Quantifying the effect of pyroxene on deformation of peridotite in a natural shear zone

Lars N. Hansen1,2 and Jessica M. Warren2

1Department of Earth Sciences, University of Oxford, Oxford, UK, 2Department of Geological and Environmental Sciences, Stanford University, Stanford, California, USA

Abstract The influence of pyroxene on upper mantle viscosity remains unclear but may occur either through its difference in strength relative to olivine or through its effect on olivine microstructure. To determine the role of pyroxene in moderating peridotite viscosity, we analyzed microstructures in paired harzburgites and dunites from a natural shear zone exposed in the Josephine Peridotite. Olivine crystallographic textures evolve similarly in harzburgites and dunites with increasing strain, indicating the operation of similar deformation mechanisms in olivine. The mean olivine grain sizes are ~1.5 times larger in dunites than in harzburgites, whereas the mean olivine subgrain sizes are a factor of ~1.2 smaller in dunites than in harzburgites. The average stresses in olivine, estimated with a subgrain-size piezometer, are inversely correlated with pyroxene volume fraction and directly correlated with the mean olivine grain size. The calculated ratio of pyroxene viscosity to olivine viscosity for each harzburgite/dunite pair varies from 1.2 to 3.3. Notably, our data indicate that olivine viscosity is non-Newtonian with a finite grain-size sensitivity. We suggest that at the conditions of the shear zone, pyroxene is more viscous than olivine but that the increase in aggregate strength with larger pyroxene fractions is offset by correspondingly smaller olivine grain sizes. Thus, major rheological weakening associated with pyroxene content may only be possible if (1) temperatures are higher than those characterizing this shear zone, increasing the viscosity contrast between olivine and pyroxene or (2) stresses are high enough that grain-size reduction promotes a transition to a deformation mechanism with extreme grain-size sensitivity, such as diffusion creep.

1. Introduction

High-temperature viscous flow in the Earth depends on the material properties of the mineral phases that make up the rocks undergoing deformation. Therefore, large-scale geodynamic processes can be related to viscous deformation of individual minerals. Because the composition of Earth's upper mantle is dominantly olivine (>60%), much progress has been made using laboratory-derived flow laws for olivine [e.g., Hirth and Kohlstedt, 2003] in modeling the dynamics of subducting slabs [e.g., Billen and Hirth, 2007; Buffett and Becker, 2012], flow in suprasubduction mantle wedges [e.g., van Keken et al., 2008; Wada et al., 2012], shear of the oceanic asthenosphere [e.g., Podolak et al., 2004; Behn et al., 2009], the formation of mantle shear zones [e.g., Warren and Hirth, 2006; Toy et al., 2010; Farla et al., 2013; Skemer et al., 2013], and large-scale convection [e.g., Sandu et al., 2011; Rozel, 2012].

Laboratory-derived flow laws are primarily calibrated for nearly monophase aggregates of olivine, but upper mantle rocks contain a significant volume fraction of secondary phases. Notably, most peridotites are made up of 30 to 40% pyroxene [Bodinier and Godard, 2007], and in depleted peridotites (i.e., harzburgites), most of that is orthopyroxene. Field-based observations of deformed peridotites indicate that orthopyroxene may play a significant role in controlling peridotite viscosity. Orthopyroxene has been suggested to influence peridotite deformation by inhibiting olivine grain growth such that grain sizes remain small and grain-size sensitive deformation mechanisms are activated [e.g., Warren and Hirth, 2006; Toy et al., 2010; Linckens et al., 2011; Skemer et al., 2013]. A counteracting effect may arise, however, from the difference in strength between olivine and orthopyroxene, because field observations, although limited, indicate that pyroxene is stronger than olivine in natural settings [Tikoff et al., 2010; Herwegh et al., 2011; Linckens et al., 2011].

Quantifying the rheological behavior of olivine-orthopyroxene aggregates is subject to significant uncertainty, mainly due to the difficulty in conducting experiments on purely orthopyroxene compositions. Laboratory measurements of the viscosity of orthopyroxene aggregates and single crystals differ by many orders of magnitude, but the relative abundance of orthopyroxene compared to olivine suggests that it is the major contributor to upper mantle viscosity. To test this hypothesis, we conducted subgrain size piezometry experiments on a primitive pyroxene-rich fraction from the Josephine Peridotite. This is an open access article under the terms of the Creative Commons Attribution-NonCommercial-NoDerivs License, which permits use and distribution in any medium, provided the original work is properly cited, the use is non-commercial and no modifications or adaptations are made.
Electron backscatter patterns and X-ray energy spectra were collected simultaneously with Oxford Instruments. Low-pressure mode was used, and the chamber pressure was maintained at 20 Pa with water vapor. Uncoated samples were analyzed in an FEI Quanta 200 scanning electron microscope with a W 60 min of polishing with colloidal silica (0.06 μm).

Microstructures within samples from the Josephine Peridotite were analyzed by mapping with electron backscatter diffraction (EBSD). Thin sections were prepared with an ultrahigh polish for EBSD analysis by polishing with diamond suspensions on lapping cloths down to a grit size of 0.25 μm. Microstructures of the two rock types allows conclusions to be drawn about the manner in which olivine deformation is affected by pyroxene content. Importantly, we use laboratory-derived relationships between olivine microstructures and stress in combination with constitutive models for two-phase aggregates to demonstrate that pyroxene affects peridotite deformation through both a grain-size effect and its strength contrast with olivine.

2. Methods

2.1. Sample Collection

The samples analyzed in this study, originally described by Warren et al. [2008], were collected in a transect across shear zone P (SZP) in the Josephine Peridotite. This peridotite massif is the well-exposed mantle section of an ophiolite in southwestern Oregon and northern California. Based on geochemical and petrological analyses and the presence of volcaniclastic sediments, the Josephine Peridotite is interpreted to have originated from a back-arc environment [Harper, 1984]. The Fresno Bench section consists of glacially polished outcrops that host at least 10 shear zones wider than 1 m [Loney and Himmelberg, 1976; Kelemen and Dick, 1995]. As illustrated in Figure 1, Fresno Bench is dominated by layered peridotites of varying pyroxene content (dunites) are interlayered with rocks with a relatively high pyroxene content (harzburgites). Both appear to have deformed at the same conditions during formation of the small-scale shear zone. Thus, a comparison between microstructures of the two rock types allows conclusions to be drawn about the manner in which olivine deformation is affected by pyroxene content. Importantly, we use laboratory-derived relationships between olivine microstructures and stress in combination with constitutive models for two-phase aggregates to demonstrate that pyroxene affects peridotite deformation through both a grain-size effect and its strength contrast with olivine.

The shear zone from which our samples originate, SZP, is approximately 60 m wide and comprises a nearly vertical shear plane at a high angle to the primary compositional layering. During deformation, the compositional layering was deflected as illustrated in Figure 1. Assuming simple shear, the orientation of layering relative to the shear plane can be used to define the magnitude of shear strain [Warren et al., 2008]. The maximum shear strain measured in SZP is ~5.25. For this study, we analyzed pairs of samples consisting of a dunite and harzburgite sampled within 1 m of each other (Figure 1). Many paired samples were collected within several centimeters of each other, and seven were directly juxtaposed and collected as part of the same hand specimen. Throughout this study, samples are analyzed and considered in harzburgite-dunite pairs as both samples making up a pair were deformed at the same thermomechanical conditions and to the same finite strain.

