

# Controls on the rheological properties of peridotite at a palaeosubduction interface: a transect across the base of the Oman-UAE ophiolite

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## **Abstract**

Studies of experimentally deformed rocks and small-scale natural shear zones have demonstrated that volumetrically minor phases can control strain localisation by limiting grain growth and promoting grain-size sensitive deformation mechanisms. These small-scale studies are often used to infer a critical role for minor phases in the development of plate boundaries. However, the role of minor phases in strain localisation at an actual plate boundary remains to be tested by direct observation. In order to test the hypothesis that minor phases control strain localisation at plate boundaries, we conducted microstructural analyses of peridotite samples collected along a ~1 km transect across the base of the Oman-United Arab Emirates (UAE) ophiolite. The base of the ophiolite is marked by the Semail thrust, which represents the now exhumed contact between subducted oceanic crust and the overlying mantle wedge. As such, the base of the ophiolite provides the opportunity to directly examine a former plate boundary.

Our results demonstrate that the mean olivine grain size is inversely proportional to the abundance of minor phases (primarily orthopyroxene, as well as clinopyroxene, hornblende, and spinel), consistent with suppression of grain growth by grain-boundary pinning. Our results also reveal that mean olivine grain size is proportional to CPO strength (both of which generally decrease towards the metamorphic sole), suggesting that the fraction of strain produced by different deformation mechanisms varied spatially. Experimentally-derived flow laws indicate that under the

inferred deformation conditions, the viscosity of olivine was grain-size sensitive. As such, grain size, and thereby the abundance of minor phases, influenced viscosity during subduction-related deformation along the base of the mantle wedge.

We calculate an order of magnitude decrease in the viscosity of olivine towards the base of the ophiolite, which suggests strain was localized near the subduction interface. Our data indicate that this rheological weakening was primarily the result of more abundant minor phases near the base of the ophiolite. Our interpretations are consistent with those of previous studies on experimentally deformed rocks and smaller-scale natural shear zones that indicate minor phases can exert the primary control on strain localisation. However, our study demonstrates for the first time that minor phases can control strain localisation at the scales relevant to a major plate boundary.

## **1 Introduction**

Earth's lithosphere consists of highly viscous tectonic plates with deformation localized within less viscous material at plate boundaries (e.g., Bercovici and Karato, 2003). However, the processes that control the development of *plate boundary-scale* shear zones remain poorly constrained. Generating strain localisation in plate-tectonic models has proven challenging, primarily due to a lack of clarity on the best rheological parameterization (e.g. Tackley, 2000). An often proposed mechanism for localized weakening is localized grain-size reduction, as exemplified by the occurrence of mylonites in ductile shear zones (e.g., Rutter and Brodie, 1988). Studies of small-scale natural shear zones (Hansen and Warren, 2015; Herwegh et al., 2011; Toy et al., 2010; Warren and Hirth, 2006) and experimentally deformed rocks (Farla et al., 2013; Hiraga et al., 2010) have demonstrated that volumetrically minor phases in polyphase rocks can control strain localisation by limiting grain-growth, enhancing and maintaining grain-size reduction, and thereby promoting grain-size sensitive creep. Theoretical models of grain-size evolution in polyphase rocks have suggested that, via this positive self-weakening feedback, minor phases play a fundamental role in the

development and long-term maintenance of weak plate boundaries (e.g., Bercovici and Skemer, 2017; Mulyukova and Bercovici, 2017).

During shearing, increased intracrystalline strain drives dynamic recrystallization and grain-size reduction (e.g., Poirier and Guillopé, 1979). Sufficient grain-size reduction can promote grain-size sensitive deformation mechanisms and associated rheological weakening (e.g., Drury, 2005). However, grain-size reduction is opposed by surface energy-driven grain growth, which increases with decreasing grain size (de Bresser et al., 1998, 2001). The steady state grain size depends on the relative contributions of grain-size reduction and grain growth (Austin and Evans, 2009; Derby and Ashby, 1987). In monophase rocks, the resulting equilibrium grain size tends to fall near the transition between the dislocation creep and diffusion creep fields, thus limiting the magnitude of rheological weakening possible through increased activity of grain-size sensitive deformation mechanisms (de Bresser et al., 1998, 2001). In polyphase rocks, however, an even finer grain size can be maintained by a well-mixed (i.e., dispersed) minor phase that pins grain boundaries, limiting grain-boundary migration and grain growth (e.g., Toy et al., 2010; Cross and Skemer, 2017). As such, by promoting and maintaining a finer grain size than would be expected in a monophase rock at a given stress, the presence of minor phases appears critical in promoting grain-size sensitive deformation and thus the dynamic rheological weakening necessary for strain localisation. This concept has been successfully implemented in analyses of plate boundary formation and stability with encouraging results (Bercovici and Ricard, 2012, 2014). However, while observations of the role of minor phases in strain localisation have been made from the thin-section to the outcrop scale, it remains to be tested with direct observation whether this process is applicable to strain localisation at an actual plate boundary.

Here we test the hypothesis that minor phases control strain localisation at plate boundaries by examining mantle rocks at a palaeosubduction interface. We analysed peridotite samples collected across a basal section of the Oman-United Arab Emirates (UAE) ophiolite. The ophiolite formed above a subduction zone during the late Cretaceous (MacLeod et al., 2013; Pearce et al., 1981; Rioux

et al., 2016). The base of the ophiolite is bound by the Semail thrust, which represents the fossilised contact between subducted Tethyan crust and overlying mantle wedge (e.g., Searle and Malpas, 1980; Searle and Malpas 1982; Cowan et al. 2014). Whereas previous studies have had to extrapolate data from experiments or natural shear zones of a small scale and/or uncertain tectonic context, the Semail thrust provides the opportunity to directly examine a former plate boundary. We used electron backscatter diffraction (EBSD) to quantify trends and relationships in grain size, modal mineral abundance, and crystallographic preferred orientation (CPO). Our results support microstructural and rheological models in which minor phases hinder olivine grain growth, which promotes diffusion creep, and thereby strongly influences strain localisation at plate boundaries.

