity matrix. The ϵ_e and ω are the symmetric and antisymmetric components of β , respectively (Fig. 2g; Britton & Wilkinson, 2012b). This approach first estimates infinitesimal rotations through an initial cross-correlation pass, followed by a virtual pattern rotation to better align the test and reference patterns, and concludes with a second cross-correlation pass to more accurately capture elastic strains and fine-tune the rotation measurements.

We calculated residual stresses from ϵ_e using Hooke's law with the stiffness tensor of olivine (Fig. 2g; Abramson et al., 1997; Wallis et al., 2019; Wilkinson et al., 2006). The elastic strains, and hence also the stresses, were measured relative to the unknown strain state at the reference point. To provide an intuitive measure of stress heterogeneity that is independent of the choice of reference point, we mean-centred each component of the stress tensor, σ_{ij} , by subtracting the mean value of each component of the stress tensor, $\bar{\sigma}_{ij}$, from the measured components at each point within each map (e.g., Mikami et al., 2015; Wallis et al., 2019, 2022). We denote the resulting values as $\hat{\sigma}_{ij}$. These values represent the local components of the mean-centred residual-stress tensor expressed in the sample reference frame. In particular, $\hat{\sigma}_{11}$ and $\hat{\sigma}_{22}$ are in-plane normal-stress components, whereas $\hat{\sigma}_{12}$ is the in-plane shear-stress component. These quantities

represent spatial stress heterogeneity within individual grains and are distinct from the bulk differential stress used in the grain-size piezometer, flow laws and deformation-mechanism maps.

We calculated GND densities, ρ_{GND} , from spatial gradients, $\nabla\omega$, in the components of the rotation tensor, ω , using Nye analysis (Fig. 2g; Nye, 1953) according to the method of Wallis et al. (2016). The noise level of ρ_{GND} revealed by HR-EBSD depends on crystal orientation and step size, and is on the order of 10^{13} m^{-2} for our maps due to their relatively small step sizes (Wallis et al., 2016, 2019).

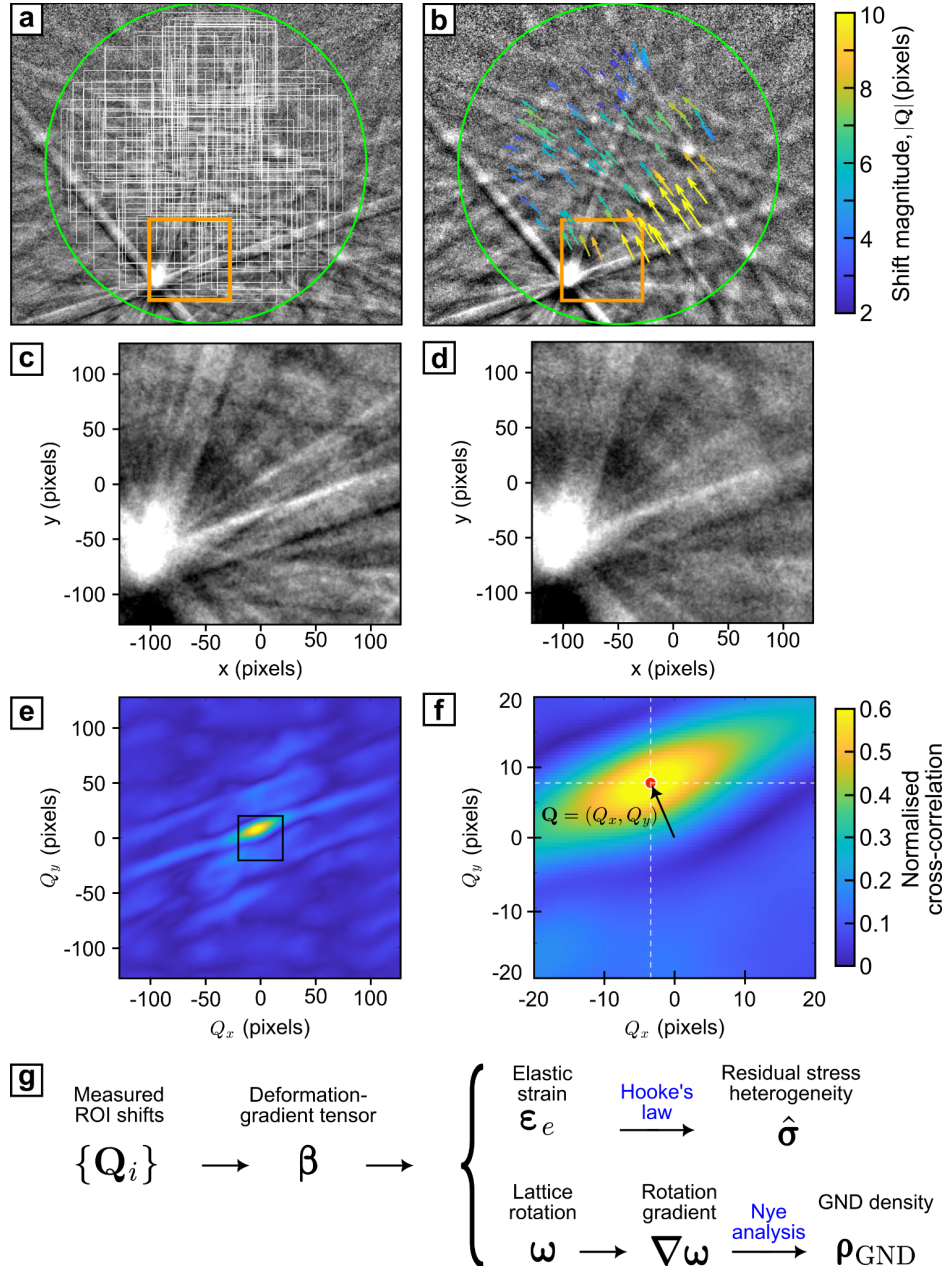


Figure 2. Conceptual workflow for determining diffraction-pattern shifts, residual-stress heterogeneity, and GND density by HR-EBSD. **(a)** Reference electron backscatter diffraction pattern showing the 100 regions of interest (ROI) used for cross correlation. The green circle marks the region within which ROI centres

were distributed, and the orange square marks the selected ROI illustrated in (c–f). **(b)** Comparison diffraction pattern showing the shift vector, $\mathbf{Q}_i$, measured for each ROI. Arrow directions indicate the directions of the measured shifts, whereas arrow colours indicate the shift magnitudes, $|\mathbf{Q}_i|$. Arrow lengths are enlarged by a constant factor for visibility. **(c)** and **(d)** are corresponding regions of the selected ROI from the reference and comparison patterns, respectively. **(e)** Normalised 2D cross-correlation function calculated from the ROIs in (c) and (d). The black square marks the area enlarged in (f). **(f)** Enlarged correlation peak. The red outlined marker and dashed white lines indicate the peak of the cross-correlation function that is fit to determine the shift, i.e., $\mathbf{Q} = (Q_x, Q_y)$. **(g)** Workflow for deriving residual-stress heterogeneity and geometrically necessary dislocation (GND) density from the measured ROI shifts. The set of ROI shifts, $\{\mathbf{Q}_i\}$, is fitted to determine the deformation-gradient tensor, β , which is decomposed into an elastic strain tensor, ϵ_e , and a lattice rotation tensor, ω . Elastic strain is converted to a residual-stress tensor, σ , using the anisotropic elastic stiffness tensor of olivine, and the resulting stress components are mean-centred to obtain residual-stress heterogeneity, $\hat{\sigma}$. Spatial gradients in lattice rotation, $\nabla\omega$, are related to GND density, ρ_{GND} , through the Nye analysis.

3.3 Statistics of stress heterogeneity

We use histograms (Section 3.3.1), normal probability plots (Section 3.3.2), and the restricted second moments (Section 3.3.3) of $\hat{\sigma}_{12}$ to quantify the intragranular stress heterogeneity in olivine within each microstructural domain. For each domain, the statistics were calculated using $\hat{\sigma}_{12}$ data aggregated from all the corresponding maps (full datasets in Supporting Information S2; Section 3.2). We chose $\hat{\sigma}_{12}$ for statistical analyses of stress heterogeneity because $\hat{\sigma}_{12}$ is the component of the stress tensor that is modified the least during the sample sectioning process and, being a shear stress, $\hat{\sigma}_{12}$ is particularly relevant to glide of dislocations (Wallis et al., 2019).

3.3.1 Probability distributions

We interrogate the mean-centred stresses by considering their probability distributions. The distribution of probability, P , of each component of $\hat{\sigma}_{ij}$ characterises the stress heterogeneity. The distribution of $\hat{\sigma}_{ij}$ is usually characterised by a peak at 0 and tails extending to stresses with greater absolute values (Fig. 3a) (Jiang et al., 2013a; Wallis et al., 2021; Wilkinson et al., 2014). The shape of the distribution can provide information on the cause(s) of the stresses. Measurement noise is typically normally distributed with stresses up to a few tens of megapascals (e.g., dashed black curve, Fig. 3a; c.f., Fig. 8a of Wallis et al., 2022). In contrast, broader distributions, particularly those with non-normal tails of $\hat{\sigma}_{ij}$ extending to greater stress magnitudes (e.g., solid red curve Fig. 3a) are indicative of significant stress heterogeneity, such as that arising from dislocations.

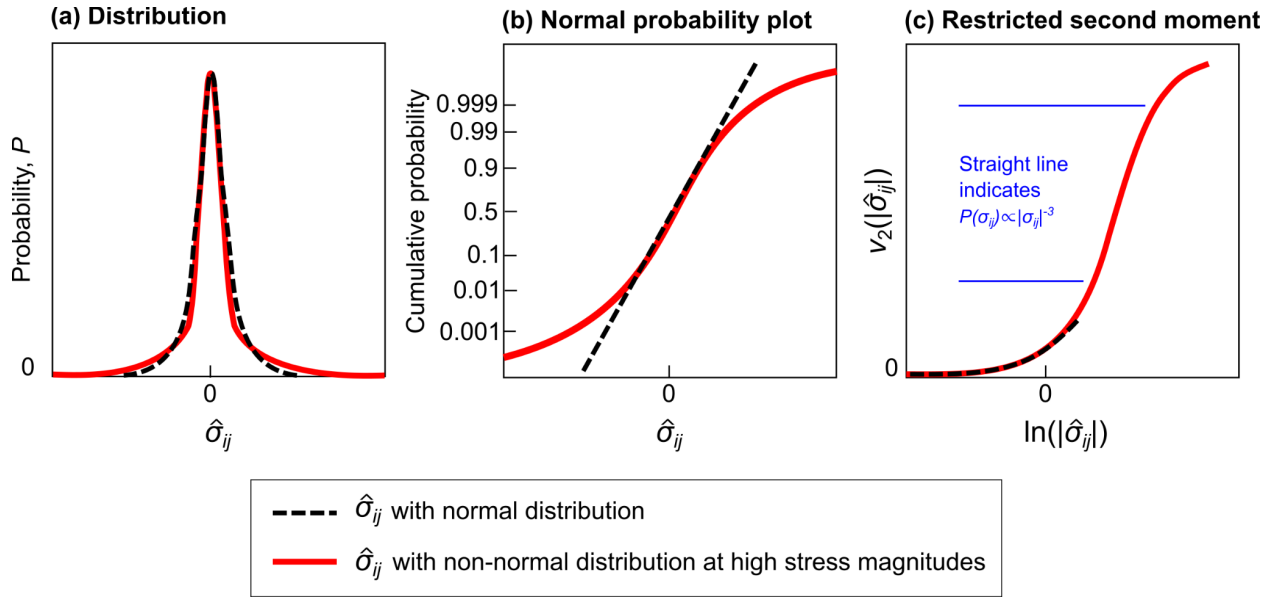


Figure 3. Schematic drawing showing (a) the probability distribution of each component of the mean-centred residual-stress tensor, $\hat{\sigma}_{ij}$, (b) a normal probability plot of $\hat{\sigma}_{ij}$, and (c) the restricted second moment of $|\hat{\sigma}_{ij}|$, $v_2(|\hat{\sigma}_{ij}|)$, as a function of the natural logarithm of $|\hat{\sigma}_{ij}|$, $\ln|\hat{\sigma}_{ij}|$, for idealised stress fields with a perfect normal distribution of $\hat{\sigma}_{ij}$ (dashed black curves) or with a non-normal distribution, in which probability decays as the inverse cube of $\hat{\sigma}_{ij}$ at high stress magnitudes (solid red curves).

3.3.2 Normal probability plot

We further interrogated the forms of the probability distributions of $\hat{\sigma}_{ij}$ with a normal probability plot, which allows assessment of whether a distribution follows, or deviates from, a normal distribution (Jiang et al., 2013a; Wallis et al., 2021; Wilkinson et al., 2014). On a normal probability plot, the cumulative-probability axis is scaled such that the cumulative probability distribution falls on a straight line (dashed black line; Fig. 3b) if $\hat{\sigma}_{ij}$ is normally distributed. For deformed crystals containing pronounced stress heterogeneity, the probability distribution of $\hat{\sigma}_{ij}$ usually departs from a straight line at high stress magnitudes indicating that the tails do not follow a normal distribution (compare the solid red curve with the dashed black line in Fig. 3b) (Jiang et al., 2013a; Wallis et al., 2021; Wilkinson et al., 2014).

3.3.3 Restricted second moments of the stress distributions

Residual stress heterogeneity within crystals can be imparted by one or more of several potential causes, including the stress fields of dislocations, changes in pressure and temperature, and chemical reactions of minerals (Holzhausen & Johnson, 1979). We can utilize the probability distribution of each component of the residual stress tensor, $P(\hat{\sigma}_{ij})$, to test whether the stress heterogeneity contains a contribution from the stress fields of dislocations (e.g., Wallis et al., 2021; Wilkinson et al., 2014). If the stress heterogeneity is due to dislocation stress fields, the high-stress portion of the probability distribution $P(\hat{\sigma}_{ij})$ should decay with the cube of the stress, i.e., $P(\sigma_{ij}) \propto |\sigma_{ij}|^{-3}$, and specifically

$$P(\sigma_{ij}) \rightarrow G^2 |b|^2 C_\sigma \langle \rho \rangle |\sigma_{ij}|^{-3}, \quad (2)$$

where G is the material dependant shear modulus; $|b|$ is the magnitude of the Burgers vector; ρ is the density of dislocations, and C_σ is the shear contrast factor depending on the types of dislocation, their

orientation, and the considered component of σ_{ij} (e.g., Kalácska et al., 2017; Wallis et al., 2021; Wilkinson et al., 2014).

