Experimental constraints on the strength of the lithospheric mantle

S. Mei,¹ A. M. Suzuki,¹ D. L. Kohlstedt,¹ N. A. Dixon,² and W. B. Durham²

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[1] To provide a better understanding of rheological properties of mantle rocks under lithospheric conditions, we carried out a series of experiments on the creep behavior of polycrystalline olivine at high pressures (\sim 4–9 GPa), relatively low temperatures (673 $\leq T \leq$ 1273 K), and anhydrous conditions, using a deformation-DIA. Differential stress and sample displacement were monitored in situ using synchrotron X-ray diffraction and radiography, respectively. Experimental results were fit to the low-temperature

plasticity flow law, $\dot{\varepsilon} = A_{\rm P} \sigma^2 \exp\left[-\frac{E_{\rm k}(0)}{RT} \left(1 - \sqrt{\frac{\sigma}{\sigma_{\rm P}}}\right)\right]$. On the basis of this analysis, the

low-temperature plasticity of olivine deformed under anhydrous conditions is well constrained by our data with a Peierls stress of $\sigma_P = 5.9 \pm 0.2$ GPa, a zero-stress activation energy of $E_k(0) = 320 \pm 50$ kJ mol⁻¹, and $A_P = 1.4 \times 10^{-7}$ s⁻¹ MPa⁻². Compared with published results for high-temperature creep of olivine, a transition from low-temperature plasticity to high-temperature creep occurs at ~1300 K for a strain rate of ~10⁻⁵ s⁻¹. For a geological strain rate of 10^{-14} s⁻¹, extrapolation of our low-temperature flow law to 873 K, the cutoff temperature for earthquakes in the mantle, yields a strength of ~600 MPa. The low-temperature, high-stress flow law for olivine in this study provides a solid basis for modeling tectonic processes occurring within Earth's lithosphere.

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1. Introduction

[2] The response of the lithosphere to gravitational and thermal loads as well as the role that the lithosphere plays in tectonic processes from subduction to earthquake generation is strongly dependent on rheological properties. For example, lithospheric strength controls the dip angle of the downgoing slab [Billen and Hirth, 2007] as well as the slab penetration of the 670 km discontinuity [Cizkova et al., 2002]. Simulations have demonstrated that lithospheric strength can control the basic mode of mantle convection, leading to either a stagnant lid regime, where the lithosphere is simply a thermal boundary layer, or a plate tectonics regime, where the lithosphere is still much stronger than asthenosphere but sufficiently weak to participate in convection [Grasset and Parmentier, 1998; Richards et al., 2001; Stein et al., 2004; Tackley, 2000]. A large strength contrast between lithosphere and asthenosphere favors the stagnant lid regime. Mars and Venus, for example, fall in this category [Solomatov and Moresi, 1996; Grasset and Parmentier, 1998; Richards et al., 2001].

[3] The flow behavior of the lithospheric mantle is in turn largely controlled by low-temperature plasticity of olivine-

²Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts, USA.

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rich rocks. Over the past few decades, numerous studies have focused on the rheological properties of mantle minerals and rocks at the high temperatures and low differential stresses appropriate for the asthenosphere [Kohlstedt and Goetze, 1974; Durham and Goetze, 1977; Chopra and Paterson, 1981; Karato et al., 1986; Borch and Green, 1989; Mei and Kohlstedt, 2000], yet relatively few investigations have explored deformation at lithospheric conditions in which flow occurs by mechanisms not encountered in high-temperature experiments [Goetze, 1978; Evans and Goetze, 1979; Raterron et al., 2004; Katayama and Karato, 2008]. Therefore, significant uncertainty exists in our knowledge of the strength of the lithosphere as a function of depth and in our understanding of associated tectonic problems [Bürgmann and Dresen, 2008].