## **2 Geologic Setting**

The Oman-UAE ophiolite (**Figure 1**) is a thrust-bound slice of oceanic lithosphere formed at *ca.* 96.4–95.5 Ma in the hanging wall of a newly initiated, northeast dipping, intraoceanic subduction zone (Pearce et al., 1981; Rioux et al., 2012, 2013, 2016; Macleod 2013). The ophiolite consists of upper-mantle peridotite overlain by a crustal section of gabbros, sheeted dykes, basaltic flows and pillows, and pelagic sediments (e.g., Glennie et al., 1973). Following subduction initiation, tholeiitic basalts, gabbros, and sediments were subducted beneath the newly forming ophiolite to form the metamorphic sole (Searle and Malpas 1980, 1982; Gnos, 1998; Cowan et al., 2014). Burial, accretion to the base of the ophiolite, and the onset of exhumation of the sole occurred within about one million years of subduction initiation (Hacker et al., 1994; Rioux et al., 2016). Following exhumation of the sole, deformation stepped to shallower levels in the foreland as convergence was accommodated by in-sequence stacking of rise, slope, and shelf sedimentary packages (e.g.; Searle, 1988). By *ca.* 79 Ma, the continental margin reached the subduction zone and was subducted to eclogite facies in the southeastern part of the ophiolite in Oman (Warren et al., 2005). The obduction process was complete by ~70 Ma and is marked by the return of shallow-marine passive margin sedimentation (Glennie et al. 1973).

For the most part, the mantle section of the ophiolite does not exhibit any obvious structure (Figure 2a). However, the basal few hundred metres of the ophiolite exhibits a mylonitic fabric, with grain size that is smaller and foliation that is more pronounced towards the base of the ophiolite (Figures 2b–d). The base of the ophiolite is marked by the Semail thrust, which placed the ophiolite over the metamorphic sole, a thrust slice of highly sheared amphibolite, quartz schist, and metachert that was scraped off the downgoing slab and accreted to the base of the mantle wedge (Figure 1).

Several structural and petrological characteristics suggest that the mylonitic fabric in the peridotites adjacent to the Semail thrust is related to the early high-temperature history of subduction. (1) Deformation was ductile and distributed over hundreds of meters. (2) In addition to olivine and orthopyroxene, the mylonitic peridotites also contain clinopyroxene and hornblende, which have previously been interpreted to be related to fluids released from the downgoing slab (Prigent et al. 2014, 2015). Microstructural evidence (discussed in more detail in *Section 4.1*) indicates that these phases were present during deformation at temperatures in excess of 650°C. (3) Although the mineral assemblage of the peridotites is not suitable for thermobarometry, thermobarometric estimates for garnet–clinopyroxene amphibolites from the metamorphic sole immediately beneath the Semail Thrust in Oman provided peak metamorphic conditions of 770–900°C and 11–13 kb (Cowan et al., 2014). U-Pb zircon ages of *ca.* 96 and 94 Ma (Rioux et al., 2016) that date peak metamorphism of the sole overlap with crystallization of the crustal section of the ophiolite at *ca.* 96.4–95.5 Ma (Rioux et al., 2013, 2012). Thus, it seems reasonable to assume that peridotites in the adjacent hanging-wall were deformed at similarly high temperatures.

While many studies have investigated the pressure, temperature, and deformation history of the metamorphic sole (e.g., Searle and Malpas, 1980, 1982; Hacker, 1990, 1994; Hacker and Mosenfelder, 1996; Gnos, 1998; Cowan et al., 2014; Rioux et al., 2016) and shear zones within the mantle section (e.g. Dijkstra et al., 2002; Michibayashi and Mainprice, 2004; Michibayashi et al., 2006; Linckens et al. 2011), relatively few have focused on the mylonitized peridotites along the base of the ophiolite. In a review of ophiolitic mantle sections, Nicolas et al. (1980) differentiate between

an upper section with a coarse-grained porphyroclastic to granular texture, and a lower section with a fine-grained porphyroclastic to mylonitic texture. They interpret the upper peridotites as preserving high-temperature asthenospheric flow beneath an oceanic spreading centre, that is overprinted by the mylonitic and porphyroclastic textures related to lower-temperature thrusting during emplacement of the ophiolite. Boudier and Coleman (1981) conducted a microstructural analysis of peridotites from a ~12 km transect across the mantle section of the ophiolite at Wadi Tayin, Oman. Olivine CPOs from this transect are characterized by [100] dominantly sub-parallel to the lineation and are interpreted to indicate slip on (010)[100] and {0kl}[100]. Boudier and Coleman (1981) additionally reported that the strength of the CPOs did not display any obvious trends with structural position. Subsequently, Boudier et al. (1988) mapped the distribution of fabrics related to either high-temperature asthenospheric flow or lower-temperature emplacement at several locations in Oman and the UAE. Unlike the previous study at Wadi Tayin, olivine CPOs weaken towards the sole and disappear in the basal mylonites (Boudier et al., 1988). In addition to harzburgite and dunite, which comprise the majority of the mantle sequence, several occurrences of lherzolite have also been described. Geochemical trends from the lherzolites have previously been interpreted to indicate some combination of melt extraction, interaction with a fluid or melt, and primary mantle heterogeneity (Takazawa et al., 2003; Khedr et al. 2013; Yoshikawa et al., 2015).