The relation between $P(\sigma_{ij})$ and $|\sigma_{ij}|^{-3}$ can be quantitatively assessed for empirical distributions through the restricted second moment, v_2 ,

$$v_2(\sigma_{ij}) = \int_{-\sigma}^{+\sigma} P(\sigma_{ij}) \sigma_{ij}^2 d\sigma_{ij}, \quad (3)$$

which quantifies the shape of $P(\sigma_{ij})$ based on the integral over restricted ranges of stress (Wilkinson et al., 2014). If the probability distribution of stress follows $P(\sigma_{ij}) \propto |\sigma_{ij}|^{-3}$, then a plot of v_2 against the logarithm of the corresponding stress component, $\ln|\hat{\sigma}_{ij}|$, should fall on a straight line (Kalácska et al., 2017; Wallis et al., 2021; Wilkinson et al., 2014). Consequently, the straight form of the plot of $v_2(|\hat{\sigma}_{ij}|)$ versus $\ln|\hat{\sigma}_{ij}|$ can be used to identify the presence of stress fields generated by dislocations (Fig. 3c).

3.4 Oxidation decoration

Portions of wall rock from sample AM35 and protomylonite from AM44 were prepared for dislocation decoration following the approach of Kohlstedt et al. (1976). Each specimen was heated in air at 900°C for 45 minutes, allowing dislocations within olivine to become decorated by oxidation. Following decoration, the samples have a thin oxidised surface rind, which was removed by polishing with 0.03 μm colloidal silica for ~30 minutes. This step exposed fresh olivine in which the decorated dislocations were retained. To minimise charging during electron imaging, the polished surfaces were coated with ~6 nm of carbon.

The decorated microstructures were examined using forescatter images in a Quanta-650F field emission gun scanning electron microscope equipped with an Oxford Instruments Symmetry S3 EBSD detector at the Department of Earth Sciences in the University of Cambridge. Operating conditions included an accelerating voltage of 30 kV and an aperture size of 100 μm . Image contrast was subsequently adjusted to maximise the visibility of the oxidised dislocations.

4. Results

4.1 Microstructural analysis using visible-light and electron microscopy

We analyse olivine grains from different domains, including wall rock, protomylonite, mylonite, and ultramylonite (Figs. 4, 5, 6; Section 2). We base our quantitative microstructural analysis on much (> 5 times) larger areas than the subregions in Figs. 5 and 6, which are smaller subsets presented for visual clarity. The full maps are available in Supporting Information S1. Microstructures of orthopyroxene, clinopyroxene, and spinel (Figs. 5a, 5f, 6a, 6f), which are also components of the Finero peridotite, are beyond the scope of this paper due to their lesser volume fractions and have been described elsewhere (Matysiak & Trepmann, 2015; Tommasi et al., 2017).

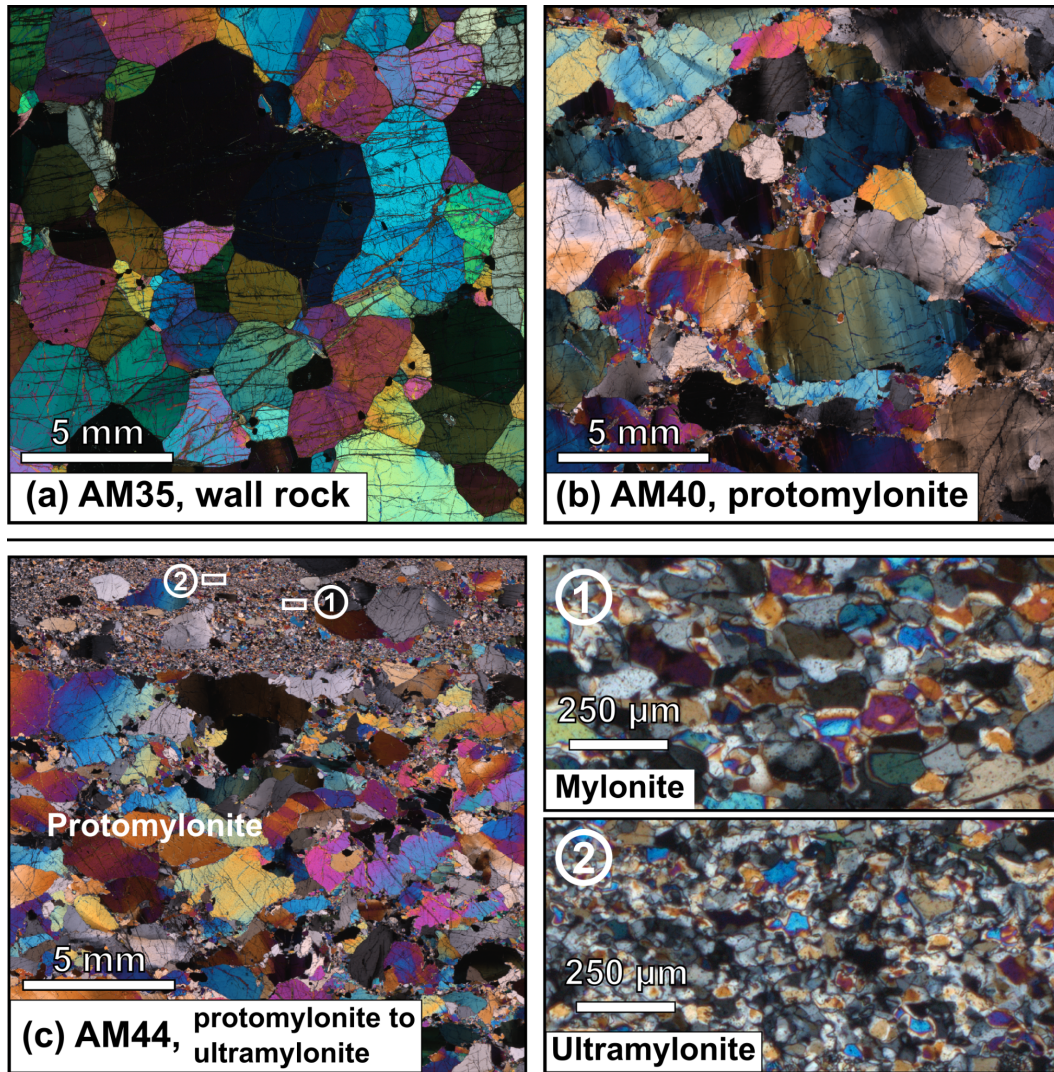


Figure 4. Polarised light photomicrographs (with crossed polarisers) of (a) AM35, containing wall rock, (b) AM40, containing protomylonite, and (c) AM44, containing protomylonite, mylonite, and ultramylonite.

4.1.1 Wall rock

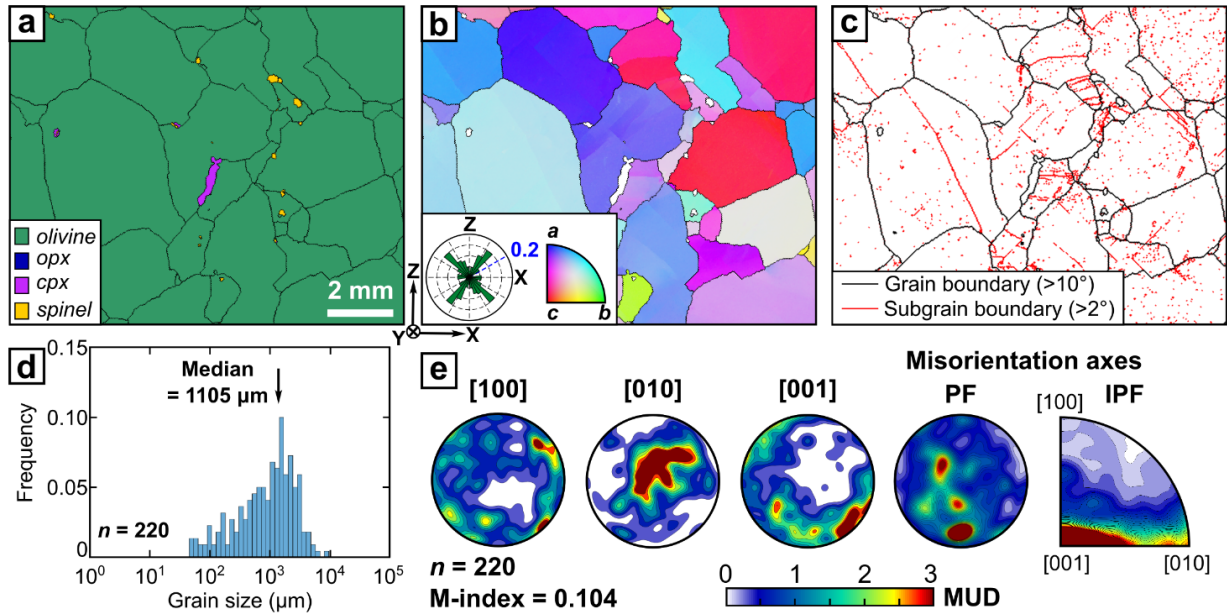
Olivine grains within the wall rock exhibit a foam-like microstructure with straight or slightly curved grain boundaries and polygonal grain shapes (Figs. 4a, 5b, 5c). The SPO of olivine grains with aspect ratios greater than 1.5 is characterised by two maxima at $\sim 90^\circ$ to each other (Fig. 5b). The distribution of grain sizes on a logarithmic scale is slightly skewed, with a median grain size of $\sim 1100 \mu\text{m}$ (Fig. 5d). Subgrain boundaries are present within a small fraction of grains (Fig. 5c). The CPO is characterised by distinct orthogonal clusters of [100], [010], and [001] (Fig. 5e). One maximum of long axes in the SPO is approximately parallel to the maximum of [100] in the CPO and the other is approximately parallel to the maximum of [001] (Figs. 5b and 5e). Misorientation axes of subgrain boundaries are distributed between [001] and [010] with the maximum near [001] (Fig. 5e).

4.1.2 Protomylonite

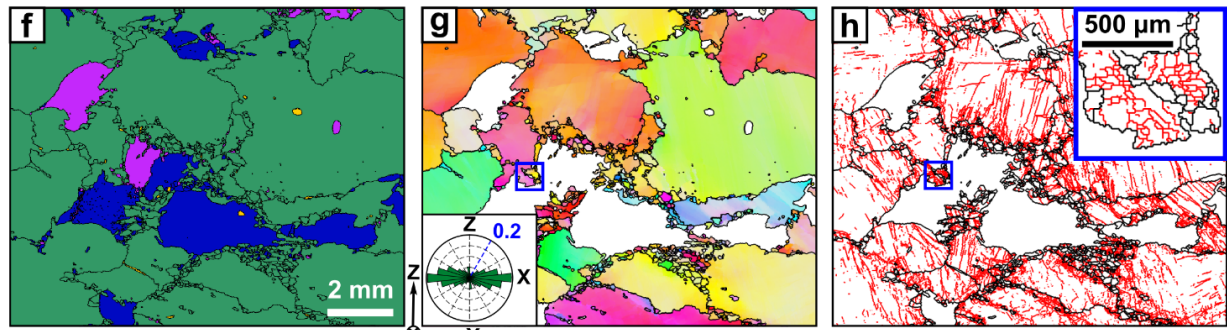
Olivine within the protomylonite consists of large grains with more irregular grain boundaries interlocking with networks of smaller grains with more regular grain boundaries (Figs. 4b, 5g, 5h). The SPO of grains with aspect ratios greater than 1.5 is characterised by a maximum subparallel to the X axis (Fig. 5g). Subgrain boundaries are common (Fig. 5h).

Figure 5j presents the median value of grain-boundary sphericity against grain size. We first identify two grain-size domains that exhibit distinct sphericity-grain size trends: grains smaller than 200 μm , for which sphericity decreases sharply with increasing grain size, and grains larger than 1000 μm , for which sphericity decreases more gradually with increasing grain size. The best fits to the two linear trends identified from the sphericity–grain size data intersect at a grain size of 292 μm (thick cross mark; Fig. 5j) and therefore we use this value as a cut-off grain size to segregate recrystallized grains from remnants of original grains. The grain-size distribution of recrystallized grains is close to log-normal with a median grain size of 67 μm (Fig. 5k). This value is similar to a value of 63 μm from the method of Cross et al. (2017). This median grain size corresponds to an estimated differential stress of ~ 20 MPa (Section 3.1.2). Recrystallized grains and remnant grains have similar CPO with [100] subparallel to the X axis and [001] subparallel to the Z axis (Figs. 5g, 5l). The CPO of the remnant grains is stronger than that of the recrystallized grains with values of M-index of approximately 0.12 and 0.06, respectively. Misorientation axes of subgrain boundaries in recrystallized grains are distributed between [001] and [010] with the maximum near [010] (Fig. 5l). Those in remnant grains have maxima near [001] and [010] (Fig. 5l).