[4] From the point of view of microphysical processes, the mechanism of deformation of a crystalline solid at low temperatures differs from that at high temperatures. At low temperatures ($T < 0.6T_{\rm m}$, where $T_{\rm m}$ is the melting temperature), plastic deformation of crystalline solids results mainly from the glide of dislocations limited by the intrinsic resistance of the lattice, with dislocation velocity governed by the rate of kink nucleation and migration [*Frost and Ashby*, 1982]. The lattice resistance to dislocation glide, often referred to as the Peierls stress, is large for a covalent or ionic-covalent solid such as olivine. Thus, the motion of dislocations will proceed only if the applied stress is high enough to overcome the strong lattice barrier. Such deformation, often referred to as low-temperature plasticity [*Frost and Ashby*, 1982], dominates at low temperatures and

¹Department of Geology and Geophysics, University of Minnesota, Minneapolis, Minnesota, USA.

high stresses. At high temperatures ($T > 0.6T_{\rm m}$), thermal energy further assists deformation through the formation and migration jogs, the latter directly involving ionic diffusion. High-temperature deformation involves not only glide but also climb and cross slip of dislocations limited by dislocation-dislocation interactions and diffusion kinetics. Consequently, the stress needed for high-temperature deformation is significantly lower than that for low-temperature deformation.

[5] Few studies of low-temperature plastic deformation of mantle rocks have been attempted because the strengths of olivine and other mantle minerals are very high, necessitating experiments at high confining pressures in order to avoid fracture of the samples. To deform rocks plastically without producing intergranular or intragranular cracks, confining pressure must in general exceed the applied differential stress, a condition referred to as the Goetze criterion [*Evans and Kohlstedt*, 1995]. Until recently, the maximum confining pressure attainable in most experimental deformation instruments was ≤ 3 GPa, thus limiting maximum differential stresses to a similar value if fracture was to be avoided.

[6] In their pioneering study, Evans and Goetze [1979] used microindentation tests to obtain a flow law describing low-temperature plasticity of olivine under anhydrous conditions. This approach was successful because the mean pressure $P_{\rm m}$ under an indenter is about twice the differential stress $(\sigma_1 - \sigma_3)$ [Johnson, 1985, p. 62], hence meeting the Goetze criterion ($\sigma_1 - \sigma_3 < P_m$) and permitting plastic flow without brittle fracture. Meade and Jeanloz [1990] added to this area by examining deformation at high pressures produced in a diamond anvil cell at room temperature. Raterron et al. [2004] further contributed by carrying out load relaxation experiments on polycrystalline olivine at high pressure under anhydrous conditions, using a cubic anvil apparatus. This latter group of researchers employed synchrotron X-ray diffraction to determine stress, and they calculated strain rate from the time rate of change of stress. In none of these studies did deformation take place under steady state mechanical and microstructural conditions, even though the results were analyzed using a strain-independent flow law. Recently, Katayama and Karato [2008] reported a low-temperature, high-stress flow law for olivine deformed under hydrous conditions in simple shear experiments in a Griggs-type deformation apparatus. Strain rate was calculated from the rate of movement of the compression piston, and stress at the end of the deformation experiment was determined from the dislocation density.

[7] With the development of the deformation-DIA (D-DIA), a solid-medium deformation apparatus capable of experiments to a confining pressure of 15 GPa [*Wang et al.*, 2003], it is now possible to carry out steady state deformation experiments that satisfy the Goetze criterion and at the same time operate under stress and temperature conditions appropriate for the lithospheric mantle. In this paper, we describe the results of our investigation of low-temperature plasticity of olivine-rich samples, using the D-DIA in conjunction with synchrotron radiation to determine pressure, differential stress, and sample displacement in situ. The result is a flow law obtained under near-steady state conditions that permits an improved understanding of the rheo-

logical properties of the lithospheric mantle and its role in tectonic processes.

2. Experimental Details

2.1. Sample Preparation and Assembly

[8] The polycrystalline samples used in our deformation experiments were fabricated from powders of San Carlos olivine plus 5% (by mass) enstatite. After mixing, all powders were dried in a vacuum oven at T = 420 K. These powders, particle size ~5 μ m, were first uniaxially cold-pressed into Ni capsules at ~200 MPa and then hydrostatically hot-pressed in a gas-medium apparatus at 300 MPa and 1473 K for 4 h to produce cylindrical rods 9.0 mm in diameter and 14.0 mm in length with a grain size of 5–10 μ m. Several samples 1.1 mm in diameter and 1.2 mm in length were cored from two such rods for D-DIA experiments. All samples were stored in the vacuum oven prior to their use in deformation experiments.

[9] The configuration of the test cell is shown in Figure 1. For each deformation experiment, the sample was wrapped with 25 μ m thick Ni foil to buffer oxygen fugacity. Ni discs of the same thickness placed at either end of the sample functioned as strain markers. The deformation column, composed of a Ni-encapsulated sample plus Al₂O₃ pistons at either end, was placed into a boron nitride sleeve, which was itself positioned inside a sleeve of graphite that served as the resistance heater. The entire assembly was placed inside a pressure medium of mullite to assure an anhydrous environment for the sample. The pressure medium in Figure 1 is a 6 mm cube of hybrid composition [*Durham et al.*, 2009], a 6 mm diameter sphere of mullite, and an outer web of unfired pyrophyllite to provide high-pressure gasketing in the D-DIA.