2.2. Microstructural Analysis

Microstructures within samples from the Josephine Peridotite were analyzed by mapping with electron backscatter diffraction (EBSD). Thin sections were prepared with an ultrahigh polish for EBSD analysis by polishing with diamond suspensions on lapping cloths down to a grit size of 0.25 μm. A final polishing step consisted of 60 min of polishing with colloidal silica (0.06 μm).

Uncoated samples were analyzed in an FEI Quanta 200 scanning electron microscope with a W filament. Low-pressure mode was used, and the chamber pressure was maintained at 20 Pa with water vapor. Electron backscatter patterns and X-ray energy spectra were collected simultaneously with Oxford Instruments.
Nordlys-F+ and X-Max 20 mm² detectors, respectively. Data acquisition was managed with the Oxford Instruments AZtec (version 2) software package. The electron beam was rastered across the sample with a step size of 15 μm, acquiring a diffraction pattern and solving for phase and crystallographic orientation at each step. As depicted in Figure 2, mapping of each sample consisted of 16 to 40 individual maps that were subsequently stitched in postprocessing into a mosaic covering ~60 mm². Map mosaics typically contained ~300 grains, and therefore, a second mosaic was constructed for each sample using a coarser step size (50 μm) to quickly increase the number of grain orientations available for texture analysis. To prevent systematic bias due to differences in procedure, any samples that were previously mapped and reported in Warren et al. [2008] were remapped as part of this study.

Postprocessing of EBSD maps was carried out with the HKL Channel5 software package. Isolated single pixels that differed by more than 10° from their neighbors were initially removed. Pixels for which indexing was not possible were assigned the average orientation of neighboring pixels if eight nearest neighbors were present. This process was iterated until no new pixels were filled. The process was then reiterated for empty pixels with seven and then six nearest neighbors. Because the subgrain-size analysis described below relies on the measurement of small changes in orientation between neighboring pixels, an orientation averaging scheme was applied (Kuwahara filter) to reduce noise. The Kuwahara filter was applied 3 times, which has been demonstrated to reduce orientation noise by a factor of 10 and enable misorientations <0.5° to be detected in Al [Humphreys et al., 2001].

To quantify the shape of crystallographic textures, we used an eigenvalue analysis of crystallographic orientation distributions recently employed by Hansen et al. [2014]. Orientation distributions were constructed with one orientation per olivine grain, and each distribution consisted of approximately 600 grains. The analysis employed follows Woodcock [1977] and calculates the eigenvalues, $S_i$, and eigenvectors, $\lambda_i$, of the orientation tensor. The eigenvector with the largest eigenvalue, $\lambda_1$, describes the average orientation of a particular crystallographic axis. Additionally, the eigenvector with the smallest eigenvalue, $\lambda_3$, describes the normal to the great circle that best
Figure 2. Electron backscatter diffraction (EBSD) data from a representative example of a paired (top) dunite and (bottom) harzburgite. For each sample, maps were constructed based on band contrast (gray scale) and crystallographic orientation (color scale). For maps of band contrast, lighter gray indicates better diffraction patterns. For maps of orientation, darker colors indicate better alignment of [100] with the shear direction. Olivine grains are colored on a green scale, and pyroxene grains (both orthopyroxene and clinopyroxene) on a blue scale. On orientation maps, gray pixels indicate regions in which orientation solutions were not attainable, which mainly corresponds to regions of low-temperature alteration. Grain boundaries (black lines) and subgrain boundaries (red lines) are included in the orientation maps based on minimum misorientations of 10° and 1°, respectively.
fits the distribution of axes. The ellipsoid describing the distribution of orientations can be further quantified through comparison of \( \ln(S_1/S_2) \) and \( \ln(S_2/S_3) \) via the shape parameter, \( K \), given by

\[
K = \frac{\ln(S_1/S_2)}{\ln(S_2/S_3)},
\]

where \( K \) is >1 for ellipsoids that are more prolate (i.e., distributions of orientations that form a cluster on a pole figure) and <1 for those that are more oblate (i.e., distributions of orientations that form a girdle on a pole figure). If the distribution of axis orientations is distinctly girdled (\( K < 1 \)), then the smallest eigenvector, \( \lambda_3 \), describes the pole to the girdle.

To quantify the strength of olivine crystallographic textures, we used both the \( M \)-index [Skemer et al., 2005] and \( J \)-index [Bunge, 1982]. The \( M \)-index is calculated from the distribution of uncorrelated misorientation axes and has a value of 0 (random) to 1 (single crystal). The \( J \)-index is calculated from the orientation distribution function and has a value of 1 (random) to infinity (single crystal). Orientation distribution functions were calculated using the MTEX toolbox for MATLAB and a kernel half width of 10° [Bachmann et al., 2010].

Arithmetic mean grain and subgrain sizes were estimated using the linear intercept method on EBSD maps. A range of minimum misorientation angles used to define subgrain and grain boundaries were explored. In general, a misorientation of 1° was used to define the lower limit of subgrain boundaries, and a misorientation of 10° was used to define the lower limit of grain boundaries. A discussion of the choice of 1° misorientations to define subgrain boundaries is provided in Appendix A. To convert between mean intercept length on a two-dimensional section and the mean grain diameter in three dimensions, a scaling factor of 1.5 was used [Underwood, 1970, pp. 80–93]. The linear-intercept method is particularly valuable compared to stereological techniques that measure bounded areas because it does not require complete closure of a boundary. Additionally, subgrain sizes in calcite measured by the linear-intercept method were determined by Volcke et al. [2006] to be the most representative of the true microstructure out of three different measurement methods.

### 3. Results

#### 3.1. Crystallographic Textures

Crystallographic textures from all analyzed harzburgites and dunites are presented in Figure 3. Data from quantitative analysis of textures are presented in Table 1. For all samples, moderate to strong textures were observed (\( J \)-index range of 1.1 to 6.8 and \( M \)-index range of 0.08 to 0.27). Distributions of [100] axes tend to form clusters, [010] axes tend to form girdles, and [001] axes exhibit relatively little alignment. With increasing strain, olivine textures in both harzburgites and dunites follow the same general evolution. Warren et al. [2008] previously demonstrated that the average orientation of [100] in harzburgites becomes increasingly parallel to the shear direction with progressive amounts of strain, indicating that a significant portion of the deformation resulted from the motion of dislocations with [100] Burgers vectors. Here we confirm that prior result and establish that the same is true of textures in dunites.

Quantification of olivine textures in both harzburgites and dunites using eigenvalue analysis is presented in Figure 4. With increasing strain, the average olivine orientations in paired harzburgites and dunites follow the same evolution, always within ~10° of each other. The average [100] orientation and the normals to [010] and [001] girdles approach the shear direction with increasing strain, although none reach values closer than ~5° to the shear direction. These values are systematically larger than those reported in Warren et al. [2008], which were calculated using the angle between the average direction and the shear plane. This difference in values may result from the average [100] direction often lying out of the \( x-y \) plane.