Wall rock



Protomylonite



Segregation of recrystallized and remnant grains using grain boundary irregularity

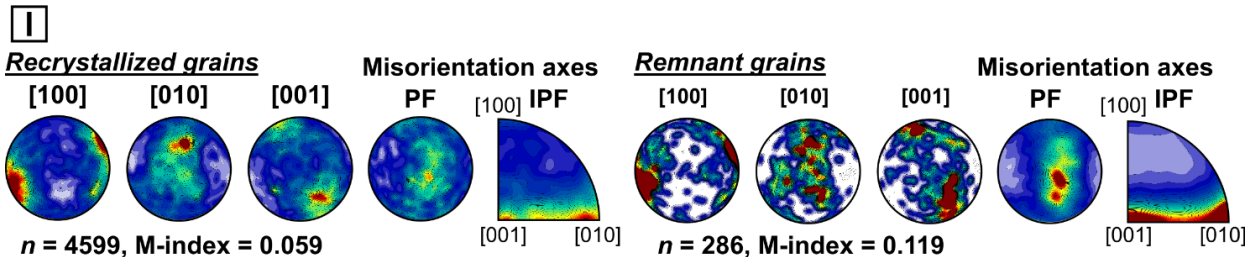
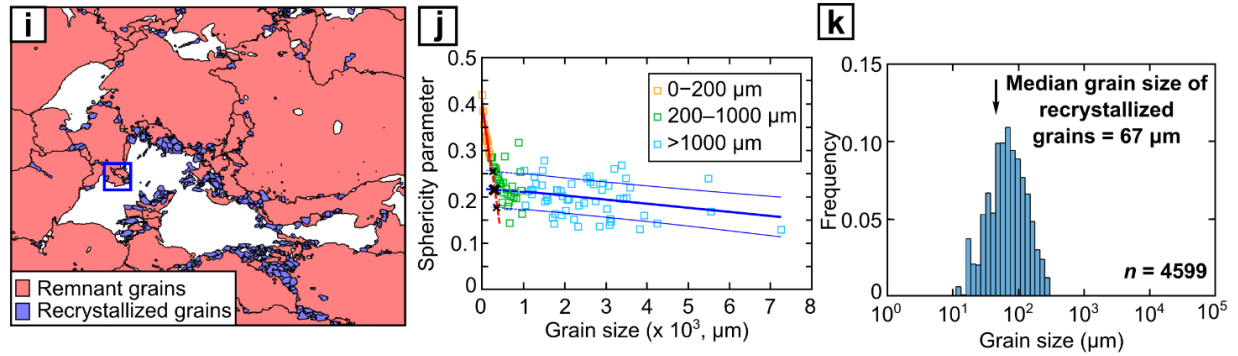


Figure 5. Microstructural analyses of wall rock (**a–e**) and protomylonite (**f–l**). Microstructural statistics are based on areas that are much larger than the subareas presented in (**a–c**) and (**f–h**). (**a**) and (**f**) are phase maps. (**b**) and (**g**) are orientation maps, in which the colour map indicates the crystallographic axes that are parallel to a specific direction as shown by the black arrows to the lower left of the box. To clearly illustrate grain orientations, (**b**) is coloured by the inverse pole figure referred to the Y axis, whereas (**g**) is coloured in reference to the Z axis. The SPO of olivine grains with aspect ratios greater than 1.5 is inset in the lower left of each of these subfigures. (**c**) and (**h**) are boundary maps. (**d**) is the grain-size distribution of olivine in wall rock plotted on a logarithmic grain-size scale. The median grain size is marked by a black arrow. (**e**) and (**i**) are pole figures (PF) of [100], [010], and [001] axes along with pole figures and inverse pole figures (IPF) of the misorientation axes of subgrain boundaries. (**j**) is a map of recrystallized and remnant olivine grains, which are segregated by the grain-boundary sphericity method (Fan et al., 2021b). (**k**) is a scatter plot of median sphericity parameter, Ψ , for each grain-size interval with a bin width of 20 μm . We fit separate linear relationships between sphericity and grain size for smaller grains ($d < 200 \mu\text{m}$) and larger grains ($d > 1000 \mu\text{m}$). Thick lines are linear best fits and thin lines are the upper and lower bounds of the 95% confidence intervals of the fits. Crosses mark the intersections of each of the corresponding lines from different grain-size classes.

4.1.3 Mylonite and ultramylonite

In mylonite, olivine grain boundaries are typically straight or slightly curved, while olivine phase boundaries exhibit more irregular shapes (Figs. 4c, 6a–6c). In ultramylonite, olivine grains have irregular grain boundaries and elongate shapes and the phase boundaries are also irregular (Figs. 4c, 6f–6h). The SPO of olivine grains is characterised by a maximum subparallel to the X axis (Figs. 6b, 6g). Subgrain boundaries are present within a few grains (Figs. 6c, 6h). The grain-size distributions are approximately log-normal with median grain sizes of 31 μm in the mylonite and 16 μm in the ultramylonite (Fig. 6d, 6i), corresponding to estimated stresses of ~ 40 MPa and ~ 80 MPa, respectively (Section 3.1.2). These stress estimates are upper bounds due to the potential effect of secondary phases pinning grain boundaries and thereby inhibiting grain growth. The CPO of the mylonite is similar to that of the protomylonite, with a maximum of [100] parallel to the X axis and a maximum of [001] at a high angle to the X axis, but it is weaker in strength with an M-index of approximately 0.01. The ultramylonite lacks a discernible CPO and has an M-index of 0.005. The misorientation axes of subgrain boundaries in olivine within the mylonite and ultramylonite exhibit near random distributions (Figs. 6e, 6j).

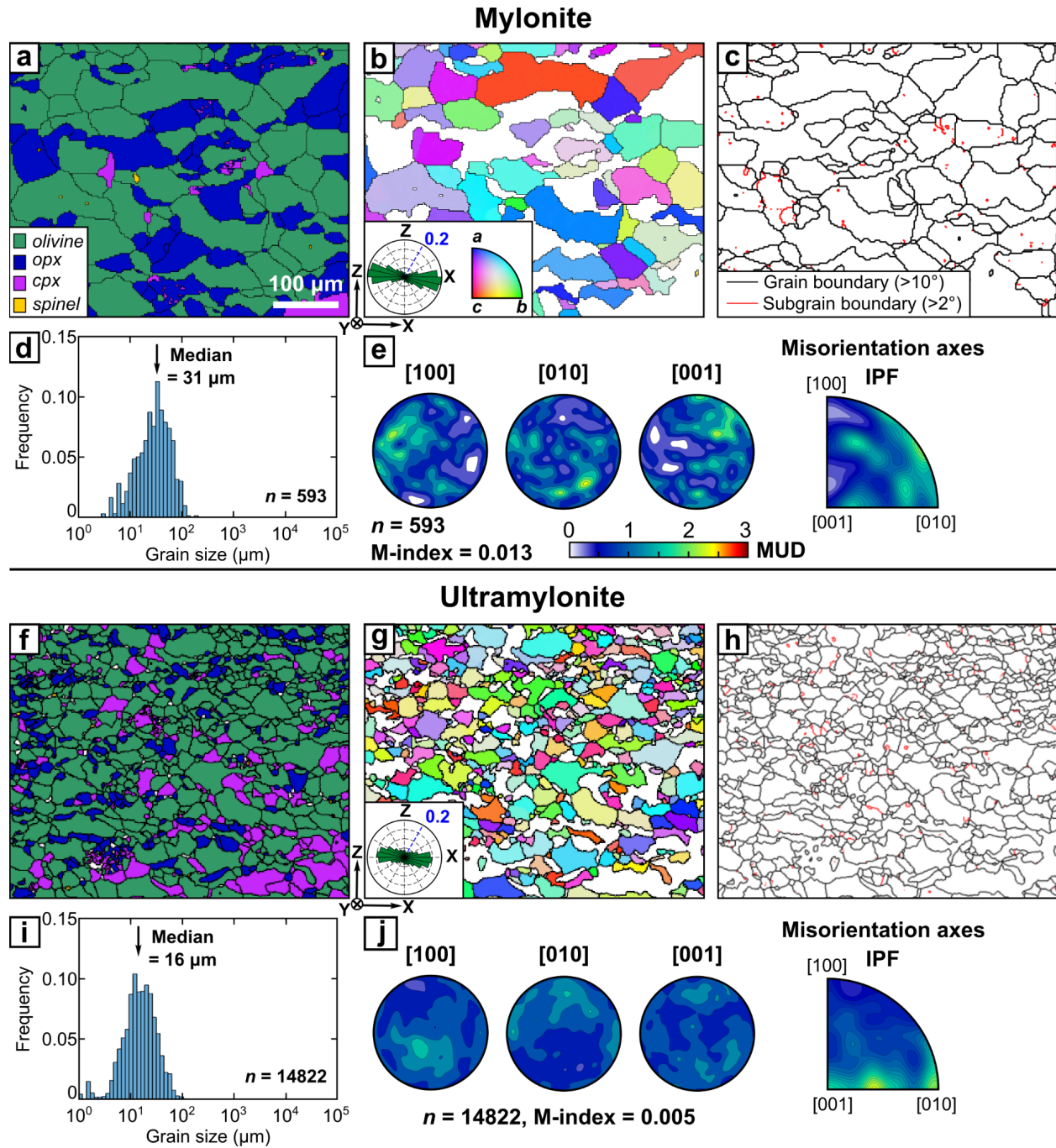


Figure 6. Microstructural analyses of mylonite (a–e) and ultramylonite (f–j). Descriptions of (a–e) and (f–j) are the same as the descriptions of (a–e) in Fig. 5, except that (b) and (g) are coloured by the inverse pole figure referred to the Z axis.

4.2 Intragranular microstructural and stress analysis from HR-EBSD data

We acquired HR-EBSD maps from 56 areas within wall rock, protomylonite, mylonite, and ultramylonite (Section 3.2). Figures 7–11 present representative datasets selected to illustrate the range of degrees of intragranular deformation within each rock type. All the HR-EBSD data are available in Supporting Information S2. In each of Figures 7–11, we present (a) foreshattered electron images that demonstrate the

surface topography and quality with a small amount of orientation contrast (Britton et al., 2018), (b) maps of the logarithm of the total density of geometrically necessary dislocations (GNDs), (c) maps of each in-plane component of the mean-centred stress tensor ($\hat{\sigma}_{ij}$), (d) raw diffraction patterns of the reference points (illustrated as pink dots in c) that demonstrate the data quality, and (e) crystal orientations displayed in pole figures.

4.2.1 Densities of geometrically necessary dislocations (GNDs)

The densities and distributions of GNDs in olivine vary between different microstructural domains. Olivine grains in the wall rock generally exhibit apparent GND densities below $\sim 10^{13} \text{ m}^{-2}$ (Fig. 7b). A few bands with GND densities of 10^{13} – $10^{13.5} \text{ m}^{-2}$ can be traced within individual grains (e.g., AM35-03, AM35-04; Fig. 7b). The protomylonite contains some grains with low apparent GND densities (close to or less than 10^{13} m^{-2}) (e.g., AM40-10; Fig. 8b. AM40-19; Fig. 9b). However, many other grains in the protomylonite are dissected by discrete bands with elevated GND densities of $\geq 10^{13.5} \text{ m}^{-2}$ (Figs. 8b, 9b). Some bands are parallel (e.g., AM40-04, AM40-08; Fig. 8b) whereas others intersect (e.g., AM40-01; Fig. 8b. AM40-13, AM40-16; Fig. 9b). Patches with elevated GND density ($10^{13.5}$ – 10^{14} m^{-2}) are infrequent and do not contain discrete bands (e.g., AM40-18; Fig. 9b). Olivine grains within mylonite and ultramylonite generally do not exhibit distinct bands with elevated GND densities, but do exhibit more diffuse linear structures locally (e.g., AM44-08, AM44-13; Fig. 10b. AM44-17, AM44-19; Fig. 11b). Instead, the mylonite and ultramylonite exhibit broader areas that have GND densities varying between $10^{13.5}$ and 10^{14} m^{-2} (Figs. 10b, 11b).

4.2.2 Spatial heterogeneity of residual stress

Olivine grains in the wall rock exhibit relatively homogeneous stress distributions, having standard deviations of approximately 40–70 MPa across each map, with occasional stress concentrations (Fig. 7c). In contrast, pronounced stress concentrations within olivine grains are common in the protomylonite, mylonite, and ultramylonite (Figs. 8c–11c). The protomylonite contains many bands of elevated stresses on the order of several hundred megapascals that vary over a few micrometres (e.g., AM40-04, Fig. 8c. AM40-13, AM40-16; Fig. 9c). Many areas within protomylonite, mylonite and ultramylonite contain diffuse (i.e., without a clear linear structure) spatial distributions of smooth stress heterogeneity, with individual mean-centred residual-stress components (e.g., $\hat{\sigma}_{12}$) spanning several hundred megapascals over length scales of approximately 1–10 μm (e.g., AM40-10, Fig. 8c; AM40-18, Fig. 9c; AM44-08, AM44-13, Fig. 10c; AM44-15, AM44-17, Fig. 11c). We note that these values describe local variations in residual stress within individual grains and do not represent the bulk differential stress applied to the rock.

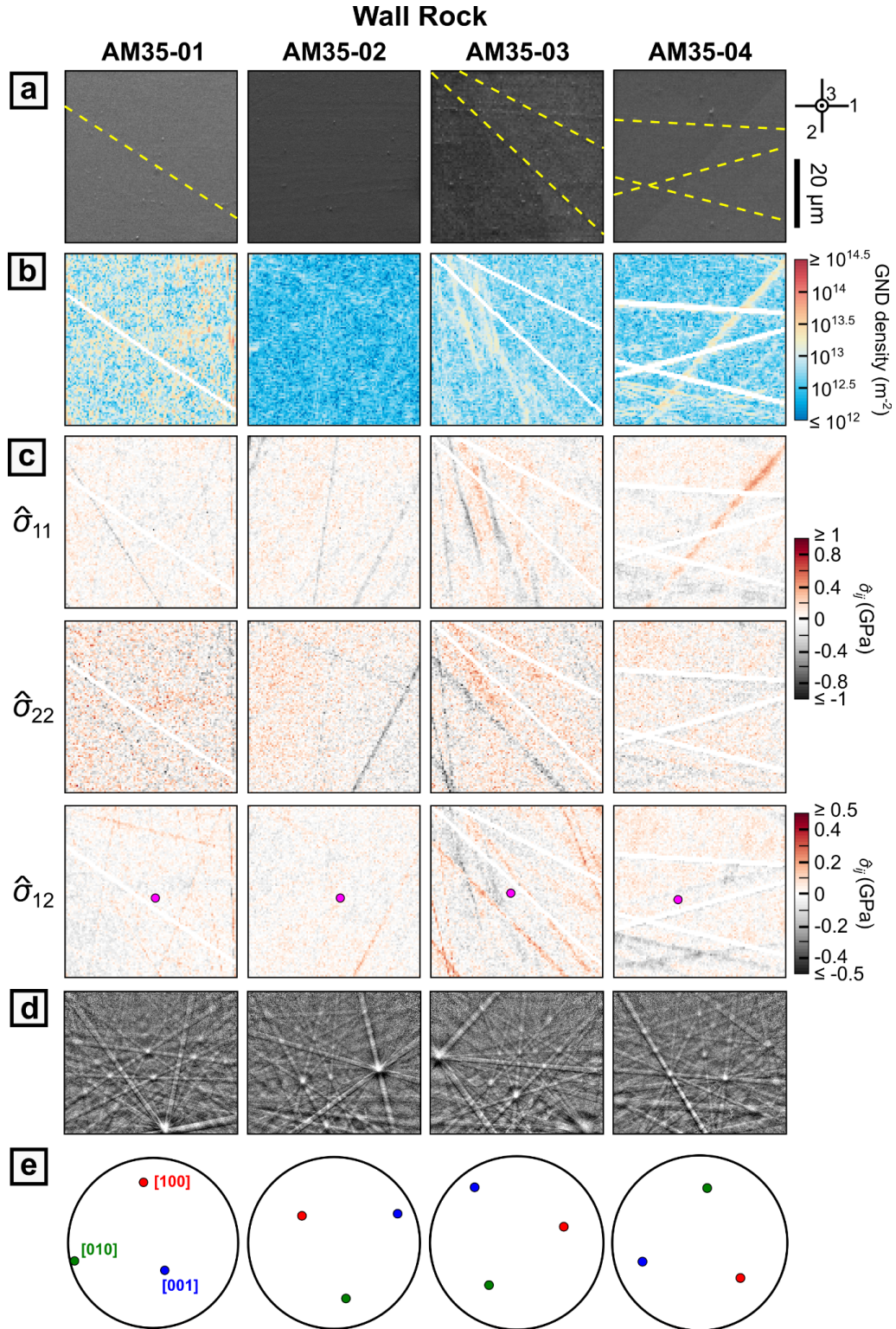


Figure 7. HR-EBSD data collected from the wall rock. **(a)** Fore-scattered electron images. Dashed yellow lines indicate surface scratches with corresponding pixels removed from analyses. **(b)** The logarithm of the total densities of geometrically necessary dislocations (GNDs) estimated from lattice rotations measured by HR-EBSD. **(c)** Maps of mean-centred residual-stress components ($\hat{\sigma}_{11}$, $\hat{\sigma}_{22}$, and $\hat{\sigma}_{12}$) calculated by

subtracting the mean value within each grain. In (b) and (c), white areas indicate pixels corresponding to surface scratches or pixels that failed the quality criteria for the cross-correlation analysis (Britton & Wilkinson, 2011); these pixels were removed from analysis. (d) Raw diffraction patterns of the reference points (pink dots in (c)). (e) Pole figures illustrating crystal orientations.