2.2. Experimental Procedure

[10] Deformation experiments were conducted in the D-DIA apparatus at synchrotron beamline X17B2, National Synchrotron Light Source, Brookhaven National Laboratory. During an experiment, a beam of white X-rays, collimated to an area of $70 \times 70 \ \mu m$, provided the source of diffracted X-rays for determining stress and pressure and also of transmitted X-rays for imaging the sample to monitor sample strain.

[11] From diffraction data, we determined pressure (i.e., mean stress) and differential stress from the elastic strain of various lattice planes measured as a function of orientation with respect to the applied stress field. This technique was initially discussed by *Singh and Balashingh* [1994] and *Singh et al.* [1998]. Application to our testing system is described in detail by *Li et al.* [2004] and *Mei et al.* [2008]. With current technology, the resolution in stress and pressure for each individual measurement is ~50 MPa.

[12] We note that different values for differential stress are obtained from analyses of diffraction from different lattice planes. This behavior occurs because the magnitude of the elastic strain depends on the orientation of the diffracting lattice plane with respect to the operative slip systems, each of which comprises a slip plane and a slip direction; consequently, there is plastic anisotropy. For olivine, given the diffraction geometry imposed by the D-DIA apparatus with a maximum allowed diffraction angle of $2\theta \approx 6.5^{\circ}$, the three



Figure 1. Sketch of sample assembly (not to scale). Olivine sample is encapsulated with $25 \ \mu m$ thick Ni foil and assembled with Al_2O_3 pistons, a boron nitride sleeve, and graphite resistance heater into a 6 mm edge length cubic pressure medium, made of mullite (inside) and pyrophyllite (outside, not shown). Assembled sample is then placed inside D-DIA for deformation.

most prominent diffraction peaks are (hkl) = (130), (131), and (112). The relationship between the value of stress deduced from lattice strain from a group of grains and the value of stress applied externally to a sample, though discussed elsewhere [*Merkel*, 2006; *Merkel and Yagi*, 2006; *Karato*, 2009], has not yet been well established. For present purposes, we approximate the stress supported by the aggregate as the arithmetic average of stresses calculated from the (130), (131), and (112) lattice reflections. Sample strain is computed from the length change of deforming samples measured from a time series of X-radiographic images with an accuracy of few micrometers. Strain rate values given here are accurate to approximately 1% of their stated values.

[13] Temperature at the sample during a deformation experiment was determined from a series of temperature versus heater power calibrations [*Durham et al.*, 2009]. For a sample deformed at several temperatures, the relative uncertainty is ± 10 K. The uncertainty in temperatures from one experiment (sample) to the next is ± 25 K.

[14] Water concentration in our samples was determined by using Nicolet Series II Magna Fourier transform infrared (FTIR) spectroscopy. The system is equipped with a microscope for achieving a high lateral spatial resolution. For a polycrystalline sample prior to deformation, an unpolarized beam ~50 μ m in diameter was focused on crack-free areas of a doubly polished sample section ~0.2 mm thick. Water content was calculated on the basis of baselinecorrected FTIR spectra using Paterson's [1982] calibration. However, reliable FTIR analyses could not be obtained from deformed olivine aggregates because slices of the friable samples had to be impregnated with epoxy to prevent disaggregation after depressurization in the D-DIA, and epoxy introduces significant uncertainties in analyses of FTIR spectra. Therefore, to check the hydrous characteristic of our deformation cell, FTIR analyses were performed on an oriented single crystal that was deformed under conditions similar to those used in the our deformation experiments on polycrystalline samples. For the single crystal, the unpolarized beam was parallel to the [010] crystal axis.

[15] At the start of an experiment, the cell was first pressurized hydrostatically and heated to the run temperature. The sample was then deformed in triaxial compression at constant pressure. Samples were deformed by pushing the top and bottom anvils toward each other. This operation not only deforms the sample but also squeezes the entire sample cell, thus increasing the confining pressure. To maintain a constant pressure throughout the duration of the experiment, we lowered the oil pressure on the side anvils by a few pascals every half hour to maintain an approximately constant confining pressure. In this way, we were able to control pressure during deformation to ± 0.5 GPa.