Evolution of the shape parameter, \( K \), for [100] and [010] axis distributions as a function of strain is also similar for both harzburgites and dunites (Figure 4d). Most [100] distributions have values of \( K > 1 \), indicating that they are more clustered than girdled, although several do exhibit values of \( K \) that demonstrate girdled [100] distributions. At low strains, [010] axis distributions are more clustered with \( K > 1 \). Increasing strain is associated with a decrease in \( K \) and more pronounced [010] girdles, in relatively good agreement with laboratory-based observations [Hansen et al., 2014]. Most dunites and harzburgites exhibit a similar evolution of \( K \), but three harzburgites (JP13-D07, 3924J03a, and 3924J09a) clearly exhibit more clustered textures than the rest of the sample suite despite relatively large shear strains (~3 to 4). Notably, these anomalous [010] distributions are
Figure 3. Crystallographic orientations of olivine grains from all samples in this study. Textures are organized with dunites on the left and harzburgites on the right and shear strain increasing downward. Dots indicate the orientations of individual grains (one point per grain) and are colored by the corresponding value of the orientation distribution function. Stars indicate the average orientation, $\lambda_1$, and squares indicate pole to the plane that best fits the distribution, $\lambda_3$. All pole figures are lower hemisphere, equal-area projections.
found in samples that also exhibit girdled [100] distributions. Olivine aggregates with similarly clustered [010] axes and girdled [100] axes have been documented in both naturally and experimentally deformed olivine aggregates [e.g., Isma...]

Table 1. Results From Textural Analysis

<table>
<thead>
<tr>
<th>Sample</th>
<th>Shear Strain</th>
<th>Rock Type</th>
<th>( \lambda_1 ) for [100]</th>
<th>( \lambda_3 ) for [010]</th>
<th>( \lambda_3 ) for [001]</th>
<th>Shape Parameter (K)</th>
<th>M-index</th>
<th>J-index</th>
<th>Symbol in Figures</th>
</tr>
</thead>
<tbody>
<tr>
<td>3923 J06</td>
<td>0.32</td>
<td>harzburgite</td>
<td>50</td>
<td>37</td>
<td>50</td>
<td>0.2</td>
<td>1.2</td>
<td>0.1</td>
<td>0.08</td>
</tr>
<tr>
<td>3923 J07</td>
<td>0.32</td>
<td>dunite</td>
<td>73</td>
<td>80</td>
<td>89</td>
<td>0.8</td>
<td>1.1</td>
<td>1.9</td>
<td>0.20</td>
</tr>
<tr>
<td>3923 J09</td>
<td>0.81</td>
<td>harzburgite</td>
<td>59</td>
<td>52</td>
<td>71</td>
<td>4.3</td>
<td>0.3</td>
<td>0.2</td>
<td>0.16</td>
</tr>
<tr>
<td>3923 J08</td>
<td>0.81</td>
<td>dunite</td>
<td>42</td>
<td>60</td>
<td>13</td>
<td>2.4</td>
<td>1.2</td>
<td>0.6</td>
<td>0.12</td>
</tr>
<tr>
<td>3923 J13</td>
<td>1.18</td>
<td>harzburgite</td>
<td>36</td>
<td>41</td>
<td>37</td>
<td>2.9</td>
<td>0.3</td>
<td>0.5</td>
<td>0.12</td>
</tr>
<tr>
<td>3923 J12</td>
<td>1.18</td>
<td>dunite</td>
<td>41</td>
<td>46</td>
<td>4</td>
<td>1.4</td>
<td>0.5</td>
<td>6.5</td>
<td>0.17</td>
</tr>
<tr>
<td>3924 J06</td>
<td>2.58</td>
<td>harzburgite</td>
<td>9</td>
<td>8</td>
<td>12</td>
<td>1.8</td>
<td>0.3</td>
<td>1.5</td>
<td>0.18</td>
</tr>
<tr>
<td>3924 J05</td>
<td>2.58</td>
<td>dunite</td>
<td>25</td>
<td>18</td>
<td>66</td>
<td>2.8</td>
<td>0.3</td>
<td>0.6</td>
<td>0.13</td>
</tr>
<tr>
<td>JP13-D07</td>
<td>2.86</td>
<td>harzburgite</td>
<td>32</td>
<td>29</td>
<td>44</td>
<td>0.7</td>
<td>1.0</td>
<td>1.6</td>
<td>0.15</td>
</tr>
<tr>
<td>JP13-D06</td>
<td>2.86</td>
<td>dunite</td>
<td>35</td>
<td>37</td>
<td>38</td>
<td>0.7</td>
<td>0.1</td>
<td>3.4</td>
<td>0.09</td>
</tr>
<tr>
<td>3924 J03a</td>
<td>3.37</td>
<td>harzburgite</td>
<td>28</td>
<td>22</td>
<td>85</td>
<td>0.5</td>
<td>1.1</td>
<td>1.3</td>
<td>0.19</td>
</tr>
<tr>
<td>3924 J03b</td>
<td>3.37</td>
<td>dunite</td>
<td>39</td>
<td>30</td>
<td>55</td>
<td>6.4</td>
<td>0.4</td>
<td>1.1</td>
<td>0.27</td>
</tr>
<tr>
<td>3924 J09b</td>
<td>3.86</td>
<td>dunite</td>
<td>17</td>
<td>14</td>
<td>26</td>
<td>1.2</td>
<td>0.1</td>
<td>0.5</td>
<td>0.18</td>
</tr>
<tr>
<td>3924 J08</td>
<td>5.25</td>
<td>harzburgite</td>
<td>27</td>
<td>29</td>
<td>30</td>
<td>1.3</td>
<td>0.3</td>
<td>0.2</td>
<td>0.14</td>
</tr>
<tr>
<td>3924 J07</td>
<td>5.25</td>
<td>dunite</td>
<td>25</td>
<td>25</td>
<td>34</td>
<td>4.4</td>
<td>0.1</td>
<td>0.3</td>
<td>0.26</td>
</tr>
</tbody>
</table>

Figure 4. Results of eigenvalue analysis assessing the evolution of olivine textures as a function of shear strain for dunites (red) and harzburgites (black). (top row) Texture orientations are quantified based on the angle between an eigenvector and the shear direction [e.g., Zhang and Karato, 1995]. The eigenvector corresponding to the average direction, \( \lambda_1 \), is used for [100] axis distributions, and the eigenvector corresponding to the normal to the best fit plane (i.e., girdle orientation), \( \lambda_3 \), is used for [010] and [001] axis distributions. (bottom row) Texture shapes are quantified using the values of K, which represent more clustered distributions for \( K > 1 \) and more girdled distributions for \( K < 1 \), with the dashed line corresponding to \( K = 1 \). The gray regions correspond to the approximate range of values determined for laboratory-based deformation of olivine aggregates [Hansen et al., 2014]. The symbol shape and color correspond to those in Tables 1 and 2.
The overall strength of olivine textures quantified with the M-index and J-index is depicted for both harzburgites and dunites in Figure 5. In agreement with the observations of Warren et al. [2008], there is no systematic increase in texture strength with increasing amounts of shear strain. However, dunites exhibit significantly more variation in texture strength as well as the strongest textures at high strain.