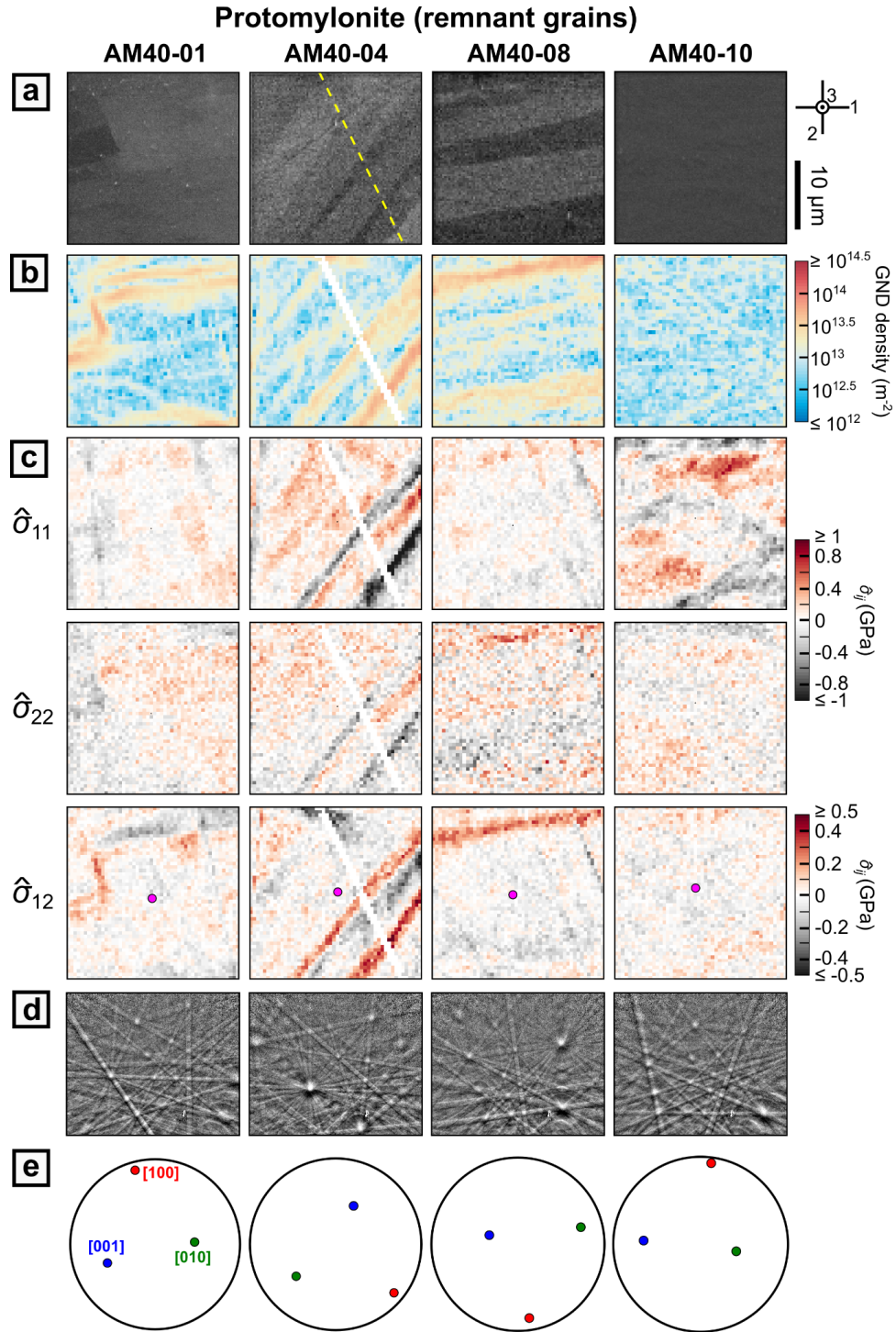


Figure 8. HR-EBSD data collected from the remnant grains within protomylonite (AM40). Descriptions of (a)–(e) are the same as those for Fig. 7.

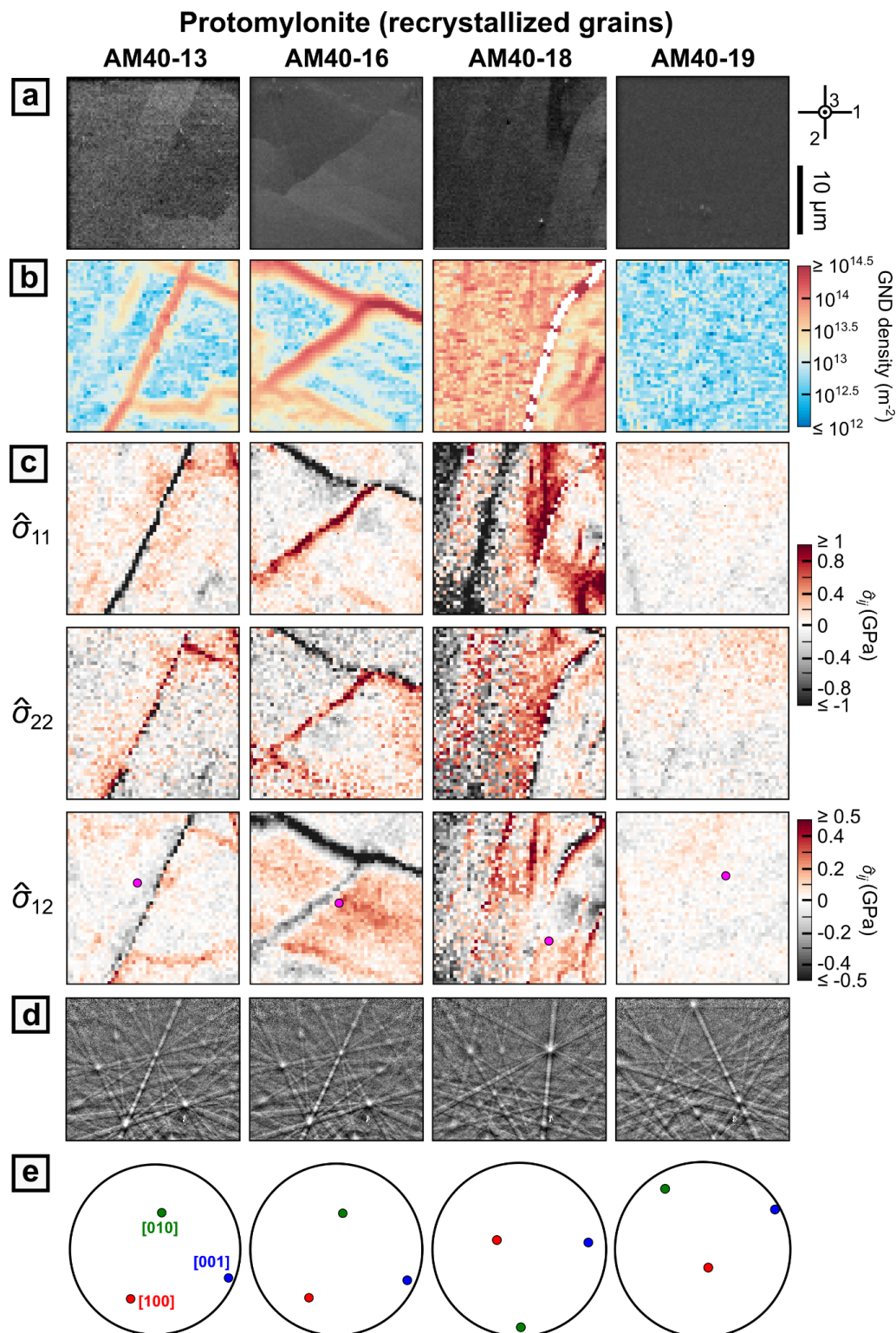


Figure 9. HR-EBSD data collected from the recrystallized grains within protomylonite (AM40). Descriptions of (a)–(e) are the same as those for Fig. 7.

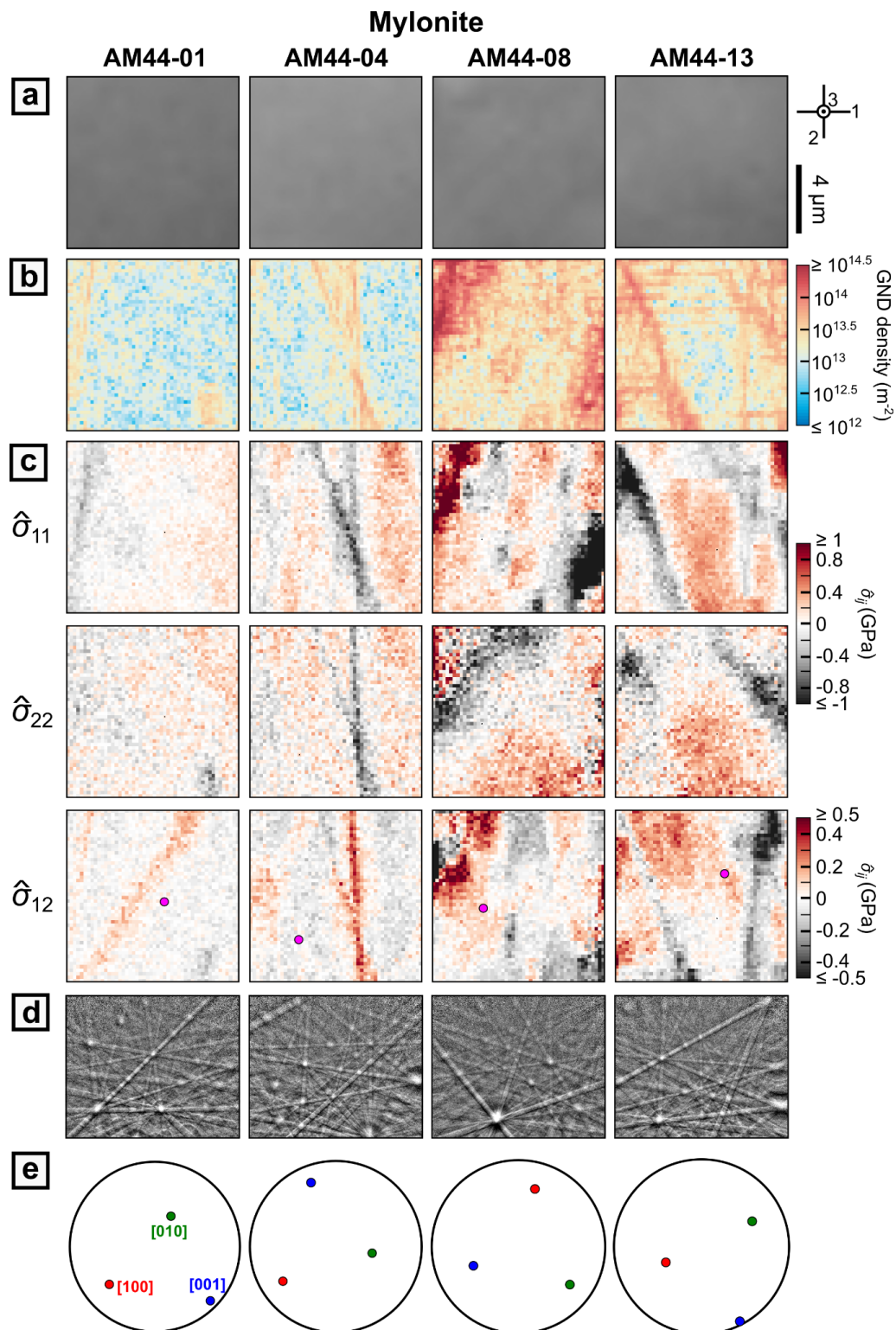


Figure 10. HR-EBSD data collected from the mylonite (AM44). Descriptions of (a)–(e) are the same as those for Fig. 7.