[16] After an experiment, the deformed sample was retrieved from the pressure medium and then encapsulated with epoxy to prevent it from falling apart. To prepare samples for microstructural observations, undeformed and deformed samples were sectioned and polished. The polishing procedure involves the use of successively finer diamond lapping films (30–0.5 μ m), followed by colloidal silica

Table 1. Summary of Experimental Conditions and Results

		Р	Ė	ε	$\sigma_{(hkl)}$ (GPa)		$\overline{\sigma}$	
Run	$T\left(\mathrm{K} ight)$	(GPa)	(10^{-5} s^{-1})	(%)	(130)	(131)	(112)	(GPa)
San129	1053	8.75	0.88	10.2	3.29	2.81	2.43	2.84
San130	1023	7.25	6.81	23.1	3.50	3.21	2.81	3.17
San149	1073	4.75	2.15	9.5	3.19	2.68	2.22	2.70
San150	873	7.54	2.78	15.2	3.52	3.30	2.69	3.17
San154 ^a	1173	(5.0)	2.41	10.0	3.15	2.86	2.40	2.80
San155 ^a	673	(8.0)	3.11	17.5	4.15	3.78	3.18	3.70
San156	773	5.20	3.49	14.0	4.02	3.67	3.11	3.60
San157	673	9.45	2.79	19.1	4.21	4.00	3.20	3.80
San165	973	5.95	2.43	17.0	3.35	3.22	2.53	3.03
San167 ^a	1073	(6.2)	2.96	17.2	3.24	2.78	2.30	2.77
San177	1173	5.20	2.46	26.1	3.02	2.70	2.28	2.67
San178 ^a	1273	(6.2)	1.61	12.5	2.61	2.28	1.91	2.27
San182	1273	5.45	2.78	28.2	2.91	2.60	1.94	2.48
San189	1073	5.48	0.14	2.6	2.79	2.42	1.82	2.34
San200	1073	4.50	0.10	7.9	2.70	2.40	1.82	2.31

^aPressure in each of these runs fluctuated; the average value of pressure is given in parentheses. Therefore, data from these runs have not been included in data analysis.



Figure 2. Scanning electron micrographs of samples (top) before and (bottom) after deformation (San130). Samples have been chemically etched to reveal grain boundaries. The deformed sample exhibits thermally etched cracks that formed during depressurization at the end of the deformation experiment. Grains in the deformed sample were generally shortened in the direction of the maximum compressive stress σ_1 . The maximum flatting strain of individual grains was approximately equal to that of the sample.

(Syton HT 50). In a few cases, surfaces of polished samples were chemically etched using diluted HF ($H_2O/HF = 10:1$ by vol) to reveal grain boundaries.

2.3. The Flow Law

[17] In the low-temperature plasticity regime, steady state deformation by dislocation glide is often described by the following relation [*Kocks et al.*, 1975, p. 141; *Frost and Ashby*, 1982, p. 6; *Karato*, 2008, p. 175]:

$$\dot{\varepsilon} = A_{\rm P0} \sigma^2 \exp\left\{-\left(\frac{H(0)}{RT}\right) \left[1 - \left(\frac{\sigma}{\sigma_{\rm P}}\right)^p\right]^q\right\},\tag{1}$$

where $\dot{\varepsilon}$ is axial strain rate, A_{P0} is a material-dependent parameter, H(0) is the zero-stress activation enthalpy, R is the gas constant, T is temperature, σ is differential stress, and σ_P is Peierls stress. The σ^2 term arises from the stress-dependence of dislocation density through Orowan's equation. The temperature dependence expressed in the exponential term arises from the need to nucleate pairs of kinks that allow dislocations to glide. The choice of p and q depends on the geometry of the obstacles that resist dislocation glide and are generally bracketed by $0 \le p \le 1$ and $1 \le q \le 2$ [*Frost and Ashby*, 1982, p. 9]. In the analysis of our data, we used values of p = 0.5 and q = 1, motivated by discussions of the shape of the Peierls potential [*Kocks et al.*, 1975, p. 142] and the mechanical model that



Figure 3. Fourier transform infrared (FTIR) spectra of an undeformed sample and a deformed sample. The undeformed (starting) sample contains $<20 \text{ H/10}^6$ Si, whereas the deformed sample has $\sim35 \text{ H/10}^6$ Si.