### **3 Methods**

#### *3.1 Sample collection and preparation*

The study area is located in the northern UAE where the ophiolite is folded and erosion has exposed the Semail thrust and underlying metamorphic sole in the Masafi window (Figure 1). Eight oriented peridotite samples were collected along a ~900 m transect oriented perpendicular to the strike of the foliation of the peridotite and metamorphic sole (Figure 1). The structurally lowest (easternmost) sample was located within a metre of the metamorphic sole. The transect spans a structural thickness of ~490 m in the direction normal to the mean orientation of foliation of the

peridotite and metamorphic sole, which corresponds to a horizontal distance of nearly 900 m (Figure 1b). As the strength and orientation of the foliation in the peridotites was variable and no lineation was observed, samples were cut with reference to the average orientation of foliation in the metamorphic sole ( $187^{\circ}/39^{\circ}$ ), which is similar to the orientation of banding in the peridotite. Although a lineation was not observable in hand samples of peridotite or in amphibolites in the adjacent sole, a well-developed north-south trending mineral-stretching lineation that is parallel to fold-axes was observed in metasedimentary units further down the section in the metamorphic sole. Thin sections and billets were oriented perpendicular to orientation of the average foliation and parallel to the average dip direction. All analyses were on polished billets, aside from LH14 and LH19, which were performed on thin sections. All specimens were polished with progressively finer diamond slurry from 30  $\mu\text{m}$  to 0.25  $\mu\text{m}$  and finished with either 0.05  $\mu\text{m}$  diamond or 0.03  $\mu\text{m}$  colloidal silica.

### 3.2 *Microstructural analysis*

Mineral abundances, CPO, and grain sizes were quantified with electron backscatter diffraction (EBSD) data. These data were collected using an FEI Quanta 650 scanning electron microscope (SEM) with a field emission gun and an Oxford Instruments AZtec acquisition system and NordlysNano EBSD camera. Uncoated samples were mapped in low-vacuum mode (50 Pa) at 30 kV accelerating voltage with a step size of 5  $\mu\text{m}$  (15  $\mu\text{m}$  for LH19). Data were processed using Oxford Instruments Channel5 software and the MTEX toolbox (e.g., Hielscher and Schaefer, 2008) for MATLAB®. Channel5 was used to remove isolated pixels ('wildspikes'), and non-indexed points surrounded by  $\geq 5$  neighbours within the same grain were filled with the average of their neighbour orientations. After processing, ~10–80% of the map consisted of non-indexed pixels (Table 1), with increasing amount of non-indexing reflecting the degree of serpentinization. Estimates of mean grain size and subgrain size were calculated in Channel5 using the line-intercept method (see Appendix A in Hansen et al. (2011)). The method was applied parallel (X-direction) and perpendicular (Z-direction) to the foliation. Grain and subgrain boundaries were defined as having misorientation angles  $> 10^{\circ}$  and  $1\text{--}10^{\circ}$ , respectively. A geometric correction factor of 1.5 was applied to convert

mean intercept-length to estimates of mean grain size and mean subgrain size for use in flow laws (Underwood, 1970, pp. 80–93). For a material with significant regions of unindexed pixels, as is the case for some samples in this study, the line intercept method is better suited to determining (sub)grain sizes than methods based on grain area (Humphreys, 2001). Intercept lengths are measured with the line continuing uninterrupted, ignoring unindexed pixels (e.g., in a serpentine vein), until it reaches an indexed pixel with a misorientation angle relative to the last indexed pixel greater than the set threshold. This method may, however, overestimate the grain size in samples with high enough degrees of serpentinization such that entire grains are missing.

EBSD data are used to estimate palaeostresses experienced by our samples. In monomineralic rocks, experimentally derived grain-size piezometers can be used to estimate differential stress (for olivine see: Karato et al., 1980; van der Wal et al., 1993). In polymineralic rocks, however, pinning due to minor phases may lead to overestimation of differential stress if grain size is maintained below that which would be attained in a monomineralic rock (Herwegh et al., 2011; Tasaka et al., 2014). Therefore, following the protocol outlined by Hansen and Warren (2015), we used the olivine subgrain-size piezometer of Toriumi (1979) to estimate differential stresses.

For each set of orientations, an orientation distribution function was determined using the automatic halfwidth algorithm in MTEX (Hielscher and Schaeben, 2008). Each orientation distribution function was calculated using one-point-per-grain for over 1000 grains. Unlike for grain-size measurements, grains were iteratively dilated in Channel5 using the orientation of adjacent pixels so that no non-indexed pixels remained. This minimized the likelihood of more than one point being included per grain due to grains being dissected by serpentine veins. Pole figures (i.e., graphical presentations of orientation distribution functions) are oriented in a geographic reference frame relative to the average foliation of the metamorphic sole ( $187^{\circ}/39^{\circ}$ ) immediately beneath the ophiolite. All pole figures are plotted with the same colour-scale, as multiples of uniform distribution (M.U.D.) ranging from 0–2.5, to facilitate comparison among them. The strength of the CPO was quantified



using the J-index (Bunge, 1982), which ranges from 1 for a uniform distribution of orientations to infinity for a single crystal.

### 3.3 Mineral abundances

Mineral abundances were calculated from the processed EBSD data using MTEX. Isolated porphyroclasts of orthopyroxene were excluded from modal abundance estimates as we are primarily interested in the volumetrically more significant matrix, which we assume controlled the rheological properties. Indexing ranged from ~20–90% and is primarily a reflection of the amount of serpentinization. Most of the serpentinization observed in the samples is interpreted to have resulted from surface processes, consistent with seismic velocities (Christensen and Smewing, 1981), as well as phase equilibria modelling, serpentine chemistry, and  $\delta\text{O}^{18}$  values for carbonate veins and ground water (Streit et al., 2012). Although pyroxenes can be serpentinized, olivine alters more rapidly (Wicks, 1969). As such, assuming that the other minerals were preserved and indexed, the non-indexed pixels are interpreted to have been primarily olivine prior to serpentinization. Unindexed pixels are not included as a ‘minor phase’ in our analyses, but instead considered to be composed of olivine when calculating modal proportions.

## 4 Results

### 4.1 Petrography

Modal abundances of different mineral phases determined from EBSD maps are presented in **Table 1**. Olivine is the most abundant phase in all samples, even when only considering the indexed regions. Minor phases comprise 4–33% of the mapped regions, and consist primarily of orthopyroxene, with less abundant clinopyroxene, hornblende, and spinel (**Table 1**).