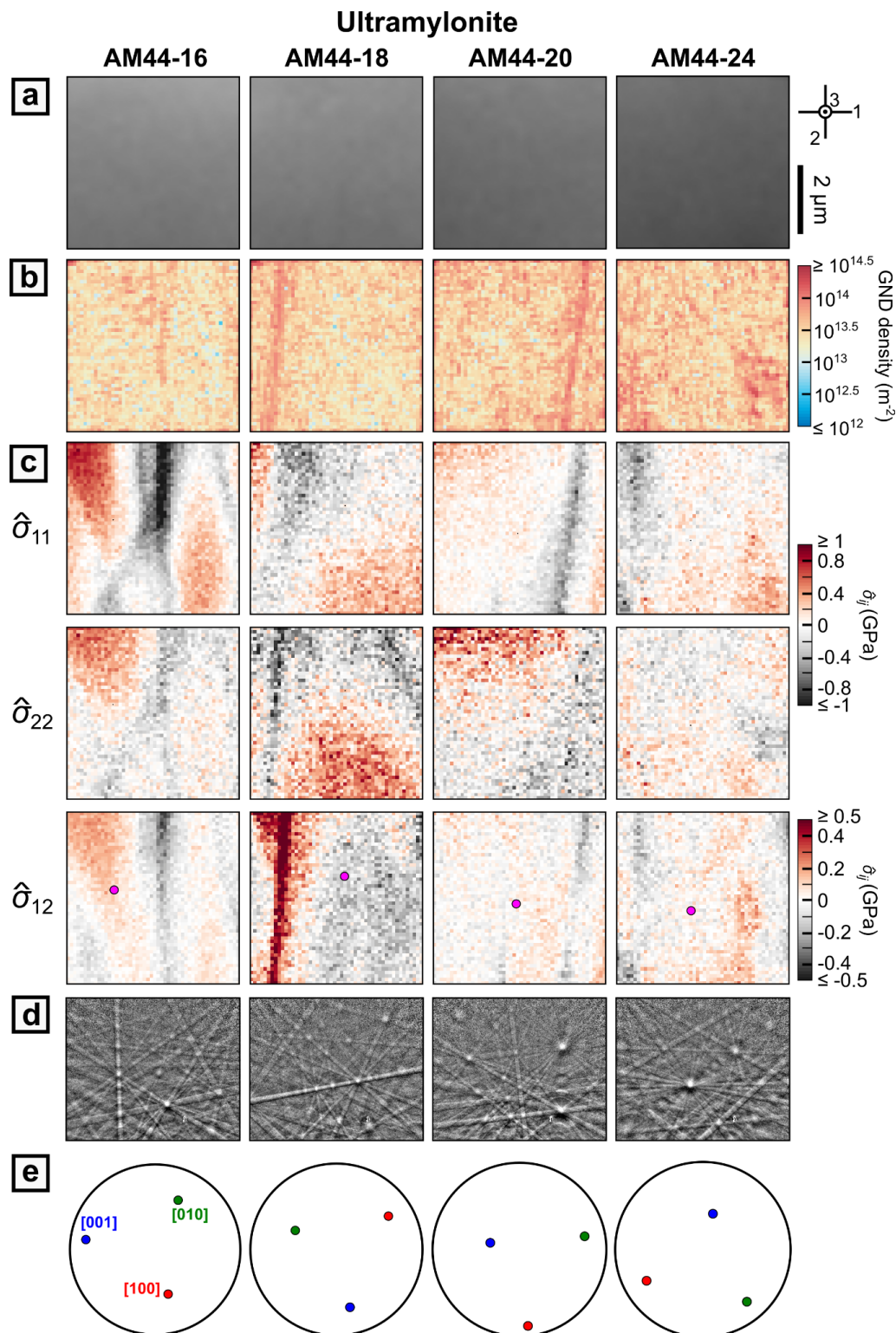


Figure 11. HR-EBSD data collected from the ultramylonite (AM44). Descriptions of (a)–(e) are the same as those for Fig. 7.

4.3 Statistics of intragranular stress heterogeneity

Relative to the distribution of $\hat{\sigma}_{12}$ in the wall rock, the distributions in the protomylonite, mylonite, and ultramylonite are broader (Fig. 12a). Specifically, the 99th percentile of the distribution of $\hat{\sigma}_{12}$ in the wall rock is 150 MPa, whereas those of the more deformed domains are in the range 250–370 MPa (Fig. 12b).

We explore the forms of the distributions of $\hat{\sigma}_{12}$ using normal probability plots and the restricted second moment. Normal probability plots of $\hat{\sigma}_{12}$ (Fig. 12b) exhibit two main characteristics that are common across the data from all domains. First, the distributions of $\hat{\sigma}_{12}$ at magnitudes less than approximately 100 MPa fall on straight lines, indicating that they are well described by normal distributions. Second, in contrast, the tails of the distributions at greater stress magnitudes deviate from the straight lines and extend to greater stress magnitudes than predicted by the normal distributions fit to the data at low stress magnitudes indicated by dashed lines. However, compared with the wall rock, the protomylonite, mylonite, and ultramylonite show a greater degree of the deviation of tails away from normal distributions (Fig. 12b). We plot the restricted second moment as a function of the natural logarithm of $|\hat{\sigma}_{12}|$ for all domains in Fig. 12c. The data from each domain follow linear relationships at stress magnitudes greater than ~200 MPa. However, for protomylonite, mylonite, and ultramylonite the straight-line proportion has a gradient that is 4–6 times greater, and extends to greater stress magnitudes, compared with that of the wall rock (Fig. 12c).

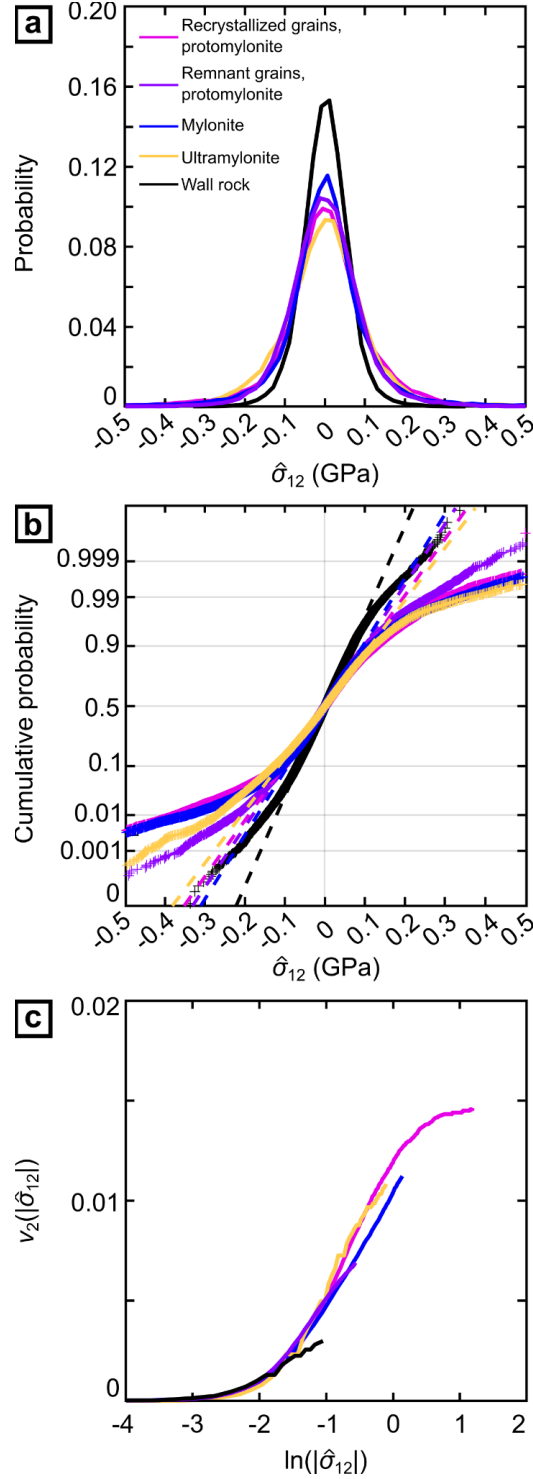


Figure 12. Statistical analyses of mean-centred in-plane residual shear-stress component, $\hat{\sigma}_{12}$, for different domains. (a) Histograms showing probability distributions of $\hat{\sigma}_{12}$. (b) Normal probability plots of $\hat{\sigma}_{12}$. (c) Restricted second moment as a function of the natural logarithm of $|\hat{\sigma}_{12}|$.

4.4 Decorated dislocations

Figure 13 presents foreshattered electron images of decorated dislocations in olivine from wall rock and protomylonite samples. In both rocks, dislocations are bright dots or short lines with varying orientations, and some occur in subgrain boundaries expressed as linear arrays of closely spaced dislocations. Where dislocations intersect the sample surface at oblique angles, they typically appear as short lines (Figs. 13b and 13f). Both straight and locally curved dislocations are evident, and in some grains they are arranged into alternating bands of higher and lower density across micrometre length scales (Figs. 13c and 13e). Dislocation densities are greater in the protomylonite (Figs. 13c–13f) than in the wall rock (Figs. 13a and 13b). The average density of dislocations is $1.7 \times 10^{11} \text{ m}^{-2}$ in the wall rock, corresponding to an estimated bulk differential stress of 28 MPa, and is $1.1 \times 10^{12} \text{ m}^{-2}$ in protomylonite, corresponding to an estimated bulk differential stress of 105 MPa.

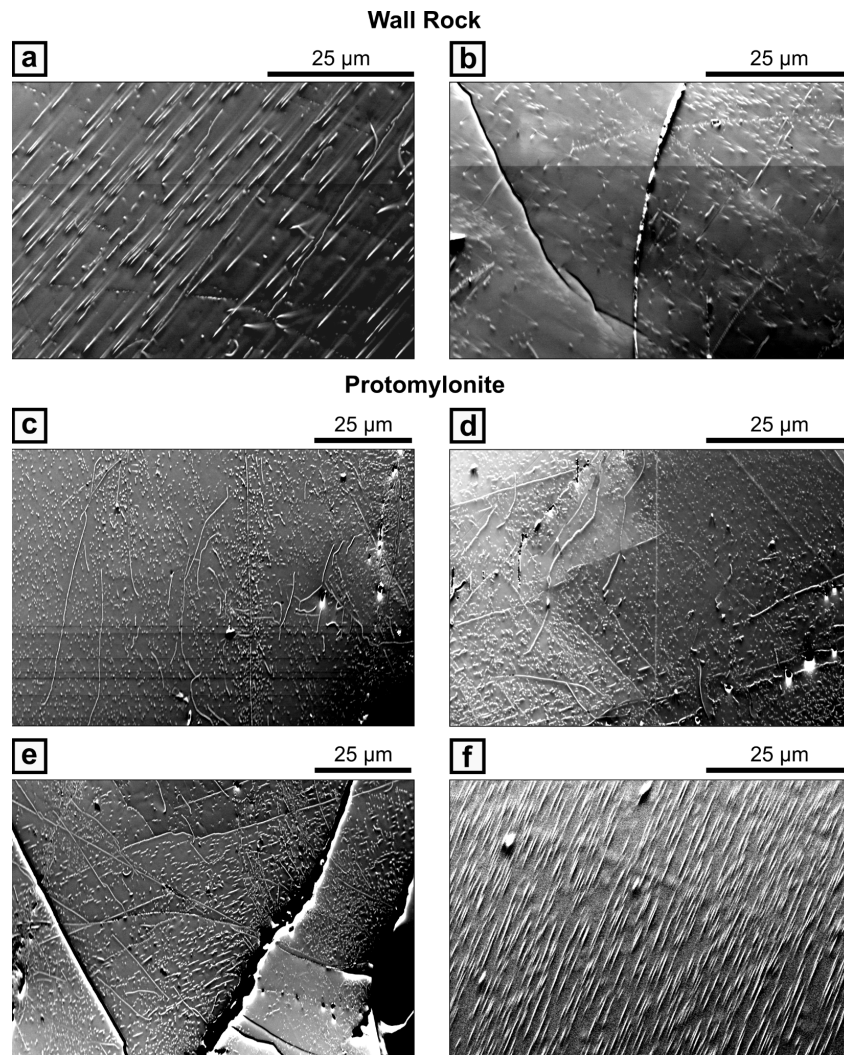


Figure 13. Foreshattered electron images of dislocations in olivine decorated by oxidation.

5. Discussion

5.1 Inferences of deformation mechanisms and recrystallization processes

5.1.1 Dislocation-mediated deformation

Dislocation microstructures are pervasive in olivine grains in the Finero peridotite, ranging from the relatively undeformed wall rock through protomylonite and mylonite to ultramylonite. Here, we use *dislocation-mediated deformation* as a term for intracrystalline deformation in which plastic strain is generated primarily by dislocation glide. Depending on stress, temperature and grain size, dislocation-mediated deformation may occur through low-temperature plasticity controlled by lattice resistance, high-temperature power-law dislocation creep, or grain-size-sensitive deformation in which dislocation glide is coupled to recovery facilitated by grain-boundary diffusion (Breithaupt et al., 2023). The latter corresponds to the rheological behaviour previously described as dislocation-accommodated grain-boundary sliding (disGBS) (Breithaupt et al., 2023; Hansen et al., 2019, 2021; Hirth & Kohlstedt, 2003; Warren & Hirth, 2006).

Oxidation decoration reveals abundant dislocations in the wall rock and protomylonite (Section 4.4) and HR-EBSD data reveal GNDs in each microstructural domain (Section 4.2.1). Deformation lamellae are also common in the protomylonite (Matysiak & Trepmann, 2015). Likewise, visible-light and EBSD analyses (Section 4.1) reveal that subgrain boundaries are present across all domains and reflect recovery processes, by which dislocations rearrange to form lower-energy configurations (Urai et al., 1986). In the wall rock, misorientation axes of subgrain boundaries parallel to [001] indicate activity of the (010)[100] slip system during early, high-temperature deformation (Section 4.1.1). In contrast, alongside misorientation axes parallel to [001], the protomylonite and mylonite also have misorientation axes of subgrain boundaries parallel to [010] (Section 4.1.2 and 4.1.3). Taken together with CPOs that have [100] approximately parallel to the lineation and [001] approximately normal to the foliation (i.e., E-type CPOs, Section 4.1.2 and 4.1.3), these observations indicate activity of the [100](001) slip system during deformation of the mylonitic rocks.

Together, these observations support the deformation history proposed by Matysiak & Trepmann, (2015). This model suggests that formation of the mylonitic series in the Finero peridotite involved an initial phase of low-temperature plasticity (i.e., dislocation glide) during transient seismic stress pulses, followed by more prolonged dislocation creep during the subsequent postseismic stress relaxation.

5.1.2 Dynamic recrystallization

Olivine grains within protomylonite, mylonite, and ultramylonite have average grain sizes that are smaller than those of the wall rock (Figs. 5 and 6; Section 4), indicating the operation of grain-size reduction processes (Biedermann et al., 2020; Matysiak & Trepmann, 2015). Olivine grains within the protomylonite have (1) a core-and-mantle structure, i.e., networks of small, recrystallized grains interlocking with large, remnant grains, (2) subgrains, which have sizes similar to those of recrystallized grains, at the margins of remnant grains, and (3) similar CPO between recrystallized and remnant grains (Figs. 5f–5l; Section 4.1.2). Previous studies on rock-forming minerals, ice, and metals suggest that core-and-mantle structures, networks of subgrains, and similar CPO between recrystallized and remnant grains are common microstructural signals of subgrain rotation recrystallization (Fan et al., 2020; Halfpenny et al., 2006; Poirier & Nicolas, 1975; Ponge & Gottstein, 1998; Urai et al., 1986; White, 1976). Olivine grains with irregular grain boundaries are common in the protomylonite (Figs. 5, 6; Sections 4.1.2, 4.1.3; Matysiak & Trepmann, 2015). Irregular grain boundaries are commonly interpreted to result from strain-induced grain-boundary migration (Fan et al., 2020; Hirth & Tullis, 1992; Karato, 1988; Stipp et al., 2010). During strain-

induced grain-boundary migration, grain boundaries are driven by intergranular heterogeneity in strain energy to migrate into the adjacent grain with the greatest strain-energy density (Urai et al., 1986). These interpretations of dynamic-recrystallization processes inferred from microstructural observations align with previous studies conducted on the Finero peridotite (Kenkmann & Dresen, 2002; Matysiak & Trepmann, 2015).