successfully described low-T deformation of Al₂O₃ and MgAl₂O₄ [*Mitchell*, 1999; *Mitchell et al.*, 1999]. Thus, equation (1) can then be rewritten as

$$\dot{\varepsilon} = A_{\rm P0}\sigma^2 \exp\left\{-\left(\frac{H(0)}{RT}\right)\left[1-\sqrt{\frac{\sigma}{\sigma_{\rm P}}}\right]\right\}.$$
(2)

3. Experimental Results

[18] Experiments were conducted at constant displacement rates ranging from 1.0×10^{-6} to 6.8×10^{-5} s⁻¹ to produce true axial strains of ~3%–28% at temperatures of 673–1273 K and pressures of 4.5–9.5 GPa. Experimental conditions and results for 15 creep tests are summarized in Table 1.

[19] Scanning electron microscope images of predeformation and postdeformation samples are shown in Figure 2. Images reveal a nearly uniform grain size distribution in the starting sample with relatively little grain growth in the deformed sample. Grain boundaries in the starting samples are generally more curved than those of deformed samples. The deformed sample shown exhibits thermally etched cracks that formed during depressurization at the termination of the experiment. Furthermore, the shapes of grains in the deformed sample changed after deformation. As shown in Figure 2, deformed grains were shortened in the direction of the maximum compressive stress σ_1 . In general, the maximum flatting strain of individual grains was approximately equal to that of the sample.

[20] The FTIR spectra in Figure 3 provide a measure of the water content of our undeformed and deformed samples. The undeformed polycrystalline sample contained <20 H/10⁶ Si. The deformed single crystal contained ~35 H/10⁶ Si.

[21] As an illustration of our deformation results, plots of pressure, sample strain, and differential stress versus time for experiment San200 are presented in Figure 4. As shown

in Figure 4a, the confining pressure of ~4.5 GPa remains approximately constant throughout the duration of the experiment (~18 h). The strain rate and stress initially increase with increasing strain and then approach steady state values in ~1 h, as demonstrated in Figures 4b and 4c. The difference in the values of stress calculated from the different (*hkl*) reflections is apparent in Figure 4c. In general, stress calculated based on diffraction from (130) planes is the highest and that calculated from diffraction from (112) planes is the lowest, the two differing by ~1 GPa. The average differential stress obtained from all three reflections is 2.3 GPa.

[22] Our experimental $\dot{\varepsilon}$ - σ -T data in Table 1 were least squares fit to the low-T flow law given in equation (2) in order to determine values for the parameters in the flow law. With the limited amount of data in our study, the pressure



Figure 4. Plots of (a) pressure versus time, (b) strain versus time, and (c) differential stress versus time for experiment San200. As indicated in Figure 4a, the confining pressure of ~4.5 GPa remains approximately constant throughout the duration of the experiment. The strain rate and stress initially increase with increasing strain and then approach quasi steady state values ~1 h after the start of the experiment, as shown in Figures 4b and 4c. Dashed lines are linear least squares fits to data points.



Figure 5. Plot of differential stress versus temperature. Raw data were fit, using a nonlinear least squares method, to equation (3). Data are normalized to a common laboratory strain rate of 3×10^{-5} s⁻¹ using the flow law, as indicated by open circles. The flow law (solid line) is plotted for comparison. The value of the Peierls stress can be obtained by extrapolating the flow law to a temperature of 0 K.

dependencies of the activation enthalpy and the Peierls stress could not be fit; thus, in equation (2), we replaced the zero-stress activation enthalpy with the zero-stress activation energy: $H_k(0) \approx E_k(0)$. Consequently, equation (2) is reformed as

$$\dot{\varepsilon} = A_{\rm P} \sigma^2 \exp\left[-\frac{E_{\rm k}(0)}{RT} \left(1 - \sqrt{\frac{\sigma}{\sigma_{\rm P}}}\right)\right].$$
(3)

[23] A least squares fit of our data to the modified constitutive equation (3) yields $A_{\rm P} = 1.4 \times 10^{-7} \, {\rm s}^{-1} \, {\rm MPa}^{-2}$, $\sigma_{\rm P} = 5.9 \pm 0.2 \, {\rm GPa}$, and $E_{\rm k}(0) = 320 \pm 50 \, {\rm kJ \ mol}^{-1}$. The resulting flow law is plotted with experimental data in Figure 5.