Examples of the spatial distribution of different mineral phases are presented as photomicrographs in **Figures 3a–c** and EBSD maps in **Figures 4a–c** (EBSD phase maps, band contrast maps, orientation maps, and local misorientation maps for all samples are provided in the

supplementary data). Orthopyroxene occurs both as porphyroclasts (Figures 3a, 4a, 4c) and as finer-grained material mixed within the matrix (Figures 4b–c). Clinopyroxene occurs both as a fine-grained and well mixed phase throughout the matrix (Figure 4b), as well as exsolution lamellae and inclusions within orthopyroxene (Figure 4a). Hornblende is colourless in thin section and occurs both as a well-mixed, fine-grained phase in the matrix (Figures 3b–c), as well as replacement rims around and within orthopyroxene (Figures 4a and 4c). Several lines of evidence indicate that hornblende was present during high-temperature deformation, rather than having formed after deformation and/or at lower temperatures. First, hornblende is cut by serpentine veins and hence predates the low-temperature serpentinization (Figure 3b). Lizardite and chrysotile, the polymorphs of serpentine present in the ophiolite (e.g., Streit et al., 2012), are both unstable above about 400 °C (Evans, 2004), which places a minimum temperature limit for hornblende crystallization. Second, hornblende exhibits lattice curvature indicative of crystal plasticity by dislocation motion (Figure 4d), consistent with deformation temperatures in excess of 650–700 °C (Berger and Stünitz, 1996). Third, asymmetrical overgrowths of hornblende on orthopyroxene (Figure 4c) indicate that deformation occurred in the stability field of hornblende, which does not overlap with that of serpentine.

#### 4.2 Grain size and subgrain size

Figures 5 presents distributions of intercept lengths used to estimate mean olivine grain size and mean olivine subgrain size. Estimates of the mean olivine grain sizes range from 98–350  $\mu\text{m}$  (Figures 5, Table 1). Three samples (LH19, LH25, LH26) were not sufficiently indexed to allow for an accurate estimate of the mean olivine subgrain size. Estimates of the mean olivine subgrain size of the remaining five samples range from 43–58  $\mu\text{m}$  (Figures 6, Table 1). The subgrain-size piezometer of Toriumi (1979) predicts corresponding differential stresses of 25–34 MPa. Estimates of the mean grain size of minor phases (excluding large and isolated porphyroclasts of orthopyroxene) range from 61–223  $\mu\text{m}$  (Table 1).

#### 4.3 *Crystallographic preferred orientations*

Olivine CPOs vary in strength but have similar shapes (Figures 6). J-index values range from 1.1 for the weakest fabrics, to 2.6 for the strongest. Aside from LH28, all samples exhibit olivine CPOs characterized by [100] point maxima that are approximately parallel to the average foliation of the metamorphic sole, and [010] and [001] orientation distributions that range from diffuse girdles to point maxima approximately perpendicular to the average foliation of the metamorphic sole. The orientations of [100] point maxima are remarkably consistent across the transect, indicating a broadly similar deformation geometry throughout the sampled region. Furthermore, [100] point maxima are oriented roughly perpendicular to the dip direction and have a similar orientation to the mineral stretching lineation observed in the metamorphic sole (Figure 6).

#### 4.4 *Microstructure as a function of position*

Microstructural characteristics described above are presented as a function of position along the transect in Figures 7a–d. Throughout this figure, sample LH28 is singled out with a red square because it exhibits an anomalously high proportion of minor phases that are primarily made up of hornblende. Mean olivine grain size generally decreases towards the sole (Figure 7a). However, the mean olivine grain size in sample LH28 (red squares in Figure 7) is only 162  $\mu\text{m}$ , which is fine relative to adjacent samples. In contrast, mean olivine subgrain size, and hence stress estimates from subgrain-size piezometry (e.g., Toriumi, 1979), remain reasonably constant along the length of the transect (Figure 7b).

Minor phases exhibit a linear increase in abundance towards the sole from about 260 m (Figure 7c). Sample LH27, located furthest from the sole has a similar minor-phase content as the samples located at 266 m from the sole. However, sample LH28, collected 413 m from the sole, contains the highest minor-phase content of all samples, due primarily to the high hornblende content. If sample LH28 is ignored due to the anomalously high hornblende content, the minor phase fraction remains reasonably constant beyond 260 m (Figure 7c).

The strength of the olivine CPO gradually decreases towards the sole, with J-index values of 2.6 furthest from the sole, and 1.2 immediately adjacent to the sole (Figures 6 and 7d). Sample LH28 (red squares in figure 7d), however, has a weak CPO relative to adjacent samples and a J-index value of just 1.1, the lowest of any of the samples. Despite similar mineralogy, low degree of indexing (i.e., high degree of serpentinization), and location along the transect, LH25 has a considerably weaker CPO than LH26, which was collected from the same location.

#### 4.5 Relationships among grain size, minor-phase abundance, and texture strength.

Although general trends can be observed with distance from the sole, there is significant covariance among microstructural characteristics. Figure 7e demonstrates that mean olivine grain size and minor-phase abundance are inversely related. If sample LH28 (red squares) is ignored, the decrease in mean olivine grain size with increasing minor-phase content is roughly linear. In contrast, no relationship is evident between mean olivine subgrain size and minor-phase abundance (Figure 7f).

Mean olivine grain size and texture strength exhibit a rough positive relationship (Figure 7g). Samples with a mean olivine grain size less than 200  $\mu\text{m}$  display reasonably constant and relatively low J-index values ranging from 1.1–1.4, whereas the three coarser samples exhibit a larger range but generally stronger texture strength, with J-index values between 1.4 and 2.6 (Figure 7g). Sample LH28 has a somewhat weak texture strength relative to other samples with similar mean olivine grain sizes, but nonetheless is not an obvious outlier in Figure 7g.

Minor-phase abundance and texture strength exhibit an inverse relationship (Figure 7h). For minor-phase abundance below ~10%, there is a large range in the texture strength, with J-index values ranging from 1.4–2.6. However, for samples with a minor-phase abundance greater than ~10%, texture strength decreases linearly with minor-phase abundance with J-index values ranging from 1.4–1.1. In contrast to being an obvious outlier in Figures 7a, 7c, 7d and 7e, sample LH28 lies along (albeit at the end) of the trend between texture strength and minor-phase abundance (Figure 7g).