### 3.2. Variations in Subgrain and Grain Sizes

Measurements of mean intercept lengths provide representations of both subgrain and grain sizes in crystalline materials [e.g., Humphreys, 2001]. Figures 6a and 6b present the measured mean intercept length as a function of the minimum misorientation angle, $\theta$, used to define the minimum misorientation angle counted as an intercept. To be clear, the linear intercept method works by defining a minimum misorientation angle, and misorientation angles equal to and above this value are taken to indicate an intercept (the beginning/end of a line segment whose length is the intercept length). Figures 6a and 6b therefore explore the manner in which the chosen definition of minimum misorientation influences the measured mean intercept length.

The mean intercept length gradually increases from near zero with increasing values of $\theta$ until an inflection point is reached at $\theta \approx 5^\circ$, above which the dependence of intercept length on $\theta$ is minimal. Above this inflection point, most of the intercepts are associated with grain boundaries. Decreasing values of $\theta$ below this inflection point incorporate increasing numbers of intercepts associated with subgrain boundaries. To measure the subgrain size, we need to know the location of all subgrain boundaries. Values of $\theta < 0.5^\circ$ should not be used as this value is at the resolution of EBSD data as determined by Humphreys et al. [2001] (see section 2.2 and Appendix A). However, if the value of $\theta$ is set too high, then subgrain boundaries with small misorientation angles are ignored. Arguably, at values of $\theta > 5^\circ$, enough subgrain boundaries are ignored that we are really only characterizing full grain boundaries. This is the basic concept plotted in Figure 6a and a major point presented by Valcke et al. [2006] for calcite. Hence, we use a value of $\theta = 1^\circ$ for subgrain boundaries (and $\theta = 10^\circ$ for grain boundaries) as this minimum misorientation is sufficiently low to capture most of the subgrain boundaries without introducing errors from noise in the orientation measurements. As described in Appendix A, a value of $\theta = 1^\circ$ also best replicates predictions of subgrain size in samples deformed in the laboratory at known stresses (for which subgrain size can be independently predicted using the piezometry discussed below).

Based on the measured intercept lengths, systematic differences in subgrain size and grain size exist between harzburgites and dunites. Figures 6c and 6d depict the mean intercept length in harzburgites normalized by the corresponding value in dunites as a function of $\theta$. For each harzburgite/dunite pair, these normalized values are greater than unity at values of $\theta < 2^\circ$ and less than unity at values of $\theta > 5^\circ$. This clear transition in...
relative intercept length demonstrates that grain sizes are larger in dunites, but subgrain sizes are larger in harzburgites. This relationship is highlighted in Figure 7, in which the mean intercept length at $\theta = 10^\circ$ is plotted as a function of mean intercept length at $\theta = 1^\circ$. The gray lines connecting harzburgite/dunite pairs all exhibit a negative slope, again demonstrating that grain sizes are larger in dunites while subgrain sizes are larger in harzburgites.

### 3.3. Estimates of Differential Stress

For viscously deforming rocks, commonly used piezometers relate recrystallized grain size to the macroscopic differential stress through

$$\sigma = \frac{B}{d^q},$$

where $\sigma$ is the differential stress, $d$ is the average size of dynamically recrystallized grains, and $B$ and $q$ are empirically derived constants. Although more sophisticated models relating grain-size evolution to thermomechanical conditions have been calibrated for olivine aggregates [Hall and Parmentier, 2003; Kellermann Slotemaker, 2006; Austin and Evans, 2007], we assume that the approach in equation (2) is sufficient for our purposes since deformation conditions vary little throughout our sample suite. In general, however, grain-size piezometers are only applicable to monophase aggregates as the presence of a secondary phase can affect the grain-size evolution of the primary phase, mainly through pinning of migrating grain boundaries [e.g., Evans et al., 2001]. As an alternative, the average stress within a single phase can be estimated using piezometers that rely on intragranular structures. Applicable structures in
olivine-rich rocks include subgrain size and dislocation density [e.g., Nicolas, 1978]. Subgrain-size piezometry is ideal for our data set as subgrain size is quickly estimated from EBSD maps (e.g., Figure 6) and can be applied to our samples regardless of pyroxene content. Subgrain size piezometric relationships generally have a similar form to equation (2), given by

$$\sigma_{ol} = \frac{C}{d_{sg}^p}.$$  

where the subscript ol indicates the property averaged among olivine grains, $d_{sg}$ is the subgrain size, and $C$ and $p$ are empirically derived constants. Such a relationship has been calibrated by various authors based on both experimentally and naturally deformed olivine-rich rocks [Goetze, 1975; Mercier, 1976; Durham et al., 1977; Toriumi, 1979; Karato et al., 1980].

Equations (2) and (3) apply equally well to monophase aggregates, so they should yield similar results for stress estimates in the Josephine dunites. Figure 8 compares stresses calculated using dunite grain size to stresses calculated using subgrain size for five different subgrain piezometric relationships. The two stress estimates are best in agreement using the piezometer of Toriumi [1979], consistent with our comparison of experiments deformed at known stresses to subgrain-size piezometers (described in Appendix A). This agreement is not surprising for the following reasons. (1) The subgrain-size piezometer of Mercier [1976] was calibrated using the measurements from standard, undecorated thin sections, in which detection of very low angle boundaries is difficult. (2) The subgrain-size piezometer of Karato et al. [1980] was presented as “a rough estimation of the stress dependence” due to a heterogeneous subgrain-size distribution. As this study was the only one of those evaluated to not assume the exponent $p = 1$, but instead empirically fit $p = 1.5$, significant error is introduced as $p$ has a large influence on extrapolation to geological stresses. (3) The piezometer of Toriumi [1979] is the only one to include naturally deformed samples in the calibration. We thus used the piezometer of Toriumi [1979] to calculate stresses in harzburgites and dunites from SZP. All calculated stresses are presented in Table 2. It is important to note that stresses estimated with this method only reflect the average stress experienced by olivine grains and not necessarily the macroscopic stress applied to the entire sample.

The grain-size and piezometric analysis reveals systematic variations as a function of pyroxene content. Both the average grain size (Figure 9b) and the average stress (calculated from the subgrain-size piezometer; Figure 9a) in olivine decrease as a function of pyroxene fraction. An increase from 0% to 35% pyroxene results in ~40% decrease in the average stress in olivine. Over this same range in pyroxene content, the average olivine grain size also decreases by ~50% (Figure 9b). As both average stress and grain size decrease with increasing pyroxene fraction, this observation indicates that these two parameters are also positively correlated, as demonstrated on the deformation mechanism maps discussed in section 4.3.

4. Discussion

4.1. Variations in Crystallographic Texture Development

In general, the development of olivine crystallographic textures is very similar in harzburgites and dunites. The average orientation of each axis does not exhibit systematic differences between these two lithologies
The observed texture evolution with increasing strain agrees well with those observed in laboratory experiments on dry, nominally melt-free olivine aggregates [Zhang and Karato, 1995; Bystricky et al., 2000; Zhang et al., 2000; Skemer et al., 2012; Hansen et al., 2014]. Additionally, the evolution of the shape of crystallographic axis distributions with increasing strain, as illustrated by the parameter $K$ (Figure 4), is similar between most harzburgites and dunites, in agreement with laboratory experiments on dunites [Hansen et al., 2014].