5.1.3 Grain-size sensitive deformation mechanisms

The sequence of deformation microstructures suggests a contribution of grain-boundary sliding (GBS) in the finer-grained domains. Olivine within both the wall rock and protomylonite exhibits well-defined CPO (Figs. 5e, 5l; Sections 4.1.1, 4.1.2). For the protomylonite, recrystallized olivine grains have a weaker CPO intensity (here defined as a lower value of M-index) compared to the remnant grains (Fig. 5l; Section 4.1.2). Moreover, olivine grains within the mylonite and ultramylonite have CPOs that are close to random, in contrast to the protomylonite with a well-defined CPO (Figs. 5l, 6e, 6i; Sections 4.1.2, 4.1.3). These observations indicate that additional processes and/or deformation mechanisms are progressively more involved in the deformation of the increasingly fine-grained portions of the protomylonite, mylonite, and ultramylonite. Previous studies on minerals have observed recrystallized daughter grains with weaker CPO compared to neighbouring parent grains, and this relationship is commonly interpreted to be the result of subgrain rotation followed by further reorientation during deformation involving GBS (Bestmann & Prior, 2003; Fan et al., 2020; Fliervoet et al., 1997; Jiang et al., 2000; Storey & Prior, 2005; Warren & Hirth, 2006). Deformation that involves GBS and progresses to high strains must also involve motion of point defects (diffusion creep) and/or dislocations (dislocation accommodated grain-boundary sliding, disGBS) to allow shape changes of the grains (Raj & Ashby, 1971). Grain shape and SPO data are not sufficient to distinguish between these additional mechanisms of intracrystalline deformation, because they both can produce grains with elongate shapes and a strong SPO, as is the case in this study (Figs. 5–6; Sections 4.1.2, 4.1.3; Ashby & Verrall, 1973; Wheeler, 2010). However, diffusion creep often results in a lack of CPO (Ashby & Verrall, 1973; Boullier & Gueguen, 1975), whilst disGBS usually produces a CPO (Hansen et al., 2012). Therefore, the lack of olivine CPO within ultramylonite suggests the operation of diffusion creep in these fine-grained domains (Warren & Hirth, 2006). Overall, the microstructures suggest that progressive grain-size reduction by subgrain-rotation recrystallization caused a progressive switch from dislocation-mediated deformation in the coarser-grained material to an increasing contribution of diffusion creep in the finer-grained domains. Shear zones commonly evolve through different deformation episodes, where deformed remnant original grains, i.e., porphyroclasts, may rather reflect a recovered initial stage of dislocation-mediated deformation, where the newly formed recrystallized grains record a subsequent stage (Matysiak & Trepmann, 2015). Nonetheless, the intragranular substructures observed in the mylonite and ultramylonite, as revealed by the EBSD and HR-EBSD data, indicate that dislocation-mediated deformation still made a partial contribution to strain accommodation in these domains.

The CPO and HR-EBSD observations are therefore complementary. CPO reflects the integrated orientation evolution of a grain population over finite strain, whereas HR-EBSD records local lattice curvature, dislocation structures, and their associated internal stresses within individual grains. A weak or absent CPO does not require a complete absence of dislocation activity; rather, it indicates that dislocation-mediated deformation was not sufficiently dominant to generate or preserve a strong CPO. In the finer-grained domains, diffusion creep may therefore have accommodated a substantial fraction of the total strain, while dislocation-mediated deformation made a subordinate but detectable contribution. The stored dislocation structures and long-range internal stresses revealed by HR-EBSD may record dislocation activity that

operated concurrently with, or at a different stage from, diffusion creep. These constraints enable us to further explore the contribution of dislocation-mediated deformation and diffusion creep to the deformation of the Finero peridotite in the following sections.

5.2 Spatial distributions of geometrically necessary dislocations and stress heterogeneity

Olivine grains within protomylonite, mylonite, and ultramylonite generally exhibit greater heterogeneity of GND density and residual stress compared to those in wall rock (compare Figs. 8–11 with Fig. 7; Section 4.2). This observation suggests that the dislocation structures, as indicated by areas with elevated GND densities (Figs. 8–11), in protomylonite, mylonite, and ultramylonite result from a distinct phase of deformation and are not inherited from the wall rock. Areas containing clear dislocation structures typically have elevated GND densities that are up to $\sim 10^{14} \text{ m}^{-2}$ (Figs. 8–11); these magnitudes are close to those of datasets collected from other natural samples (i.e., Oman-UAE ophiolite; Wallis et al., 2022), but are less than those of datasets collected from experimentally deformed samples ($< \sim 10^{15} \text{ m}^{-2}$) (e.g., Wallis et al., 2020, 2021).

We note that our measured GND densities ($10^{13.5}–10^{14} \text{ m}^{-2}$) are greater than the average dislocation densities ($10^{12}–10^{13} \text{ m}^{-2}$; Section 4.4 and Matysiak & Trepmann (2015)) measured by oxidation decoration. Following Wallis et al. (2022) and Wiesman et al. (2024), we provide two explanations to rationalise the difference between GND density and decorated-dislocation density:

(1) Decorated dislocations and GND density measure different populations of dislocations. The densities calculated from decorated dislocations are those of free dislocations within subgrain interiors, whereas the densely packed dislocations along subgrain boundaries cannot be reliably measured. In contrast, GND densities determined from orientation data include dislocations in subgrain boundaries due to their influence on lattice curvature. Moreover, GND measurements can reveal dislocations that do not thread to sample surfaces during decoration (Wallis et al., 2016).

(2) GND maps have a noise background. Previous studies suggest GND density measured from undeformed silicon and olivine single crystals using HR-ESBD technique is on the order of $10^{12}–10^{13} \text{ m}^{-2}$ at step sizes of $0.1–1 \mu\text{m}$ as used in this study (Jiang et al., 2013b; Ruggles et al., 2016; Wallis et al., 2019; Wallis et al., 2016). Dislocations within subgrain interiors can sometimes be resolved by HR-EBSD but they are often below the noise level. We suggest the apparent GND density below 10^{13} m^{-2} (corresponding to the blue colour in Figs. 7b–11b) in our dataset mostly results from noise. Areas with GND densities above 10^{13} m^{-2} (corresponding to the brown to red colour in Figs. 7b–11b) are robust measurements above the noise level.

The spatial distributions of GND density and stress heterogeneity observed from olivine grains within the Finero peridotite are similar to those obtained from naturally and experimentally deformed samples reported in previous studies (Wallis et al., 2017, 2021, 2022; Wiesman et al., 2024). Residual stress is heterogeneous in deformed domains, varying in magnitude by hundreds of megapascals over distances of a few micrometres (Figs. 8c–11c; Section 4.2.2). Moreover, areas of localised, elevated stresses are generally spatially correlated to structures with elevated GND densities (Figs. 8–11). Together, these observations suggest that dislocations are the main source of intragranular stress heterogeneity (Britton et al., 2013; Britton & Wilkinson, 2012b; Hansen et al., 2021; Wallis et al., 2020, 2021) and we explore this interpretation in more detail below.

5.3 Origins of stress heterogeneity

We apply statistical analysis to infer the origin of stress heterogeneity observed in Finero peridotite (Fig. 12; Section 4.3). The probability distribution of $\hat{\sigma}_{12}$ is broader for olivine within protomylonite, mylonite, and ultramylonite compared to that within the wall rock (Fig. 12a; Section 4.3). This observation is similar to those from previous experimental studies, where deformed olivine aggregates have greater stress heterogeneity compared with undeformed samples that underwent the same decompression and cooling (Wallis et al., 2020, 2021). The shape of the normal probability plot (Fig. 12b) can test how close the distribution of stress (Fig. 12a) is to a normal distribution (Section 3.3.1). Whilst the central portions of the stress distributions fall on a straight line indicating that they are normally distributed, all microstructural domains exhibit tails that deviate from a straight line (i.e., depart from a normal distribution) and extend to greater stress magnitudes. This effect is widely observed in minerals and metals containing dislocations (e.g., Jiang et al., 2013a; Kalácska et al., 2017; Wallis et al., 2021; Wilkinson et al., 2014). Analysis of the restricted second moment of a stress distribution (Fig. 12c) further assesses the source of intragranular stress by focusing on the form of the tails at high stress magnitudes (Section 3.3.1). On the plot of restricted second moment as a function of the natural logarithm of $|\hat{\sigma}_{12}|$, $\ln(|\hat{\sigma}_{12}|)$, the stress distributions fall on straight lines at stress magnitudes greater than ~ 200 MPa (Fig. 12c; Section 4.3). This observation indicates that the high-stress portions of the probability distributions, $P(\sigma)$, decay with $|\sigma|^{-3}$; this form is expected of stress fields of a population of dislocations (Section 3.3.1; Kalácska et al., 2017; Wilkinson et al., 2014). The slope of the straight-line portion is proportional to the total dislocation density (Kalácska et al., 2017; Wilkinson et al., 2014). The straight-line proportions for olivine within protomylonite, mylonite, and ultramylonite are 4–6 times steeper than that of the wall rock (Fig. 12c; Section 4.3). Therefore, this difference in slope is consistent with the factor of six difference in the density of decorated dislocations between the wall rock and protomylonite (Section 4.4). The curve for recrystallized grains in the protomylonite departs from a straight line above ~ 1.2 GPa. This phenomenon is commonly observed and has been inferred to result from the spatial averaging of stresses with the highest magnitudes close to the cores of dislocations over the finite interaction volume illuminated by the electron beam in the SEM (Kalácska et al., 2017; Wallis et al., 2021, 2022). Lastly, correlation of the spatial distributions of GND density and intragranular stress heterogeneity also indicates that dislocations are a key source of stress heterogeneity (Section 5.2). As the typical length scale of stress heterogeneity in the mylonitic domains is on the order of a few micrometres, whilst the average dislocation spacing is less than a micrometre, the stress heterogeneity constitutes long-range internal stress, by which the dislocations undergo long-range elastic interactions.

5.4 Implications for transient creep during post-seismic deformation

5.4.1 Deformation-mechanism map

Matysiak & Trepmann (2015) inferred that many of the microstructural elements recorded in the mylonitic rocks of the Finero peridotite formed as a result of sudden increases in differential stress and subsequent relaxation related to seismic cycles during exhumation. Our new data, which focus on characterising the intragranular microstructure of olivine, reveals extensive dislocation substructure as well as long-range internal stresses that can be attributed to these dislocations. These observations evidence dislocation-mediated deformation involving backstress and kinematic hardening (Hansen et al., 2019, 2021; Wallis et al., 2020, 2021), and therefore support the inclusion of dislocation-generated backstress in a constitutive description of dislocation-mediated deformation. In this section, we quantify the potential contribution of dislocation-mediated deformation to both the steady-state and transient rheological behaviour, comparing

a model of this deformation process by Breithaupt et al. (2023) against a description of diffusion creep. We describe the ingredients of this model below; a more detailed exposition is given in Sections S3.4–3.7 of the Supporting Information.

Breithaupt et al. (2023) develop their model of dislocation-mediated deformation in two stages; firstly fitting experimental data for dislocation creep and secondly fitting data gathered in both dislocation creep and disGBS regimes. In the former case, they fitted parameters describing dislocation recovery facilitated by pipe diffusion, characterising grain-size-insensitive dislocation creep. In the latter case, which they term the full model, they fitted parameters describing dislocation recovery facilitated by both pipe diffusion and grain-boundary diffusion using a broader dataset, which can also predict a grain-size sensitive dislocation-mediated rheological behaviour. In both cases, they also constrained a flow law for dislocation glide that incorporates the effect of long-range dislocation interactions. In the deformation-mechanism maps and constitutive calculations presented below, we use their full model. In their analysis comparing model predictions to steady-state experimental data, their formulation of recovery facilitated by pipe diffusion received the greatest experimental support, with recovery facilitated by grain-boundary diffusion being less well constrained. The transient rheological behaviour predicted by their model is most sensitive to the formulation of dislocation glide, which was tested in detail by Hansen et al. (2021). Thus, we use their full model as an exploratory implementation to probe the mechanical behaviour of continental lithospheric mantle during transient creep following sudden increases of stress imposed by earthquakes, because it couples dislocation glide, evolving dislocation density and backstress, and dislocation recovery within a single framework.

We pair this description of dislocation-mediated deformation with a model for the steady-state and transient rheological behaviour of diffusion creep in olivine (Section S3.6 of the Supporting Information). Warren & Hansen (2023) note that there is considerable uncertainty regarding flow laws for diffusion creep of olivine; we follow their advice and use the steady-state flow law for diffusion creep fitted by Hansen et al. (2011) for internal consistency with the model of dislocation-mediated deformation, as both were fitted using common data. To model the transient rheological behaviour of diffusion creep, we use the extended Burgers model (Faul & Jackson, 2015; Jackson & Faul, 2010; Rudge, 2026). The extended Burgers model is underpinned by a relaxation spectrum, which describes the fractional compliance that relaxes at different timescales. The shape of this relaxation spectrum, typically modelled as a power law, plays a key role in controlling the transient viscosity. Our understanding of the relaxation spectrum of transient diffusion creep has been substantially revised in recent years. Whereas in the model of Jackson & Faul (2010) transient creep had a weaker grain-size sensitivity than steady-state diffusion creep, more recent experimental (Qu, 2022) and theoretical work (Rudge, 2026) suggests that they instead have the same grain-size sensitivity. This distinction matters when extrapolating to larger grain sizes than those used in laboratory studies, as we do here. We follow the parameterisation of Rudge (2026) when constructing the diffusion-creep relaxation spectrum for our model, which relates key parameters characterising the relaxation spectrum to the steady-state flow law. A key parameter not constrained by Rudge (2026) is the lower limiting timescale of the relaxation spectrum, which we assume to be equal to the timescale of EAGBS. We utilise an expression from Ashby (1972) to parameterise the intrinsic viscosity of grain-boundary sliding in terms of the steady-state flow law for diffusion creep. This expression likely provides a lower bound on this viscosity and therefore the lower limiting timescale. As a result, our choice maximises the initial transient strain rate following step changes in stress and thus provides an upper bound on the initial contribution of diffusion creep to the transient rheological behaviour (Sections S3.5 and S3.6 of the Supporting Information). At

intermediate timescales, the transient rheological behaviour is relatively insensitive to the choice of lower limiting timescale, thus reducing the impact of uncertainty in this parameter (Section S3.6 of the Supporting Information). In addition to models of dislocation-mediated deformation and diffusion creep, we also incorporate a model of EAGBS, using the intrinsic grain-boundary viscosity that we calculate and the dissipation strength constrained by the recent simulations of (Li & Rudge, 2026; Section S3.5 of the Supporting Information).