## 5 Discussion

### 5.1 Controls on strain localisation along the base of the Oman-UAE ophiolite

Studies of experimentally deformed rocks and outcrop-scale natural shear zones indicate that minor phases can control strain localisation (Farla et al., 2013; Hansen and Warren, 2015; Herwegh et al., 2011; Hiraga et al., 2010; Toy et al., 2010; Warren and Hirth, 2006, Tasaka et al., 2014, Tasaka et al., 2017). These small-scale studies are often used to infer the importance of minor phases in the formation of lithospheric-scale shear zones, and theoretical models have suggested that minor phases are crucial in the formation of tectonic plate boundaries (Bercovici and Ricard, 2014; Bercovici and Skemer, 2017). The Oman-UAE ophiolite provides a unique opportunity to directly examine strain localisation at the larger and more relevant scale of a former plate boundary. In this section, we discuss the trends and relationships among minor-phase abundance, olivine grain size, and CPO strength from the mantle section at the base of the Oman-UAE ophiolite. As olivine is the most abundant and generally weakest phase in the upper mantle, it is often presumed to control the macroscopic rheology (Karato and Wu, 1993; Zimmerman and Kohlstedt, 2004). Furthermore, the recent experiments of Tasaka et al. (2013) demonstrate that, if olivine grain size is kept constant, even an increase in pyroxene fraction of 30% (roughly the maximum observed here) only modifies the macroscopic viscosity in diffusion creep by a factor of  $\sim 2$  (see Figure 10 in Tasaka et al., 2013). As presented below, this difference is insignificant compared to the changes in viscosity associated with changing the olivine grain size. The following discussion therefore focuses on olivine rheology, and in particular, the effect of minor-phases in moderating the strength of olivine.

### 5.2 The influence of minor phases on olivine grain size

Our results resolve an interplay between minor-phase content and olivine grain size along the base of the Oman ophiolite. Based on our initial observation in [Figure 7e](#), we suggest the inverse relationship between olivine grain size and minor-phase content is a result of minor phases pinning olivine grain boundaries. If true, our microstructural data should follow the general Zener equation,

*olivine grain size* =  $c \times \text{minor-phase grain size} / (\text{fraction of minor phases})^m$ , where  $c$  is a constant of order 1 and  $m$  is a factor related to the spatial distribution of the dispersed phase (for reviews, see Evans et al., 2001; Herwegh et al., 2011). Based on theoretical development and computer simulation,  $m$  is predicted to take on values of 0.33 to 1. **Figure 8** presents our results using a value of  $m = 0.52$ , following the empirical determination by Tasaka et al. (2014) for olivine-pyroxene mixtures. Our data are consistent with this form of the Zener equation and in agreement with previous observations from natural peridotites (Linckens et al. 2011; Tasaka et al. 2013, 2014; Hansen and Warren 2015), supporting our hypothesis that the presence of minor phases controlled olivine grain size along the base of the mantle wedge.

While the role of minor phases in limiting grain size is fairly well understood, the mechanism by which minor phases become mixed (i.e., dispersed) is still somewhat enigmatic. Hypotheses that have been proposed to explain mixing include nucleation of one phase in regions dominated by another phase as a result of stress-induced chemical potential gradients (Tasaka et al., 2017), nucleation of phases from a fluid in cavities opened by grain-boundary sliding (Precigout and Stünitz, 2016), or mechanical production of new grains that are transported along grain boundaries of another phase as a result of surface tension and local stresses at the boundaries (Bercovici and Skemer, 2017). It is possible that multiple mechanisms were responsible for phase mixing in our samples. Clinopyroxene and hornblende, on the one hand, are interpreted to be produced from metasomatic fluids associated with the subducting slab (Prigent et al., 2014, 2015; Yoshikawa et al., 2015). Considering these phases are not generally seen as porphyroclasts and are primarily observed in well-mixed regions (**Figures 4b and c**), it is possible they were generated by a mechanism similar to that proposed by Precigout and Stünitz (2016). Orthopyroxene, on the other hand, does occur as porphyroclasts, and small olivine grains appear to penetrate into the larger orthopyroxene grains (**Figure 4a**). The location of these olivine grains does not appear to be controlled by orthopyroxene grain or subgrain boundaries, which is in better agreement with the model of Tasaka et al. (2017).

### 5.3 *Deformation mechanisms and rheology of the subduction interface*

We suggest that the grain-size reduction described above leads to weakening through a transition in deformation mechanism. We first use CPO strength to qualitatively assess variations in the relative contributions of dislocation creep and diffusion creep to the total strain rate. Deformation accommodated largely by the motion of dislocations typically leads to development of a CPO (e.g., Karato, 1988). In contrast, diffusion creep is presumed to produce a random or weak CPO (Fliervoet et al., 1999; Rutter et al., 1994). Although some questions remain regarding the effectiveness of diffusion creep in modifying an existing CPO (Wheeler, 2009), recent experiments on olivine-pyroxene mixtures by Tasaka et al. (2017) demonstrated that CPO strength decreases with increasing contributions of grain-size sensitive deformation to the strain rate. Therefore, the positive relationship between texture strength and mean olivine grain size in our samples (Figure 7g) indicates that a mechanism other than dislocation creep was promoted by fine grain sizes. Given the influence of minor phases on olivine grain size (Figures 7e and 8), this provides an explanation for the inverse correlation between minor-phase content and texture strength (Figure 7h).