The similar textural evolution of harzburgites and dunites suggests that olivine is deforming by the same microphysical mechanism. Numerical simulations have found that subtle changes in the contribution of different deformation mechanisms to the total strain can affect the resulting texture in an observable manner [e.g., Kaminski et al., 2004]. Therefore, we suggest that the rheological behavior of olivine in harzburgites is the same as in dunites. Furthermore, the laboratory experiments in which olivine textural evolution was documented by Hansen et al. [2014] were conducted under conditions that promoted dislocation-accommodated grain-boundary sliding (disGBS) as the dominant deformation mechanism. Thus, the similarity in olivine textural evolution between our samples and the experimentally deformed aggregates implies that olivine flow laws calibrated for disGBS [e.g., Hirth and Kohlstedt, 2003; Hansen et al., 2011] are applicable to our samples. The potential operation of this deformation mechanism is discussed further in section 4.2.

Three harzburgite samples (JP13-D07, 3924J03a, and 3924J09a) exhibit texture shapes that are distinctly different from those observed throughout the rest of the sample suite. The primary difference is in the degree of girdling in the [100] and [010] axis distributions. In these three harzburgites, [100] and [010] axis distributions tend to be girdled and clustered, respectively, which is opposite to the [100]
and [010] axis distributions of the rest of our samples. This difference in the degree of girdling probably accounts for the lower $M$- and $J$-indices in these harzburgites relative to their corresponding dunites (Figure 5). We suggest two possible explanations for these differences. First, clusters of [010] axes and girdles of [100] axes are expected for deformation dominated by axial compression. Similar textures have been observed in both numerical [Tommasi et al., 1999] and experimental [Aveallemant and Carter, 1970; Hansen et al., 2011; Keefner et al., 2011] investigations of compressional deformation of olivine aggregates. Thinning of harzburgite layers during the formation of SZP could potentially introduce a component of compression, but it is not clear by what mechanism thinning would be localized to a small subset of the harzburgite layers sampled. Alternatively, similar textures have been associated with natural [Higgie and Tommasi, 2012, 2014] and experimental deformation of olivine-rich rocks in the presence of a melt phase [Holtzman et al., 2003]. Although shear zones in the Josephine Peridotite are thought to be related to the presence of a partial melt, harzburgite layers are suggested to be melt poor relative to dunite layers [Kelemen and Dick, 1995]. Thus, locally high melt fractions in a select number of harzburgite layers seems improbable. The mechanism for developing the observed anomalous textures is therefore currently unclear, and for the subsequent mechanical analysis, we assume that deformation in all samples occurred by the same deformation mechanism.

4.2. Impact of Pyroxene on Rock Viscosity

4.2.1. Relative Viscosities of Olivine and Pyroxene

A key parameter for predicting mantle rheology is the viscosity of pyroxene relative to olivine. Our measurements of stresses in olivine can be used in a mechanical analysis of olivine-pyroxene aggregates to derive the relative viscosity of pyroxene. This derivation relies on two important characteristics of the SZP peridotites. First, macroscopic structures—such as the lack of boudins (Figure 1)—imply that harzburgites and dunites have similar viscosities. Second, crystallographic textures indicate that olivine is deforming by the same deformation mechanism in both harzburgites and dunites. These observations simplify what is otherwise a significantly underconstrained problem. Here we assess the effect of pyroxene content on the rheological behavior of peridotite by casting the relationships between the macroscopic stress and strain rate in terms of the average stresses within olivine grains. In this analysis, we consider all pyroxene to be a single phase for simplicity, as clinopyroxene occurs at relatively small modal fractions (< 5%) in our sample suite.

We are primarily interested in two ratios, the ratio between the viscosities of pyroxene grains and olivine grains, $X$, and the ratio between the viscosities of harzburgites and dunites, $\Sigma$. These parameters are defined as

$$X = \frac{\eta_{px}}{\eta_{ol}} = \frac{\sigma_{px}\epsilon_{col}}{\epsilon_{px}\sigma_{col}}$$

(4)
and

$$\Sigma = \frac{\eta_{hz}}{\eta_{dn}} = \frac{\sigma_{hz}\varepsilon_{dn}}{\sigma_{dn}\varepsilon_{hz}} \quad (5)$$

where $\eta$ is the viscosity, $\varepsilon$ is the strain rate, and the subscripts px, ol, hz, and dn denote the values for pyroxene grains, olivine grains, harzburgites, and dunites, respectively. Because we observe that the harzburgite/dunite layering in SZP undergoes homogeneous deformation and does not exhibit measurable boudinage or folding (Figure 1), we assume that the viscosity of any given harzburgite is equal to that of its accompanying dunite [e.g., Ramsay, 1980; Herwegh et al., 2011]. An additional assumption is required regarding the distribution of stress and strain rate among the phases within a polyphase rock. Two extreme bounds can be defined, either strain rate within any harzburgite or dunite is homogeneously distributed among all phases (Taylor model) or stress is homogeneously distributed among all phases (Sachs model). Although laboratory observation and theoretical consideration suggest that the aggregate viscosity can be intermediate to these models, especially when the fractions of the two phases are nearly equal [e.g., Ji et al., 2001], we simplify our analysis by first considering the case of homogeneous strain rate.

For the case in which strain rate is homogeneously distributed among all phases, $X$ reduces to

$$X^T = \frac{\sigma_{px}}{\sigma_{ol}} \quad (6)$$

where the superscript $T$ denotes the Taylor model assumption. For this condition, the stresses in harzburgites and dunites can be related to the volume factions of pyroxene and olivine, which is described by

$$\sigma_{hz} = f_{px,hz}\sigma_{px,hz} + f_{ol,hz}\sigma_{ol,hz} \quad (7)$$

and

$$\sigma_{dn} = f_{px,dn}\sigma_{px,dn} + f_{ol,dn}\sigma_{ol,dn} \quad (8)$$

where $f$ is the volume fraction of a particular phase and the combinations of subscripts denote the phase in a specific rock type.

Equations (4) through (8) can be combined to yield

$$\Sigma^T = \frac{a}{b} \left( \frac{f_{px,hz}X + f_{ol,hz}}{f_{px,dn}X + f_{ol,dn}} \right), \quad (9)$$

where

$$a = \frac{\sigma_{ol,hz}}{\sigma_{ol,dn}} \quad \text{and} \quad b = \frac{\varepsilon_{ol,hz}}{\varepsilon_{ol,dn}}. \quad (10)$$

Because of the apparent isoviscous deformation of harzburgites and dunites, $\Sigma$ is taken to be unity, and equation (9) can be rearranged to define $X^T$ as

$$X^T = \frac{abf_{ol,hz} - f_{ol,dn}}{f_{px,dn} - abf_{px,hz}} \quad (11)$$

Values of $a$ can be calculated directly from estimates of stress in olivine (Table 2) and range from 1.05 to 1.30 in our samples. Because the deformation of harzburgite/dunite layering is observed to be homogeneous, $b$ is taken to be unity, and therefore $X^T$ can be calculated directly from microstructural measurements of $a$, $f_{ol}$, and $f_{px}$. The assumption of homogeneous strain rate ($b = 1$) also means that values of $X^2$ are not influenced by nonlinearity in the relationship between stress and strain rate as the relative viscosity is contained in the ratio of stresses ($a$). Values of $X^2$ are given in Table 2 and range from 1.1 to 3.3, indicating that the viscosity of pyroxene is up to ~3 times more viscous than olivine at the deformation conditions relevant to this sample suite.