It is noteworthy that all of the transient deformation mechanisms that we consider (dislocation-mediated deformation, diffusion creep and EAGBS) have memory that can be parameterised in terms of backstresses or internal stresses (Sections S3.4, S3.5, and S3.7 of the Supporting Information). In the case of dislocation-mediated deformation, the backstress corresponds to the long-range elastic interactions among dislocations that we have characterised above. In the case of EAGBS and diffusion creep, the internal stresses correspond to the grain-scale stress heterogeneity. As transient diffusion creep necessarily involves a spectrum of internal stress, we present internal stresses averaged over the full spectrum. We also consider effective stresses, defined to be the difference between the applied stress and the relevant backstress or internal stress. The effective stresses drive transient creep and thereby control transient strain rates.

Our analysis assesses the extent to which dislocation-mediated deformation contributes to the transient response following a stress increase during a seismic event relative to other transient deformation mechanisms, as well as how this contribution evolves with time. In this respect, it extends the analysis of Wallis et al. (2022), who considered only the instantaneous change in viscosity controlled by dislocation-mediated deformation following a step change in stress. The calculations are used to explore order-of-magnitude changes in viscosity during postseismic deformation using the latest models under development, whilst acknowledging that the parameter values and predicted strain rates can be refined as future experimental results are generated. The analysis that follows utilises observations from the Finero peridotite as a guide to input values but is intended to provide a simple exposition of the basic characteristics of transient creep in such contexts more generally.

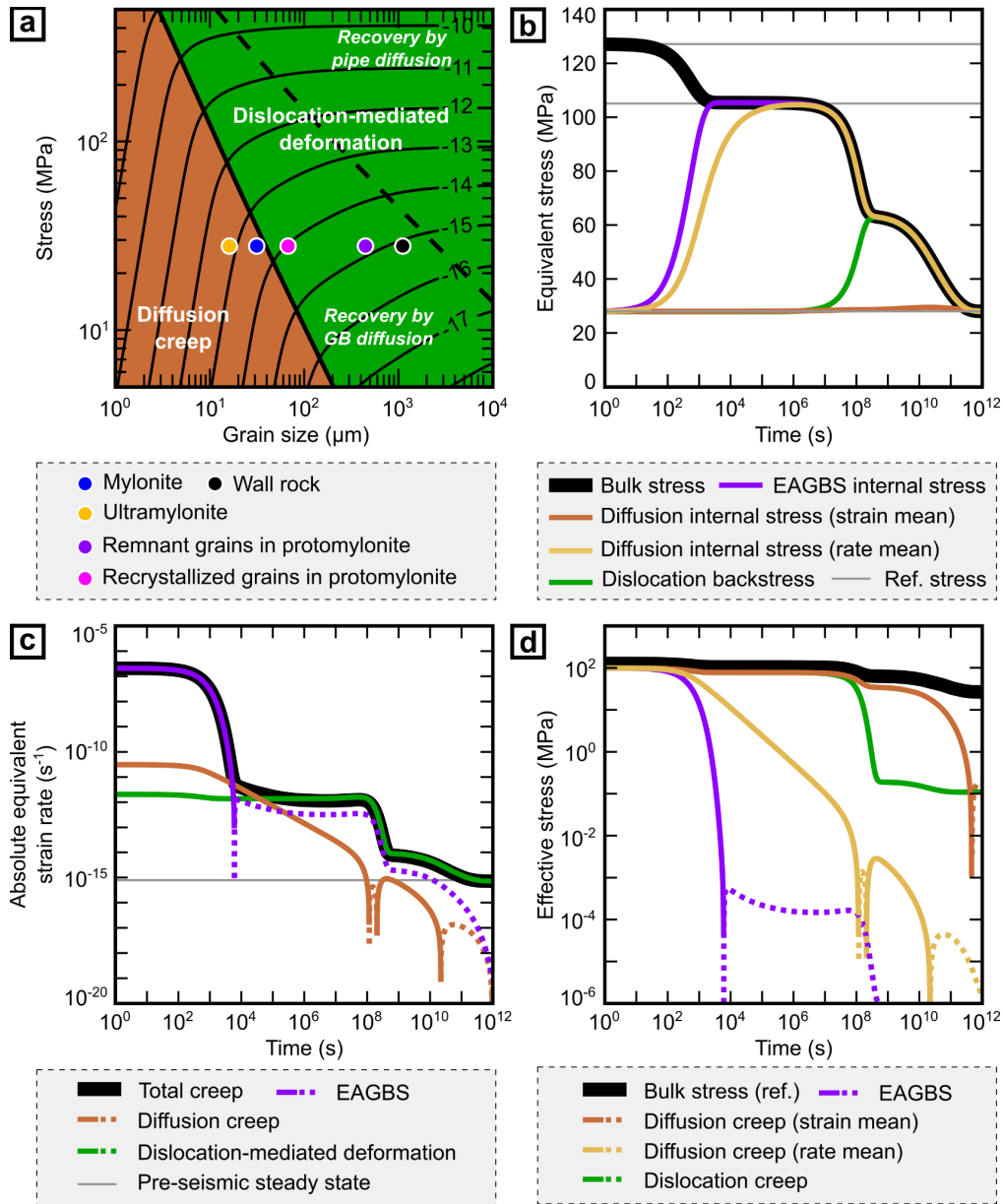


Figure 14. Exploratory predictions obtained from the transient-creep model incorporating transient dislocation-mediated deformation, transient diffusion creep and EAGBS (Sections S3.4-S3.8 of the Supporting Information). **(a)** Deformation-mechanism map at steady-state before the seismic event. The thick solid black line represents the deformation-mechanism boundary, where the strain rates contributed by dislocation-mediated deformation and diffusion creep are equal. Points represent stress and grain size for ultramylonite, mylonite, recrystallized grains in protomylonite, remnant grains in protomylonite, and wall rock. **(b)** The transient evolution of equivalent stress as a function of time during stress relaxation after the seismic event. Also plotted are the backstress from dislocation-mediated deformation, and internal stresses from EAGBS and diffusion creep. The horizontal thin grey lines mark the preseismic background stress (28 MPa), the piezometric reference used as the approximate stress level after rapid EAGBS relaxation (105 MPa), and the imposed stress immediately after seismic loading (127 MPa). As internal stresses in diffusion creep form a spectrum, averages over this spectrum are presented. The averages we

present correspond to the degree of strain accommodated by transient diffusion creep and the average internal stress that sets the net transient diffusion-creep strain rate. **(c)** The transient evolution of equivalent strain rate as a function of time for dislocation-mediated deformation, EAGBS, and diffusion creep. Also plotted is the total strain rate calculated across all mechanisms (thick black line). Negative strain rates are plotted using dotted lines. The horizontal thin grey line corresponds to the equivalent net strain rate inferred to operate before the seismic event, i.e., the steady-state strain rate before the seismic event. **(d)** The transient evolution of effective stress for each mechanism as a function of time. Negative effective stresses are plotted using dotted lines. For diffusion creep, the two averages over the spectrum of diffusion creep internal stresses are plotted. The total equivalent stress is plotted as a thick black line.

We use a deformation-mechanism map (De Bresser et al., 2001; Frost & Ashby, 1982) to constrain the relative contributions of dislocation-mediated deformation and diffusion creep to the total strain rate at steady state (Fig 14a). To construct this map, we estimate the strain rates associated with dislocation-mediated deformation and diffusion creep for each pair of stress, σ , and grain size, d . We choose a temperature, T , of 800°C, which is representative of that during deformation of the Finero peridotite (Altenberger, 1995; Lu et al., 1997). All stress values shown in the deformation-mechanism map and transient-creep figures are bulk equivalent stresses, which equal differential stresses in the axial calibration geometry. The transient calculation is performed in simple shear using the conversions in Section S3.3 of the Supporting Information. On the map, we superpose the grain size-stress pairs for ultramylonite (yellow), mylonite (blue), recrystallized grains in protomylonite (pink), remnant grains in protomylonite (purple), and wall rock (black). We assume that, at steady state, the wall rock and deformed domains are subject to a common background stress. For mylonite and ultramylonite, the mixing of olivine with other phases (Figs. 6a, 6f; Section 4.1.3) has potential to pin grain boundaries and maintain a grain size that is less than the piezometric value (Herwegh et al., 2011). The occurrence of this effect is supported by the difference between the 67 μm median size of recrystallized olivine in the protomylonite, which has mostly monophase grain boundaries and the median grain sizes of 31 μm and 16 μm of olivine in the mylonite and ultramylonite, which have abundant interphase boundaries (Section 4.1). Using the grain-size piezometer developed for pure olivine (Qi et al., 2021), these grain sizes correspond to differential stresses of approximately 20, 40, and 80 MPa for the recrystallized olivine in the protomylonite, mylonite, and ultramylonite. Consequently, the grain-size piezometer may overestimate the stress in mylonite and ultramylonite. We suggest that the pinning effect is less significant in protomylonite as most of the recrystallized olivine grains are in contact with olivine (Fig. 5f; Section 4.1.2). Thus, the ~20 MPa stress estimated for the recrystallized olivine in the protomylonite provides a more appropriate grain-size-based estimate of the background stress than the higher values obtained for the mylonite and ultramylonite. Independently, the decorated-dislocation density in the relatively undeformed wall rock yields a similar background-stress estimate of ~28 MPa (Section 4.4). We therefore use a common background stress of 28 MPa as a controlled comparison among the representative domain parameter sets.

Under this common-stress assumption, the positions of the samples on the deformation mechanism map allow us to assess the relative contributions of dislocation-mediated deformation and diffusion creep at steady state. We note that the deformation-mechanism boundary (thick solid black line; Fig. 14a) does not represent a discontinuous switch between mutually exclusive mechanisms. It marks the conditions at which the steady-state strain rates contributed by diffusion creep and dislocation-mediated deformation are equal. Both mechanisms contribute in the vicinity of the boundary, although their relative proportions differ on either side. The stress-grain size data of wall rock (black) and remnant olivine grains in protomylonite

(purple) lie to the right of the deformation-mechanism boundary, indicating a significant contribution of dislocation-mediated deformation. The stress-grain size data of olivine grains in ultramylonite (yellow) lie to the left of the deformation-mechanism boundary, indicating a significant contribution of diffusion creep. The stress-grain size data for mylonite (blue) and recrystallized olivine grains in protomylonite (pink) lie close to the deformation-mechanism boundary, with the former slightly within the diffusion-creep field and the latter slightly within the field of dislocation-mediated deformation. Thus at steady state, dislocation-mediated deformation remains an important contributor to deformation; however, its relative contribution decreases with decreasing grain size, consistent with the microstructural evidence (Section 5.1.3). This interpretation is further supported by (1) the observations of dislocations and dislocation substructures (i.e., subgrain boundaries and structures with elevated GND densities) (Figs. 5h, 6c, 6h, 8b–11b, 13; Sections 5.1.2, 5.2), and (2) the interpretation from the statistics of stress heterogeneity that dislocations are a key source of intragranular stress heterogeneity in all microstructural domains (Section 5.3).

Grain size evolves with progressive deformation, and the microstructures indicate grain-size reduction associated with subgrain-rotation recrystallization and phase mixing (Section 5.1.3). Under the common-background-stress assumption, this evolution corresponds to movement towards smaller grain sizes on the deformation-mechanism map, with an increasing relative contribution from diffusion creep. The sequence of plotted grain sizes therefore illustrates a possible evolution path with accumulated strain over longer timescales. This longer-term grain-size evolution is distinct from the individual transient-creep event modelled below, during which grain size is held constant.

5.4.2 The evolution of viscosity during transient creep

We use the coarse grain size of the wall rock and remnant grains of the protomylonite as an example to demonstrate the predicted viscosity evolution during the first instance of transient creep following a nearby seismic event, i.e., a sudden increase of applied stress. We choose a temperature of 800°C (Section 5.4.1) and assume that the grain size of the wall rock ($d = 1100 \mu\text{m}$) does not change during a single transient-creep event due to the small strains involved.

Figure 14b–d presents the model predictions, with the details of the calculation provided in Section S3.8 of the Supporting Information. In short, we assume that the seismic event loads the stress beyond its steady-state value. After the seismic event, we allow stress relaxation to occur. To simulate the subsequent relaxation, we maintained a constant externally imposed total strain rate corresponding to the pre-pulse steady-state strain rate, while allowing the strain rate produced by each deformation mechanism, stress, and all backstresses/internal stresses to evolve. We perform this calculation with a simple-shear geometry and convert all shear stresses to equivalent stresses for the figures that we present here.

At steady state before a seismic event, we assume that the equivalent stress is $\sim 28 \text{ MPa}$, estimated from the dislocation density of wall rock (Section 4.4; Fig. 14b). We assume that the internal stresses associated with diffusion creep and EAGBS are equal to the initial stress prior to the seismic event, consistent with the assumption that prior to the seismic event the system was at steady-state. In contrast, for dislocation-mediated deformation, the backstress is not expected to be equal to the applied stress at steady-state. Our steady-state calculation indicates that the initial (equivalent) backstress was less than the prior equivalent applied stress by about 10^{-1} MPa , i.e., that the effective stress driving dislocation-mediated deformation before the seismic event was about 10^{-1} MPa . We use this inferred dislocation backstress as an initial condition in our calculation of transient creep. We imposed an instantaneous change in applied equivalent stress to 127 MPa , to simulate a sudden stress increase following a seismic event. This initial stress was

chosen so that the peak stress following rapid EAGBS relaxation was approximately 105 MPa (Section S3.8 of the Supporting Information). This latter value is estimated from the decorated-dislocation density in the protomylonite (Section 4.4), which is inferred to preserve the high-stress stage, after EAGBS relaxation, most clearly. In particular, remnant grains in this domain contain slip bands and kinks indicative of low-temperature plasticity that formed during the stress pulse and were not completely erased during subsequent postseismic relaxation by dislocation-mediated deformation (Matysiak & Trepmann, 2015). If the decorated-dislocation density in these remnant grains was reduced by later static annealing, 105 MPa represents a lower bound on the peak stress recorded by the protomylonite.