An increase in the fraction of strain produced by grain-size sensitive deformation at a given stress can result in rheological weakening and provide a mechanism for strain localisation (e.g., Drury, 2005; Warren and Hirth, 2006). To constrain the activity of different deformation mechanisms, we constructed deformation mechanism maps using the experimentally derived flow laws for wet olivine deforming by diffusion creep and dislocation creep as given by Hirth and Kohlstedt (2003; Figure 9a). Hirth and Kohlstedt (2003) argued that dislocation-accommodated grain-boundary sliding (disGBS) is not important under hydrous conditions. Alternatively, Ohuchi et al. (2015) have more recently developed a flow law for disGBS in olivine under hydrous conditions based on high-pressure experiments. Their results, however, indicate that the contribution of disGBS to the total strain rate is insignificant at the deformation conditions expected along the base of the mantle wedge (low temperature and low stress). As such, we have not included disGBS in our calculations. Strain rates were calculated using a temperature of 800°C and a confining pressure of 600 MPa. This pressure and

temperature roughly correspond to peak temperature and intermediate pressure estimates from the hanging wall metamorphic sole (Cowan et al., 2014). The wet flow laws also include a term for hydrogen concentration (or water fugacity). Although the water content has not been precisely constrained in our dataset, the presence of hornblende is consistent with water-saturated conditions. At 800 °C and 600 MPa, the solubility of hydrogen in olivine is  $\sim 300 \text{ H}/10^6 \text{ Si}$  (Kohlstedt et al., 1996; Zhao et al., 2004). To assess the effect of water on strain rate, Figures 9b and c present predicted strain rates using both the wet (Hirth and Kohlstedt, 2003) and dry (Hirth and Kohlstedt, 2003; Hansen et al., 2011) flow laws. Unlike the wet flow law of Hirth and Kohlstedt (2003), the dry flow laws of Hansen et al. (2011) and Hirth and Kohlstedt (2003) include a contribution from disGBS in addition to dislocation and diffusion creep. The wet flow law for water saturated olivine predicts an increase in strain rate of approximately two orders of magnitude relative to the dry flow laws (Figure 9b). Importantly, the strain rate increases along the transect towards the sole, a feature that is apparent regardless of the water content used (Figures 9b and c). Given the presence of hornblende, which suggests that significant amounts of water were present during deformation, and that the overall trends are not drastically affected by varying the water content, we have used a water concentration of  $300 \text{ H}/10^6 \text{ Si}$  in the flow laws for all samples (Figure 9). It is worth noting, however, that the heterogeneous distribution of hornblende (Figure 3) is consistent with channelized flow of a hydrous fluid (e.g. Plummer et al. 2016). Thus, it is possible that hydration of olivine varied along the section.

Our calculated deformation mechanism map is presented in Figure 9a. The map is colored by the fraction of the total strain rate due to dislocation creep to emphasize that the transition from dislocation creep to diffusion creep is gradational and does not reflect an abrupt ‘switch’ in deformation mechanism (e.g., Skemer and Hansen, 2016). Dislocation creep continues to operate at the same strain rates for a given stress in the diffusion creep field in stress-grain-size space. However, the proportion of the total strain rate due to dislocation creep decreases with decreasing grain size as the efficiency of diffusion creep increases. Our samples are plotted in Figure 9a using measured olivine grain sizes and stress estimates from subgrain-size piezometry. All samples fall to the left of



the nominal boundary between deformation mechanisms (50% diffusion creep and 50% dislocation creep). Aside from LH27, which was located furthest from the sole thrust, all of our samples are finer grained than predicted by grain-size piezometry (white dashed line; Van der Wal et al., 1993). The predicted percent of the total strain rate due to dislocation creep ranges from 3–46% and generally decreases towards the sole (Figure 9d). Despite similar estimated stresses, this increase in diffusion creep is associated with an increase in predicted strain rates by a factor of  $\sim 7.5$  ( $4 \times 10^{-14}$  –  $3 \times 10^{-13} \text{ s}^{-1}$ ) towards the metamorphic sole (Figures 9b and c). We also calculate that apparent viscosities decrease by a factor of  $\sim 10$  ( $10^{22}$ – $10^{21} \text{ Pa s}$ ) within  $\sim 100 \text{ m}$  of the metamorphic sole (Figures 9e and f).

Our estimated strain rates are reasonably consistent with geochronological constraints for both the ophiolite and metamorphic sole, as well as pressure estimates from the sole. Geochronology (Rioux et al., 2012, 2013, 2016) and thermochronology (Hacker, 1994) of the crustal sequence and metamorphic sole indicate that high-temperature deformation along the base of the ophiolite lasted approximately 1 Myr. The subducting slab reached depths of  $\sim 40 \text{ km}$  (Gnos, 1998; Cowan et al., 2014). For a dip of  $45^\circ$  and perpendicular convergence, these constraints correspond to  $\sim 56 \text{ km}$  of movement along the subduction interface and an upper bound for the strain rate of  $\sim 10^{-12} \text{ s}^{-1}$  for a 1 km wide shear zone.

The results from our microstructural analysis are consistent with the hypothesis that minor phases pinned olivine grain boundaries and limited grain-boundary migration, which hindered olivine grain growth and promoted diffusion creep. This transition in deformation mechanism had a weakening effect on olivine (Figure 9a). As such, we suggest it was ultimately the distribution of minor phases that controlled the rheological weakening at the base of the mantle section. Similarly, Linckens et al. (2011) demonstrated a correlation between minor-phase abundance, grain size, and the rheological properties of peridotite from the vertical, strike-slip Hilti shear zone located within the middle of the mantle section in Oman. Here, we have demonstrated that minor phases also influenced strain localization during subduction-related deformation along the base of the ophiolite. Our results

are consistent with theoretical models that predict minor phases are necessary for the persistence of weak zones (i.e., plate boundaries) within the lithosphere (e.g., Berocovici and Ricard 2012, 2014; Mulyukova and Bercovici, 2017).

Although generally stronger than olivine (e.g., Hansen and Warren, 2015), the net effect of increasing pyroxene content along our transect was a viscosity reduction. This conclusion is in contrast to the results of Hansen and Warren (2015). They examined a shear-zone formed under relatively dry conditions with a significant portion of the deformation occurring by disGBS. In the disGBS regime, they suggested the intrinsic strength of pyroxene roughly offsets any grain-size related weakening due to the relatively small grain-size sensitivity of disGBS. Thus, the hydrous conditions in the mantle section indirectly promoted significant weakening because diffusion creep is enhanced relative to disGBS (at this temperature and stress).