These values depend on the assumption that dunites and harzburgites are isoviscous, which we assume based on the lack of folding or boudinage among harzburgite and dunite layers. Some constraints on this assumption can be derived from theoretical and experimental investigations of the mechanics of folding. Biot [1961] suggested that a linear viscous layer would not undergo significant folding if its viscosity was less than 100 times that of the matrix. However, it has been demonstrated experimentally and numerically...
[e.g., Hudleston, 1973; Llorens et al., 2013] that folding occurs at much smaller viscosity contrasts (such as viscosity contrasts of 10:1). Therefore, the lack of observable folding in the Josephine Peridotite is a demonstration of very low viscosity contrast between harzburgites and dunites. To place bounds on possible values of that viscosity contrast, we suggest that folding with a ratio of amplitude to layer thickness of 0.1 is a conservative estimate of the magnitude of folding that might be resolvable in the Josephine Peridotite. Following Schmalholz and Podladchikov [2001], see their Figure 6, such folding would correspond to a maximum possible viscosity contrast between harzburgites and dunites of ~2. This value reduces to ~1.2 assuming ductile rheology and a stress exponent of 3. A further reduction in predicted viscosity contrast would arise from considering multiple competent layers (with a spacing on the order of the layer thickness) rather than a single competent layer [Schmid and Podladchikov, 2006]. Notably, Tikoff et al. [2010] investigated viscosity contrast among ultramafic units in the Twin Sisters massif using folding geometry and similarly determined that harzburgites and dunites were effectively isoviscous.

Our results are only marginally affected by the error in the assumption of harzburgites and dunites being isoviscous. We can assess this effect by varying $\Sigma^T$ and $b$ in equation (9), which were set to unity in the above analysis. No change in $X$ occurs as long as $\Sigma^T = b^{-1}$, which corresponds to the case in which any difference in viscosity of harzburgites and dunites results entirely from differences in strain rates rather than stresses. If we consider microstructural measurements from the samples with the highest value of $X$ [3924 J05 and 3924 J06], allowing the values of $\Sigma^T$ and $b$ to vary within a factor of 1.2 yields a range of values for $X$ between 0.3 and 10.7. Therefore, our estimates of the contrast in viscosity between olivine and pyroxene are likely constrained to within 1 order of magnitude.

Values of $X$ can potentially be calculated assuming a homogeneous stress distribution among phases (Sachs model). However, because harzburgites and dunites are observed to deform both homogeneously and isoviscously, the Sachs model would suggest that $\sigma_{ol,dn} = \sigma_{ol,hz}$. This is clearly not the case since (1) the subgrain size (and therefore stress) measured in each dunite is larger than that in the corresponding harzburgite (Figure 7) and (2) a systematic variation in stress is observed with changing volume fraction of pyroxene (Figure 9a). Thus, we conclude that of the two extreme bounds, the Taylor model best describes the observed microstructures. This suggestion agrees with recent laboratory experiments on mixtures of forsterite and enstatite that indicate that the aggregate viscosity is best approximated with the Taylor model over a wide range in pyroxene content [Tasaka et al., 2013].

We compare our estimates of $X$ to laboratory-derived flow laws in Figure 10. A number of flow laws have been calibrated for orthopyroxene [Raleigh et al., 1971; Ross and Nielsen, 1978; Mackwell, 1991; Ji et al., 2001; Bystricky et al., 2011; Ohuchi et al., 2011; Tasaka et al., 2013]. In Figure 10a, we plot the viscosity predicted by these flow laws as a function of temperature. We assume a differential stress of 10 MPa, which is within the range of stresses calculated for olivine in this study (Figure 9). Clearly, there is significant disagreement between published orthopyroxene flow laws, covering a range of viscosities of 5 to 20 orders of magnitude. Figure 10a also presents olivine viscosity calculated from a constitutive model composed of flow laws for diffusion creep, dislocation creep, disGBS, and low-temperature plasticity [Hirth and Kohlstedt, 2003; Mei et al., 2010; Hansen et al., 2011]. In Figure 10b, we calculate the ratio of pyroxene viscosity to olivine viscosity, $X$, for each pyroxene flow law. We also plot the range of values estimated from our observations of SZP peridotites, which cover a temperature range of 900–1100°C [Loney and Himmelberg, 1976; Harding, 1988] and an order of magnitude range in $X$. Despite the relatively large uncertainties in our field data, our results are only explained by a single pyroxene flow law, that of Bystricky et al. [2011]. Importantly, this is the only orthopyroxene flow law that intersects the olivine flow law in the 800–1400°C temperature range, with this intersection indicating that orthopyroxene is the strong phase at temperatures approximately <1150°C. The suitability of the flow law from Bystricky et al. [2011] likely results from the use of a high-resolution gas-medium apparatus for the experiments. Other experimental studies yielding pyroxene flow laws were influenced by a variety of factors, including a phase transformation from orthoenstatite to protoenstatite at low confining pressure [Mackwell, 1991; Ji et al., 2001], uncertainties in stress and temperature associated with using a solid-medium apparatus [Raleigh et al., 1971; Ross and Nielsen, 1978; Ohuchi et al., 2011], and elevated water contents [Ross and Nielsen, 1978]. For completeness, we also plot flow laws for Fe-free forsterite and enstatite from Ji et al. [2001] and Tasaka et al. [2013] in Figure 10. These studies predict a similar value of $X$ to that determined here. However, we note that these experiments were conducted under conditions in
4.2.2. The Role of Grain Size in Moderating Viscosity

The rheological analysis above suggests that pyroxene is several times more viscous than olivine at the conditions relevant to SZP. However, high pyroxene fractions (up to 35% in our sample suite) do not demonstrably change the macroscopic viscosity, as evidenced by the lack of folding or boudinage. Thus, we suggest that pyroxene also affects rock viscosity through its influence on the olivine grain size, which counteracts the inherent strength of pyroxene through a reduction in the viscosity of olivine.

Secondary phases can influence primary phase grain size evolution by impeding the migration of grain boundaries [e.g., Evans et al., 2001; Herwegh et al., 2011], a theory which stems from the early work of Zener [Smith, 1948]. In this scenario, the grain size of the primary phase depends on the grain size, volume fraction, and topology of the secondary phase. Several studies have quantitatively demonstrated that the grain size of olivine in deforming, fine-grained peridotites is controlled by the pyroxene microstructure [Linchens et al., 2011; Tasaka et al., 2013, 2014]. Based on these observations, a significant pyroxene fraction will tend to yield smaller average olivine grain sizes than expected for monophase aggregates. If the deformation mechanism is grain-size sensitive, then the addition of pyroxene would conceivably reduce the viscosity of olivine indirectly through its effect on the olivine grain size. Pinning of olivine grains by pyroxene has been indirectly through its effect on the olivine grain size. Pinning of olivine grains by pyroxene has been
suggested in field-based, laboratory-based, and numerically based studies as a mechanism for extreme rheological weakening and therefore the initiation of lithospheric-scale shear zones [Warren and Hirth, 2006; Skemer et al., 2010; Toy et al., 2010; Bercovici and Ricard, 2012; Farla et al., 2013].