The results of our transient-creep calculation are presented in Figures 14b–d. In Figure 14b, we examine the relaxation of equivalent stress as a function of time. We also plot the evolution of the backstress arising in dislocation-mediated deformation, as well as the internal stresses associated with EAGBS and diffusion creep. In the case of diffusion creep, in detail, we model a spectrum of internal stresses corresponding to the relaxation spectrum; the plotted internal stresses correspond to the averages across this spectrum; the first relating to the net strain accommodated by this mechanism (Equations S30 and S36 of the Supporting Information) and the second corresponding to the relevant average internal stress for calculating transient strain rate (Equations S37 and S38 of the Supporting Information). In Figure 14c, we examine the strain rate arising from each of these mechanisms as a function of time. In Figure 14d, we examine the evolution of the effective stresses for each mechanism, defined as the difference between the equivalent applied stress and the backstress/internal stress for each mechanism.

The initial increment of transient deformation is predicted to be dominated by EAGBS, which has an equivalent strain rate of $2 \times 10^{-7} \text{ s}^{-1}$. This value is controlled by the viscosity of grain-boundary sliding. The predicted initial strain rates following the seismic step for diffusion creep and dislocation-mediated deformation are substantially slower at $\sim 3 \times 10^{-11} \text{ s}^{-1}$ and $\sim 2 \times 10^{-12} \text{ s}^{-1}$, respectively. The initial strain rate due to transient diffusion creep is primarily controlled by the choice of lower limiting timescale for the diffusion-creep relaxation spectrum (Section S3.6 of the Supporting Information). By choosing that this timescale is equal to that of elastically-accommodated grain-boundary sliding, we have maximised the transient strain rate from diffusion creep in this time period. The initial strain rate arising from dislocation-mediated deformation can be rationalised by considering the dislocation backstress, which controls the rate of dislocation-mediated deformation through the flow law for dislocation glide. Prior to the seismic event, the dislocation backstress was approximately equal to the equivalent applied stress, leading to a small effective stress of approximately $\sim 10^{-1} \text{ MPa}$ acting on the dislocations. The seismic event increases the equivalent applied stress to approximately $\sim 10^2 \text{ MPa}$, increasing the effective stress to a similar value. Thus, the seismic event increases the effective stress acting on dislocations by approximately three orders of magnitude, leading to a three-orders-of-magnitude increase in the strain rate due to dislocation-mediated deformation relative to the steady-state value prior to the modelled seismic event.

Over the time interval 10^2 – 10^3 s , the equivalent bulk stress decreases from its initial value towards 105 MPa. This time interval is congruent with the characteristic timescale of EAGBS predicted by the model, $6 \times 10^2 \text{ s}$. Simultaneously, the internal stress associated with EAGBS rises significantly, reducing the effective stress driving this mechanism and therefore its strain rate. This increase in internal stress also indicates that EAGBS is accommodating strain. These observations are consistent with EAGBS acting as the dominant transient creep mechanism over this time interval and thereby controlling this initial reduction in bulk equivalent stress. As the characteristic timescale of EAGBS corresponds to the lower limiting timescale of the diffusion-creep relaxation spectrum in our model, the strain rate associated with diffusion

creep begins to decrease at this time. This reduction in strain rate is accompanied by an increase in the rate-averaged internal stress of diffusion creep. However, the rate-averaged internal stress approaches the applied stress more slowly than the internal stress associated with EAGBS. Moreover, the strain-averaged internal stress associated with diffusion creep has not changed significantly by this time, reflecting the fact that transient diffusion creep has not accommodated any significant strain.

By 6×10^3 s, the internal stress associated with EAGBS has approached the equivalent stress, and so the strain rate associated with this mechanism has decreased drastically, becoming negative. This negative strain rate is a consequence of the fact that the EAGBS effective stress also becomes negative, as the internal stress must unload while other mechanisms facilitate continuing stress relaxation. At this time, the equivalent stress has relaxed from the imposed value of 127 MPa to ~ 105 MPa, primarily through EAGBS. The largest positive contribution to the creep strain rate is now provided by diffusion creep, which is predicted to have a strain rate of $3 \times 10^{-12} \text{ s}^{-1}$ compared to a strain rate of $1 \times 10^{-12} \text{ s}^{-1}$ from dislocation-mediated deformation. However, the strain rate associated with diffusion creep continues to decrease, approximately following the “Andrade creep” law, in which strain rate decreases following the relationship $\dot{\epsilon} \propto t^{-2/3}$ reflecting the corresponding decrease in the effective stress defined using the rate-averaged internal stress of diffusion creep. This Andrade creep arises from the internal stresses associated with the faster relaxing modes saturating at the applied stress. However, the strain-averaged internal stress of diffusion creep, which is most sensitive to the slower relaxation modes, remains close to its initial value, indicating that diffusion creep is playing a very limited role in the transient. In contrast, both the strain rate and backstress associated with dislocation-mediated deformation remains relatively constant at this time. In addition, the stress of the system also remains relatively constant, indicating that the strain rates associated with diffusion creep and dislocation-mediated deformation are too slow to appreciably relax the stress over these timescales.

By 10^7 s (about 4 months), the backstress associated with dislocation-mediated deformation begins to rise, the equivalent bulk stress begins to relax, and dislocation effective stress—the difference between these quantities—falls over the interval 10^7 – 10^8 s. These observations indicate that transient dislocation-mediated deformation is playing the dominant role in controlling the stress relaxation over this time interval. The increase in dislocation backstress is primarily controlled by the process of dislocation storage. In the model of Breithaupt et al. (2023), dislocation storage is sensitive to a single parameter, the plastic modulus, which characterises the rate of increase in dislocation backstress with progressive strain accommodated by dislocation-mediated deformation. During this main hardening phase, the transient rheological behaviour of dislocation-mediated deformation is primarily controlled by two factors: the increase in dislocation backstress controlled by the plastic modulus and the properties of the flow law for dislocation glide. Over this time interval, the strain rate arising from diffusion creep briefly becomes negative. This reversal reflects the relatively rapid reduction in the equivalent stress driving deformation, which causes the effective stress associated with the faster relaxing modes to become negative. Note that strain-averaged internal stress of diffusion creep is comparatively unchanging over this time interval, indicating that despite this change in the overall direction of transient diffusion creep, it continues to play a very minor role in the overall transient rheological behaviour.

By $\sim 10^9$ s (about 30 years), the convergence between the equivalent stress and dislocation backstress is nearly complete, such that the dislocation effective stress has fallen to $\sim 2 \times 10^{-1}$ MPa, of the same order of magnitude as its initial value before the seismic event. As a result of this convergence, the strain rate arising from dislocation-mediated deformation falls to $9 \times 10^{-15} \text{ s}^{-1}$. This value is greater than the preseismic strain rate, due to the structure of the flow law for dislocation glide. The flow law for dislocation glide depends

on the product of the dislocation effective stress and dislocation density. At this time, the dislocation density is approximately five times its preseismic value, and although the dislocation effective stress has reduced substantially from the initial stages of the transient, it is approximately double its preseismic value. Together, these effects maintain a dislocation-mediated strain rate approximately one order of magnitude faster than its preseismic value. However, this remaining enhancement is relatively modest compared to the approximately three-orders-of-magnitude increase in dislocation-mediated strain rate immediately following seismic loading relative to its preseismic value.

At 10^9 s, the total creep strain rate is approximately $7 \times 10^{-15} \text{ s}^{-1}$, about ten times the preseismic value, and subsequently continues to decrease at a slower rate than during the previous time interval. As discussed above, this enhancement reflects the elevated dislocation density and the remaining elevation in effective stress, although the dislocation effective stress has decreased to the same order of magnitude as its preseismic value. In contrast to the previous time interval, the transient rheological behaviour at these times largely reflects the properties of dislocation recovery, which acts to reduce the dislocation density back towards its preseismic event value. These properties are less well constrained than those of dislocation glide, which acted as a primary control on the rheological behaviour of the previous time interval, and thus there is greater uncertainty in our predictions of the stress relaxation for these long times. Over this interval, the fast relaxing modes of transient diffusion creep have adjusted to the new equivalent applied stress, and thus transient diffusion creep reverts to its prior Andrade-creep behaviour. At later times, continued stress relaxation produces a second interval of negative diffusion-creep strain rate, approximately between 2×10^{10} s and 10^{12} s, and its magnitude remains small relative to the dislocation-mediated strain rate. By 10^{12} s (approximately 32000 years), the enhanced dislocation density has largely recovered; the overall strain rate and applied stress have returned close to their preseismic values.

The overall transient rheological behaviour of our model can be summarised as follows. Initially, EAGBS dominates, but its potential for transient strain is limited, and so it also shuts off rapidly. Despite tying our model of transient diffusion creep to EAGBS, and thereby maximising its initial transient strain rate within the adopted parameterisation, transient diffusion creep makes an extremely limited contribution. Instead, at times beyond ~ 1 year, transient dislocation-mediated deformation controls further transient creep. The initial transient strain rate due to this deformation mechanism reflects the increase by three orders of magnitude in the effective stress due to seismic loading. The subsequent transient rheological behaviour of this mechanism comprises two parts. The first part is responsible for approximately half of the further stress relaxation following the initial EAGBS response, and reflects the properties of dislocation glide, which links the enhanced effective stress to transient strain rate, and dislocation storage, which controls the rate of increase in dislocation backstress and associated decrease in dislocation effective stress. The greatest variation in transient viscosity occurs in this part. This phase is followed by a second more gradual part, in which the more modest enhancement in strain rate relative to that before the seismic event is controlled by the elevated dislocation density and remaining elevation in effective stress, and the transient rheological behaviour is controlled by the properties of dislocation recovery. The predicted evolution resembles aspects of the relaxation observed in some olivine deformation experiments, which exhibit either both phases of relaxation or only the last phase depending on the crystal orientation(s) and deformation conditions (Cooper et al., 2016; Hansen et al., 2021; Hein et al., 2025; Ohuchi et al., 2024).

For coarse-grained olivine ($d=1100 \text{ }\mu\text{m}$), representative of the wall rock, our exploratory transient-creep model predicts substantial evolution in viscosity following a seismic stress increase (Section 5.4). Together, model predictions and microstructural analyses (Sections 5.1–5.3) suggest that the deforming lithospheric

mantle, which is usually at a steady state, is characterised by low effective stress and therefore is sensitive to a sudden change of stress introduced by seismic events. The dislocation-mediated contribution exhibits an initial reduction in apparent viscosity of approximately 3 orders of magnitude during post-seismic deformation and the subsequent evolution is controlled by long-range elastic interactions among dislocations. The majority of stress relaxation is controlled by the properties of dislocation glide and dislocation storage. EAGBS acts to relax some of the seismic loading at short time intervals, but is limited in the degree of stress that it can relax. Our model suggests that diffusion creep makes a very limited contribution to the transient rheological behaviour, at least for the coarse-grained olivine at the conditions investigated here ($d=1100\text{ }\mu\text{m}$; $T=800^\circ\text{C}$).

6. Conclusions

1. In the protomylonitic peridotite, olivine grains exhibit subgrains and a strong CPO, indicating that deformation was largely accommodated by dislocation-mediated processes. In the mylonitic to ultramylonitic peridotite, the olivine CPO is weak to absent, consistent with a substantial contribution from diffusion creep involving grain-boundary sliding. However, weak CPO does not imply an absence of dislocation activity. GNDs, subgrain structures, and dislocation-generated stress heterogeneity revealed by HR-EBSD demonstrate that dislocation-mediated deformation also contributed to strain accommodation in these finer-grained domains, although its relative contribution may decrease with decreasing grain size.
2. HR-EBSD maps of olivine reveal that pronounced intragranular stress heterogeneity is spatially correlated with regions of elevated GND density and the form of the probability distributions of the stress heterogeneity is consistent with that expected of a population of dislocations. The typical length scale of stress heterogeneity is a few micrometres, which is greater than the average dislocation spacing and therefore constitutes long-range internal stress. These observations provide direct evidence that long-range elastic interactions among dislocations occurred in olivine deformed in the continental lithospheric mantle. The similarity of these characteristics to those observed in laboratory experiments (Hansen et al., 2021; Wallis et al., 2017, 2020, 2021; Wiesman et al., 2024) supports the relevance of dislocation backstress and associated kinematic hardening in the continental lithospheric mantle.
3. We model the steady-state rheological behaviour as well as the evolution of the transient rheological behaviour following a seismic event. Prior to the seismic event at steady-state, the wall rock and remnant grains in the protomylonite lie within the dislocation-mediated deformation regime, the recrystallized grains in the protomylonite plot just inside the dislocation-mediated deformation field, the mylonite plots slightly within the diffusion-creep regime, and the ultramylonite lies well inside the diffusion-creep regime. This distribution indicates that the significance of dislocation-mediated deformation decreases with decreasing grain size. In all cases, the dominant dislocation recovery mechanism is predicted to be recovery facilitated by grain-boundary diffusion.
4. Our exploratory application of a transient creep model, incorporating EAGBS, transient diffusion creep, using an adaptation of the extended Burgers model from Rudge (2026), and transient dislocation-mediated deformation, using the full model of Breithaupt et al. (2023), together with parameters guided by the Finero peridotite predicts a significant evolution in mantle viscosity following a stress increase. Our results demonstrate that the strength of the lithospheric mantle after major earthquakes is highly transient and sensitive to stress changes. This transient evolution is

complex, reflecting an interchange between different deformation processes. Immediately following a sudden stress increase, the effective viscosity of the olivine-rich mantle drops, reflecting a transient weakening. This initial transient rheological behaviour results primarily from EAGBS, which initially produces an equivalent strain rate of $\sim 2 \times 10^{-7} \text{ s}^{-1}$. The degree of transient weakening is controlled by the grain-boundary viscosity. At longer timescales on the order of years, dislocation-mediated deformation dominates the transient rheological behaviour. The degree of transient weakening reflects the change in dislocation effective stress, which is drastically increased by seismic loading relative to a small initial value before the seismic event. This increase in effective stress raises the dislocation-mediated strain rate by approximately three orders of magnitude relative to its original value, to $\sim 2 \times 10^{-12} \text{ s}^{-1}$. The transient evolution proceeds through two phases. The first phase is controlled by dislocation glide and dislocation storage. In this phase, the dislocation density increases and internal backstress builds up, gradually reducing the effective stress and causing the mantle to progressively restrengthen. After this phase, the mantle remains weaker than its preseismic state, with the dislocation-mediated strain rate reduced to approximately $9 \times 10^{-15} \text{ s}^{-1}$ at 10^9 s , approximately one order of magnitude above its preseismic value. The second stage consists of a more muted variation in viscosity controlled by the properties of dislocation recovery. Conventional steady-state flow laws alone are insufficient to describe the behaviour. Instead, incorporating transient-creep mechanisms and their associated backstress or internal stresses is essential for accurately modelling postseismic mantle flow and viscosity evolution.

Data Availability Statement

All data used in this study are openly available from Figshare (Fan et al., 2026).

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Declaration of competing interest

The authors declare no competing interests.

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