#### *5.4 Orientation of CPOs and strain geometry at the subduction interface*

For the deformation conditions expected along the subduction interface, it is expected that olivine would have [100] axes aligned approximately parallel to the transport direction (see Figure 4 in Skemer and Hansen (2016)). However, Figure 6 demonstrates that olivine CPOs exhibit [100] axes at a high angle to the dip direction and more closely aligned with the stretching lineation in the metamorphic sole, particularly for the lower most sample. Four possible explanations can account for the difference between the orientation of the alignment of [100] axis and that of the dip direction. (1) It is possible that formation of the olivine CPOs predated deformation along the subduction interface. However, the high degree of olivine recrystallization, systematic trends in the olivine CPO towards the metamorphic sole, orientation of the foliation parallel to the subduction interface, and evidence of deformation in the presence of hornblende all indicate that deformation and development of the CPO resulted from shearing along the subduction interface. (2) Another explanation is that the CPOs developed during high-stress and wet deformation conditions (Jung and Karato, 2001; Jung et al. 2006), which has been demonstrated to produce CPOs with [100] axes aligned perpendicular to the

shear direction and the shear-plane normal (B-type fabric). However, the development of a B-type fabric requires significantly higher stresses than those estimated here by subgrain-size piezometry, even if the relatively low temperature of deformation is taken into account (Karato et al., 2008). (3) It is possible that the CPO developed during oblique convergence. However, the strain rates required to subduct the sole to 40 km in only 1 million years if subduction had a dominant strike-slip component are unrealistically high. Furthermore, thrusts, axial planes, and fold axes generally trend north-south, indicating that shortening was oriented east-west. (4) Our preferred explanation is that deformation was constrictional during (near) perpendicular convergence with a strong component of trench-parallel extension. Constrictional deformation is consistent with the well developed mineral stretching lineation in observed metasedimentary rocks in the metamorphic sole.

### *5.5 Tectonic Implications*

The role of minor phases in localising strain at the base of the mantle wedge has important implications for subduction zone processes. Theoretical models (e.g., Bercovici and Ricard 2012, 2014; Mulyukova and Bercovici, 2017) have predicted that, by limiting grain growth and promoting grain-size sensitive deformation mechanisms, minor phases drive a positive self-weakening feedback that allows for the persistence of lithospheric weak zones. In particular, few geological processes are capable of reversing phase mixing. Thus, the presence and distribution of minor phases plays a critical role in the development and long-term maintenance of plate boundaries. Orthopyroxene was the most abundant minor phase in seven of eight samples in our transect. The large-scale spatial distribution of orthopyroxene is most easily explained by pre-existing mantle heterogeneity. In contrast, hornblende, the most abundant minor phase in one sample, and clinopyroxene have previously been interpreted to form during metasomatism of the mantle wedge by slab derived fluids (Prigent et al., 2015, 2014; Yoshikawa et al., 2015). Therefore, fluid transfer can yield additional localized weakening through crystallization of minor phases that limit olivine grain growth and enhance diffusion creep. These fluids may additionally weaken the mantle through 1) the reduction of the crystal plastic strength of nominally anhydrous minerals, such as olivine, through incorporation of intracrystalline hydrogen

(e.g., Mei and Kohlstedt, 2000a, 2000b) or 2) crystallization of weaker phases such as serpentine and chlorite. Thus, weakening and localization in the mantle near the subduction interface appears controlled by both pre-existing heterogeneity and infiltration of fluids from the slab.

In contrast to weakening of the mantle wedge, concomitant dehydration of the slab would have had a strengthening effect. The magnitude of the strengthening in the slab, however, is uncertain as there is a paucity of experimental data on the rheological influence of dehydration of metabasites. Fluid infiltration and amphibolitisation of gabbro (i.e., approximately the reverse reaction to that observed in the sole) has been demonstrated to have a significant weakening effect, primarily from the hydration of plagioclase (Getsinger et al., 2013).

The contrasting rheological evolution of the mantle wedge and downgoing slab has previously been proposed to control the progressive deepening of the subduction interface and associated accretion of slices of the downgoing slab (i.e., the future metamorphic sole) to the hanging-wall peridotites (e.g. Hacker, 1990; Agard et al., 2016). Agard et al. (2016) suggested that strain localizes where there is a large contrast in rheologies. In the model of Agard et al. (2016), the downgoing slab is initially stronger than the serpentinized mantle wedge and strain localizes at the interface between the two. However, with increasing depth and temperature the mantle wedge becomes less serpentinized and stronger, whereas the slab becomes hotter and weaker. At some critical point the downgoing slab and mantle wedge have similar viscosities and the locus of highest strain migrates into the slab. Here we have demonstrated that the abundance of minor phases exerts an additional and significant rheological control along the base of the mantle wedge not accounted for in previous models. Thus, dehydration and strengthening of the slab and concomitant metasomatism and weakening of the mantle wedge would enhance the strength contrast across the subduction interface, thereby limiting coupling to greater depths.

While our results demonstrate that the heterogeneous distribution of minor phases influences plate boundary development, they also raise questions as to the origin and extent of petrological

heterogeneities in the mantle. Further insight would be gained by grid sampling (at the mm through to the km scale) across the mantle section of the ophiolite in bog-standard peridotite and characterizing the distributions of grain size, composition, water, CPO, and modal mineral abundances on each scale.

## **6 Conclusions**

Models of strain localisation at plate boundaries rely on the extrapolation of data from experimentally deformed rocks, small-scale natural shear zones, and shear zones in which the tectonic context is ambiguous. Direct observation of the strain localization at plate boundaries is difficult due to the paucity of suitable exposure. We examined samples of peridotite collected across the base of the Oman-UAE ophiolite to elucidate the rheological controls on a palaeosubduction interface and use laboratory-based constraints to identify strain localization.