Figure 9b demonstrates that the average olivine grain size in our samples depends on the pyroxene content. We can further quantify this effect using the Zener relationship, $d_{ol} \propto d_{px}/f^Z_{px}$, where $Z$ is the Zener parameter and is dependent on the topology of the pinning phase. Following Tasaka et al. [2014], we compare our data for harzburgites to those of field and laboratory studies in Figure 11 using a value of $Z = 0.52$. Our data are in reasonable agreement with previous studies in which olivine grain size was determined to be controlled by pyroxene [Linckens et al., 2011; Tasaka et al., 2013, 2014], indicating that the pyroxene content in harzburgites in SZP is the primary control on olivine grain size.

We suggest that the reduction in olivine grain size associated with increased pyroxene fraction reduces the olivine viscosity (due to grain size sensitive deformation) by enough to nullify any increase in viscosity associated with the strength of pyroxene. In the previous section, subgrain size piezometry was used as a basis for estimating pyroxene viscosity relative to olivine viscosity. This analysis did not assume that the olivine viscosity in harzburgites is the same as the olivine viscosity in dunites. In fact, the observation of values of $a \neq 1$, in conjunction with the assumption that deformation is homogeneous, necessitates that the viscosity of olivine differs between harzburgites and dunites. To quantify the degree to which differences in olivine grain size affect the macroscopic viscosity, we use the standard power law relationship that describes the solid-state creep of most geological materials:

$$\dot{\varepsilon} = A^{\alpha n},$$

(12)

where $A$ is a material constant that depends on a variety of state variables including temperature and oxygen fugacity, $n$ is the stress exponent, and $p$ is the grain size exponent, all for a given deformation mechanism. Combining equations (10) and (12), $a$ can be expressed in terms of parameters that are measurable from the microstructure and the two exponents, $n$ and $p$,

$$a = \left( \frac{d_{ol,nw}}{d_{ol,dn}} \right)^{\frac{1}{2}}.$$

(13)

Figure 12 presents $p$ as a function of $n$, calculated by rearranging equation (13) and using measured values of $a$ and $(d_{ol,nw}/d_{ol,dn})$ from the Josephine harzburgite/dunite pairs. We calculated combinations of $p$ and $n$ used in flow laws for diffusion creep, dislocation creep, and dislocation-accommodated grain-boundary sliding (disGBS) (Figure 12). Our field measurements suggest that olivine deformation in SZP is best in agreement with flow laws that include $n > 1$ and $p > 0$. Thus, we suggest that deformation of olivine at conditions relevant to SZP is controlled by disGBS.

We explore two sets of flow law parameters for disGBS in Figure 12 ($n = 2, p = 2$ and $n = 3, p = 1$). These values are based on the two regimes proposed by Langdon [1994], in which the rheological behavior stems from either the absence or presence of subgrains. Both types of rheological behavior have been observed in laboratory experiments on olivine-rich aggregates. Hiraga et al. [2010] measured flow law parameters with $n = 2.5$ and $p = 1.5$ for aggregates of forsterite + periclase and $n = 1.5$ and $p = 1.5$ for aggregates of forsterite + diopside + enstatite. The results of Hiraga et al. [2010] are in relatively good agreement with the no-subgrain
model \( (n = 2, p = 2) \) from Langdon [1994], especially considering the lack of any observed subgrain structure. In coarser-grained samples, Hansen et al. [2011] and Hansen et al. [2012] measured flow law parameters with \( n = 2.9 \) and \( p = 0.7 \) for aggregates of San Carlos olivine and \( n = 4.1 \) and \( p = 0.7 \) for aggregates of Fe-rich olivine, respectively. The results of Hansen et al. [2011, 2012] are in relatively good agreement with the with-subgrain model \( (n = 3, p = 1) \) of Langdon [1994], especially considering the prevalence of subgrain structure. The well-developed subgrain structure in the peridotites from SZP (Figure 2) indicates that the with-subgrain model is more applicable to olivine in SZP (Figure 12). Furthermore, the strength of olivine textures and the details of textural evolution in peridotites from SZP are in much better agreement with those from the experiments of Hansen et al. [2011, 2012] than those from Hiraga et al. [2010]. Therefore, we suggest that the dominant deformation mechanism in olivine in SZP is most analogous to that operating in the experiments of Hansen et al. [2011, 2012].

We compare our mechanical and microstructural results to published olivine flow laws in the deformation mechanism map presented in Figure 13. For simplicity, we use dry, melt-free flow laws, although small water contents \((-300 \text{ ppm H}/\text{Si})\) have been documented in SZP [Skemer et al., 2013]. All of our data plot near the center of the disGBS field, in agreement with our prediction of the stress and grain-size exponents above. Furthermore, the data are contained in a narrow enough range of stresses and grain sizes that changes in the relative activity of different deformation mechanisms are not predicted to be significant. Thus, we suggest that at these conditions, the main effect of pyroxene on rock viscosity is through the intrinsic grain-size sensitivity of disGBS rather than the relative activities of different deformation mechanisms. Data from dunites cluster near the olivine grain-size piezometer from Karato et al. [1980], and data from harzburgites all plot at grain-sizes smaller than the piezometer. This difference between harzburgites and dunites gives rise to an apparent direct relationship between grain size and stress, in contrast to the inverse relationship typically presented in piezometry. However, as the increased pyroxene content in harzburgites acts to pin olivine grain boundaries, the typical inverse relationship between recrystallized grain size and stress is only expected if the second phase content is constant. The observed relationship among the grain size, second phase content, and the grain-size piezometer is similar to that observed in gabbro mylonites by Mehl and Hirth [2008]. Importantly, the enlarged area depicted in Figure 13b illustrates that our data approximately lie on a line of constant strain rate \((-10^{-12.5} \text{s}^{-1})\), which is consistent with the assumption that strain rates are similar between both harzburgites and dunites (at the layer scale) and olivine and pyroxene (at the grain scale).

The above analysis indicates that increasing pyroxene fraction results in a decrease in olivine viscosity. Such a phenomenon has been evoked as a mechanism for shear zone formation at a variety of scales [Warren and Hirth, 2006; Skemer et al., 2010; Toy et al., 2010; Bercovici and Ricard, 2012; Farla et al., 2013]. The peridotites examined here represent a special case in which the reduction in olivine viscosity is balanced by the addition of more viscous pyroxene. However, some experimental results indicate that this balance is not always achieved since a minimum in aggregate viscosity can occur at intermediate pyroxene fractions [Hitchings et al., 1989; Tasaka et al., 2012]. We propose three scenarios in which the net effect of pyroxene on viscosity may be more substantial. First, as demonstrated by Figure 10 and supported by the field evidence of Tikoff et al. [2010], higher values of \( X \) are increasingly likely at lower temperatures. Therefore, temperatures higher than those relevant to SZP will reduce \( X \) and promote pyroxene-associated weakening. Second, the Zener relationship presented in Figure 11 may be modified by temperature [Lincens et al., 2011] or differential
An increased dependence of olivine grain size on pyroxene fraction will also lead to pyroxene-associated weakening. Third, an increase in grain-size sensitivity of olivine deformation can be realized through a transition to diffusion creep, which will also enhance pyroxene-associated weakening.