4. Discussion

[24] Analyses of the FTIR spectra in Figure 3 yield relatively small water concentrations of \sim 35 H/10⁶ Si. This amount of water is below that required to affect the deformation behavior of olivine [*Mei and Kohlstedt*, 2000]. Thus, we conclude that our samples were deformed under effectively anhydrous conditions.

[25] Our flow law for a quasi steady state deformation of olivine under low-temperature, high-stress, anhydrous conditions is compared in Figure 6 with published results obtained under similar conditions from microindentation [*Goetze*, 1978; *Evans and Goetze*, 1979] and load-relaxation tests [*Raterron et al.*, 2004]. The high-temperature flow law for creep of olivine reported by *Chopra and Paterson* [1981] and modified by *Hirth and Kohlstedt* [1996] is included for comparison. Values of $E_k(0)$, σ_P , *p*, and *q* for the low-temperature plasticity from all three studies are



Figure 6. Comparison of our low-temperature, high-stress flow law with those reported by *Goetze* [1978], *Evans and Goetze* [1979], and *Raterron et al.* [2004]. The high-temperature flow law from *Chopra and Paterson* [1984], as modified by *Hirth and Kohlstedt* [2003], is included for comparison.

compared in Table 2. On the basis of Figure 6, for a laboratory strain rate of 3×10^{-5} s⁻¹, our low-temperature flow law yields strengths greater than those obtained by *Goetze* [1978] and *Evans and Goetze* [1979], which in turn are greater than those obtained by *Raterron et al.* [2004].

[26] The value of the Peierls stress of 5.9 GPa obtained from our D-DIA data is somewhat smaller than the values of ~9.0 GPa determined in microindentation experiments [*Evans and Goetze*, 1979] and substantially smaller than the value of 15.4 GPa obtained from load relaxation tests by *Raterron et al.* [2004]. The cause of the discrepancy amongst these three values is not certain, but we note that the two previous studies of low-temperature plasticity employed deformation techniques for which deformation did not occur in the steady state, whereas our experiments closely approximated steady state flow, as illustrated in Figure 4. The three studies of low-temperature plasticity of olivine yield quite different values for zero-stress activation energy, as summarized in Table 2. The zero-stress activation energy

 Table 2. Comparison of Flow Law Parameters Obtained From

 Three Low-Temperature Studies

Flow Law Parameter	This Study ^a	Indentation Test ^b	Relaxation Test ^c
Peierls stress, σ_p (GPa)	5.9 ± 0.2	9.1	15.4 ± 1.0
Zero-stress activation energy, E(0) (kJ mol ⁻¹)	320 ± 50	~500	564 ± 89
p	≡1/2	≡1	2/3
q	=1	≡2	2

^aThis study: Parameters were obtained from fit to equation (3).

^bEvans and Goetze [1979]: Parameters were obtained from fit to equation (1) without σ^2 term.

^c*Raterron et al.* [2004]: Parameters were obtained from fit to equation (1) without σ^2 term.



Figure 7. Extrapolation of our low-temperature plasticity flow law from laboratory to lithospheric conditions at T =873 K, shown by the solid line. The dashed lines indicate the bounds associated with the uncertainty in zero-stress activation energy.

determined in our study, 320 kJ mol⁻¹, is significantly smaller than those determined from microindentation tests, 500 kJ mol⁻¹ [Evans and Goetze, 1979], and load relaxation tests, 564 kJ mol⁻¹ [Raterron et al., 2004]. The latter two values are similar to the activation energy obtained for high-temperature creep of olivine, ~530 kJ mol⁻¹ [Chopra and Paterson, 1981]. A difference between the value of zero-stress activation energy for low-temperature plasticity and the activation energy for high-temperature creep is not surprising as the mechanisms of deformation are different between these two regimes. At high temperatures, deformation involves dislocation climb as well as dislocation glide. If deformation is rate-limited by climb, the activation energy for deformation will include an extra contribution from the formation and migration of jogs. At the same deformation regime, the difference between the zero-stress activation energy obtained in our study and that reported previously may be related to differences in the testing methods (i.e., transient deformation versus steady state flow). Moreover, in previous studies [Goetze, 1978; Evans and Goetze, 1979; Raterron et al., 2004], the σ^2 term in equation (1) was not included; omission of that term results in a higher value for the activation energy.

[27] As illustrated in Figure 6, the high-temperature flow law dramatically overpredicts the strength of olivine in the lower temperature regime. Our results indicate a transition from low-temperature to high-temperature creep near 1300 K for a strain rate of $\sim 10^{-5}$ s⁻¹.