Microstructural analyses reveal systematic changes in olivine grain size, minor-phase content, and olivine CPO. We demonstrate that minor phases pinned olivine grain boundaries and limited grain growth. The resultant finer grain size promoted diffusion creep and led to rheological weakening. We calculate that viscosity and strain rate respectively decrease and increase by approximately an order of magnitude towards the base of the ophiolite. Our data indicate that this rheological weakening was the result of more abundant minor phases near the base of the ophiolite. Orthopyroxene is the most abundant minor phase in all samples. The distribution of orthopyroxene is interpreted to reflect compositional heterogeneity that predates deformation. Unlike orthopyroxene, hornblende and clinopyroxene are interpreted to have crystallised during metasomatism of the mantle wedge by fluids from dehydration of the downgoing slab. Thus, in addition to hydrating and weakening olivine, metasomatic fluids released from the downgoing slab indirectly influenced strain localization by promoting the crystallization of minor phases that limited olivine grain size and promoted grain-size sensitive deformation mechanisms. Furthermore, based on our microstructural evidence and recent laboratory deformation experiments, we suggest that the presence of a hydrous fluid promoted diffusion creep relative to disGBS, enhancing the grain-size sensitivity and associated weakening.

While observed in experimentally deformed rocks and at the outcrop scale, this is the first demonstration that minor phases controlled strain localisation at a plate boundary. Moreover, our interpretations are consistent with theoretical models that suggest the distribution of minor phases strongly influence strain localisation along plate boundaries.

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## **Figure Captions**

Figure 1. (a) Simplified geological map of the northern United Arab Emirates (modified after Farrant et al. (2012)) and elevation model showing the map location on the northeast margin of the Arabian peninsula. (b) Cross-section across the uppermost sole and basal peridotites with locations of analysed samples. Along the western margin of the Masafi window the Semail thrust has an apparent normal sense because the transport direction is to the west and the Semail thrust has been affected by later (post-emplacement) folding. Foliation is indicated by dashed lines. Distance above the sole was estimated using the average foliation of the upper sole and banding in peridotites, which approximate the orientation of the thrust contact. Horizontal distance is taken from the sole-peridotite contact.

Figure 2. Field photos of the Oman-UAE ophiolite mantle section. (a) Most of the mantle section lacks obvious lineations and foliation. (b–c) The basal kilometre of the mantle peridotites displays a foliation that increases in intensity towards the base. (d) The lowermost few metres of the ophiolite exhibit a well developed mylonitic fabric.

Figure 3. Representative photomicrographs displaying: (a) an orthopyroxene porphyroclast surrounded by a fine-grained olivine matrix in sample LH18, (b) orthopyroxene, olivine, and hornblende cut by serpentine in sample LH28, and (c) olivine, hornblende, spinel, and serpentine in sample LH19. Ol (olivine), Opx (orthopyroxene), Cpx (clinopyroxene), Hbl (hornblende), Sp (spinel), Srp (serpentine).

Figure 4. EBSD maps. (a) Phase map overlain on band contrast map of sample LH18 displaying an orthopyroxene porphyroclast with clinopyroxene exsolution lamellae and hornblende replacement along fractures. Olivine appears to have penetrated and mixed with orthopyroxene along the bottom of the porphyroclast. (b) Phase map overlain on band contrast map of sample LH17 displaying olivine with well-mixed (dispersed) minor phases (orthopyroxene, clinopyroxene, and hornblende). (c) Phase map overlain on band contrast of map LH28 displaying porphyroclasts of orthopyroxene overgrown by hornblende and a finer-grained matrix of olivine, orthopyroxene, and hornblende. (d) Map of local misorientation in hornblende averaged within 3x3 pixel kernels. Abbreviations: Ol (olivine), Opx (orthopyroxene), Cpx (clinopyroxene), Hbl (hornblende), Sp (spinel). Gray pixels are mostly serpentine that was not indexed during EBSD analysis.

Figure 5. Histograms of olivine grain size (left column) and subgrain size (right column) from intercepts measured parallel (x-direction) and perpendicular (z-direction) to foliation. Samples are ordered according to distance from the sole. The average grain sizes in Figures 7, 8, and 9a, and Table 1 are an average of the x- and z-direction. A geometric correction factor of 1.5 was applied to convert intercept length to grain size.

Figure 6. Olivine pole figures for all samples illustrating the change in CPO from with distance from the sole. Distance from sole and J-index value are included for reference. Pole figures are oriented within a geographic reference frame, as schematically shown at the bottom. All figures are contoured for multiples of uniform distributions (M.U.D.) ranging from 0 to 2.5.

Figure 7. Microstructural relationships. Subgrain sizes were not calculated for 3 samples due to insufficient indexing. In all plots, the red square corresponds to LH28.

Figure 8. Mean olivine grain size as a function of mean minor-phase grain size and minor-phase volume fraction. Following Tasaka et al. (2014), we use a value of  $m = 0.52$  for the minor phase volume fraction. Our samples are plotted along with data from synthetic olivine-pyroxene aggregates (Tasaka et al. 2013), as well as data from natural shear zones in Oman (Linckens et al., 2011, 2015; Tasaka et al., 2014) and the Josephine peridotite (Hansen and Warren, 2015). The solid line is the best fit derived by Tasaka et al. (2014).

Figure 9. (a) Deformation mechanism map for olivine using the wet flow laws for diffusion creep and dislocation creep (Hirth and Kohlstedt, 2003). Our samples are plotted using the measured grain size, and stress estimates from subgrain-size piezometry. The colour scale corresponds to the fraction of the total strain rate produced by dislocation creep. Flow laws were calculated using 800 °C, a confining pressure of 600 MPa, and 300 H/10<sup>6</sup>Si. Black lines are contours for constant strain-rate. White dashed

line is the grain-size piezometer of Van der Wal et al. (1993). (b) Predicted strain rates and (c) strain rates normalized to sample LH27 using the flow laws for wet olivine (black circles) and dry olivine (gray circles) as a function of distance from the sole. (d) Fraction of dislocation creep for wet olivine as a function of distance from the sole. (e) Predicted viscosities and (f) viscosities normalized to sample LH27 for wet olivine as a function of distance from the sole. Sample LH28 is displayed as a red square in all panels.

Table 1. Results from microstructural analysis

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