Our data in Figure 13 fall well within the disGBS regime, such that a transition to diffusion creep does not appear possible with increasing pyroxene fraction. However, decreasing the temperature or increasing the stress would make the diffusion creep regime significantly more accessible. Predicting the viscosity of peridotites therefore relies on accurate knowledge of the viscosity contrast between olivine and pyroxene, the dependence of olivine grain size on pyroxene fraction, and the conditions necessary to activate different deformation mechanisms.

5. Conclusion

The role of secondary phases in controlling the viscosity of mantle rocks is one of the major difficulties in using laboratory-derived models of viscous deformation to describe geodynamic processes at geological scales. We use a set of harzburgites and dunites from a shear zone exposed in the Josephine Peridotite to gain insight into the role of pyroxene in the deformation of peridotites.

Several key characteristics of these peridotites allow us to quantify the viscosity of pyroxene relative to that of olivine. First, dunites and harzburgites were collected in pairs from within just a few centimeters of each other so that comparisons could be made between rocks that were deformed at very similar conditions. Second, olivine crystallographic texture evolution is similar in both harzburgites and dunites, suggesting that similar deformation mechanisms are operating in both rock types. Third, harzburgite and dunite layers exhibit no folding or boudinage, which indicates that they have similar viscosities and therefore dramatically simplifies subsequent mechanical analysis.

Mapping of the grain and subgrain structures of olivine and pyroxene with EBSD reveals that olivine subgrains are smaller in dunites than in harzburgites, but olivine grains are larger in dunites than in harzburgites. Because subgrain size can be quantifiably related to the differential stress during deformation using an experimental calibration, we conclude that coarse-grained olivine in dunites experienced higher stresses than fine-grained olivine in harzburgites. Using estimates for the differential stress experienced by olivine based on subgrain-size piezometry, our rheological analysis indicates that pyroxene has a viscosity up to 3 times greater than olivine viscosity at the conditions of the shear zone investigated. These results additionally help to determine the appropriate pyroxene flow law for application to lithospheric conditions.
Although pyroxene affects the aggregate viscosity through the contribution from its inherent viscosity, the role of pyroxene in moderating the grain size of olivine has an equally important effect. Our data indicate that (1) olivine at the conditions of SZP is deforming by dislocation-accommodated grain-boundary sliding, a grain-size sensitive deformation mechanism, and (2) olivine and pyroxene grain sizes follow a Zener relationship, indicating that olivine grain size is primarily controlled by the pyroxene volume fraction. Thus, the addition of pyroxene decreases the viscosity of olivine (and therefore of the aggregate) by decreasing the mean grain size of olivine, counteracting the effect of the increased volume fraction of stronger pyroxene. Previous studies have suggested that Zener pinning of olivine can lead to major rheological weakening. Based on our results, we suggest that major weakening may only be possible with either significantly different temperatures than characteristic of SZP, leading to larger viscosity contrast between olivine and pyroxene or large enough macroscopic stresses to reduce both pyroxene and olivine grain sizes enough to promote a transition to a deformation mechanism with extreme grain-size sensitivity, such as diffusion creep.

Appendix A

We calculate the average stresses among olivine grains using the relationship between subgrain size and differential stress. Several subgrain size piezometers have been calibrated [Goetze, 1975; Mercier, 1976; Durham et al., 1977; Toriumi, 1979; Karato et al., 1980], but predicted stresses can vary significantly among the different

![Figure A1](image-url)
calibrations. Significant variation may occur due to differences in grain size measurement techniques or the definition of a subgrain boundary. We assume that the most appropriate grain size measurement technique is the linear intercept method (see Appendix A in Hansen et al., 2011) and address the best definition of a subgrain boundary below.

We carried out our own subgrain-size piezometer calibration on experimentally deformed olivine. Subgrain sizes were measured from EBSD maps (e.g., Figure A1a) of six aggregates of San Carlos olivine deformed at known conditions in a gas-medium deformation apparatus [Hansen et al., 2011]. The detection of subgrain boundaries is depicted in Figure A1b and the measurement of linear intercepts in Figure A1c. We tested three different misorientation angles (0.5°, 1.0°, and 2.0°) in defining the lower limit for a subgrain boundary. Misorientations of 2° and larger are easily resolvable with EBSD, but such a high value will likely result in some boundaries being missed during grain-size measurement. Misorientations of 0.5° are near the angular resolution, even after using a Kuwahara filter [Humphreys et al., 2001].

In Figure A1d, we plot the measured subgrain size for each minimum misorientation as a function of the stress measured during the experiment. These values are compared to predictions from three different subgrain-size piezometers [Mercier, 1976; Toriumi, 1979; Karato et al., 1980]. We suggest that subgrain sizes are best predicted using a minimum misorientation of 1.0° and the piezometer of Toriumi [1979]. Although there is some subjectivity in choosing the best combination of minimum misorientation and piezometer, comparison between Figure A1 and Figure 8 demonstrates that the combination we have chosen allows reasonable prediction of stresses at both laboratory and geologic conditions.

Acknowledgments

This work was greatly improved through discussions with Take Hiraga, Greg Hirth, Peter Huddleston, Peter Kelemen, Miki Tasaka, and Phil Skemer. We also thank Misha Bystricky for sharing pyroxene flow law parameters prior to their publication. The manuscript was greatly enhanced by comments from Andreas Kronenberg and Holger Stünitz. Sample collection was facilitated by a field course led by Greg Hirth and Peter Kelemen. Scott Sitzman provided expert technical advice on data collection and processing. The Barton Lab in the Department of Plant Biology at Stanford University is thanked for access to their SEM, which is supported by the Carnegie Institution of Washington. Requests for EBSD data sets can be made by contacting the corresponding author. This work was supported by Stanford University, NSF grant EAR-1255620 to JM Warren and award 123/718 from the John Fell Oxford University Press Fund to LN Hansen.

References


Hansen, M.-G., P. D. Bons, A. Griera, E. Gomez-Rivas, and L. A. Evans (2013), Single layer folding in simple shear,
Rozel, A. (2012), Impact of grain size on the convection of terrestrial planets,
Linckens, J., R. H. C. Bruijn, and P. Skemer (2014), Dynamic recrystallization and phase mixing in experimentally deformed peridotite,
Linckens, J., M. Herwegh, O. Müntener, and I. Mercolli (2011), Evolution of a polymineralic mantle shear zone and the role of second phases in
Kaminski, É., N. M. Ribe, and J. T. Browaeys (2004), D-Rex, a program for calculation of seismic anisotropy due to crystal lattice preferred


Smith, C. S. (1948), Grains, phases, and Interfaces: An Interpretation of Microstructure,