5. Geological Implications

[28] The strength of the lithospheric mantle can be predicted by extrapolating our flow law to a lithospheric strain rate. As illustrated in Figure 7, extrapolation of our flow law from a laboratory strain rate of 3×10^{-5} s⁻¹ to a geological strain rate of 1×10^{-14} s⁻¹ yields a strength for the lithospheric mantle at 873 K (the cutoff temperature for earthquakes in the mantle) of ~ 600 MPa. The dash lines in the Figure 7 show the uncertainties for the estimated strength (± 200 MPa) if the uncertainties of activation energy are taken into account.

[29] To further illustrate the application of our flow law describing low-temperature deformation of olivine-rich rocks under anhydrous conditions, a strength envelope for the lithospheric mantle is presented in Figure 8. For comparison, the low-temperature flow laws of Goetze [1978], Evans and Goetze [1979], and Raterron et al. [2004] are included. The *Raterron et al.* [2004] relationship predicts a much weaker lithosphere than does our flow law. In contrast, the strength profiles calculated using the flow laws from Goetze [1978] and Evans and Goetze [1979] indicate strengths greater (less) than that obtained from our lowtemperature flow law at depths less (greater) than 40 and 50 km, respectively. We note, however, that the effects of pressure on the flow strength of olivine through the activation enthalpy and the Peierls stress were not determined in any of these investigations of low-temperature plasticity.

[30] The lithospheric plate, as determined from our experimental results shown in Figure 8, has a depth-averaged strength of ~0.3 GPa. This is consistent with that considered by *Billen and Hirth* [2007] and *Cizkova et al.* [2002] to be the limiting strength in models of slab decent



Figure 8. Strength as a function of depth for the oceanic lithosphere deforming at 10^{-14} s⁻¹, modified from *Kohlstedt and Mackwell* [2009]. Oceanic geotherm used in this plot is from the work of *Turcotte and Schubert* [2002, p. 184] with $T_{surface} = 273$ K. a, b, c, and d: extrapolation of low-*T* plasticity flow laws from the works of *Raterron et al.* [2004], *Evans and Goetze* [1979], *Goetze* [1978], and this study, respectively. e: extrapolation of high-*T* creep flow law from the work of *Chopra and Paterson* [1984], as modified by *Hirth and Kohlstedt* [2003].

in the mantle, where a Peierls or similar mechanism is required to offset the unreasonably high strength of dislocation and diffusion creep mechanisms at low temperature [*Cizkova et al.*, 2002]. This depth-averaged strength is consistent with that obtained from the results of *Evans and Goetze* [1979] and twice that predicted by the flow law of *Raterron et al.* [2004]. A lithosphere composed of dry olivine-rich rocks is thus required over a weaker lithosphere of wet olivine [*Katayama and Karato*, 2008] to explain shallow dip angles of the downgoing slab [*Billen and Hirth*, 2007] as well as penetration of the 670 km discontinuity [*Cizkova et al.*, 2002].

6. Conclusions

[31] Results from our near-steady state measurements of the rheological behavior of polycrystalline olivine at low temperatures, high stresses, and high pressures using a D-DIA indicate the following:

[32] 1. Below ~1300 K at laboratory strain rates $(10^{-6} \text{ to } 10^{-5} \text{ s}^{-1})$, under anhydrous conditions, low-temperature plasticity dominates over high-temperature creep mechanisms.

[33] 2. The flow stress in the low-temperature plasticity regime is significantly less temperature sensitive than reported in earlier studies using microindentation and load relaxation techniques [*Evans and Goetze*, 1979; *Raterron et al.*, 2004].

[34] 3. Our low-temperature plasticity flow law predicts a strength of ~600 MPa for anhydrous lithospheric mantle at 873 K and 10^{-14} s⁻¹. This value is similar to that obtained from the flow law of *Evans and Goetze* [1979] and is sixfold stronger than that calculated from the flow law of *Raterron et al.* [2004].

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D. L. Kohlstedt, S. Mei, and A. M. Suzuki, Department of Geology and Geophysics, University of Minnesota, 310 Pillsbury Dr. SE, Minneapolis, MN 55455, USA. (meixx002@umn.edu)

N. A. Dixon and W. B. Durham, Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, 77 Massachusetts Ave., Cambridge, MA 02139, USA.