Flow strength of wet quartzite in steady-state dislocation creep regimes

Lucy Lu^1 and Dazhi Jiang¹

¹Western University

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Abstract

We quantitatively investigated the flow laws of wet quartzite in steady-state dislocation creep regimes by considering both the dependence of the activation enthalpy on pressure and dependence of the stress exponent on slip systems. From a critically-selected set of creep experiments of wet quartzite with microstructures and c-axis fabrics suggesting steady-state dislocation creeps, we obtained two endmember flow laws corresponding respectively to dominant prism slip and dominant basal slip systems. To characterize the dislocation creep of wet quartzite by a continuous combination of prism and basal slips commonly observed in nature and experiments, we developed a self-consistent micromechanics-based homogenization approach. Our results reconciled the large discrepancies in flow law parameters for wet quartzite determined from different creep experiments and are broadly consistent with microstructures and c-axis fabrics from nature and experiments as well as theoretical considerations.



Figure4



Figure2

1	Flow strength of wet quartzite in steady-state dislocation creep regimes											
2	Lucy X. Lu ^a , Dazhi Jiang ^{a,b,1}											
3	^a Department of Earth Sciences, Western University, London Ontario, Canada N6A 5B7											
4	^b State Key Laboratory of Continental Dynamics, Department of Geology, Northwest University,											
5	Xi'an 710069, China											
6	Key points:											
7 8	• Best-fit flow laws derived from high-quality creep experiments of wet quartzite with consideration of pressure and slip-system dependence											
9 10	 Micromechanics-based homogenization method developed for the creep of wet quartzite by simultaneous basal <a> and prism <a> slips 											
11 12	• Results reconciling discrepancies among flow laws derived from experiments and consistent with observations and theoretical considerations											

¹ corresponding author: djiang3@uwo.ca

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Abstract

We quantitatively investigated the flow laws of wet quartzite in steady-state dislocation creep 14 regimes by considering both the dependence of the activation enthalpy on pressure and 15 dependence of the stress exponent on slip systems. From a critically-selected set of creep 16 experiments of wet quartzite with microstructures and c-axis fabrics suggesting steady-state 17 18 dislocation creeps, we obtained two endmember flow laws corresponding respectively to dominant prism <a> slip and dominant basal <a> slip systems. To characterize the dislocation 19 creep of wet quartzite by a continuous combination of prism <a> and basal <a> slips commonly 20 21 observed in nature and experiments, we developed a self-consistent micromechanics-based homogenization approach. Our results reconciled the large discrepancies in flow law parameters 22 for wet quartzite determined from different creep experiments and are broadly consistent with 23 microstructures and c-axis fabrics from nature and experiments as well as theoretical 24 considerations. 25

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27

Plain Language Summary

Understanding the rheology of continental crust is a fundamental problem related to the physics 28 29 of Earth and geodynamics. Our current understanding of the rheology of continental crust is based on the high-temperature and high-pressure creep experiments on wet quartzite in 30 31 recognition that quartz is a common and strength-controlling mineral in the crust. These 32 experimental data are commonly fitted into a power-law flow law, which is incorporated in many geodynamic models. Unfortunately, past creep experiments on wet quartzite yield very different 33 34 quartz flow law parameters. We critically select 19 high-quality creep experiments on wet 35 quartize and demonstrate that the current dataset is consistent with two endmember flow laws,

corresponding to dominant basal <a> and dominant prism <a> slip systems. In both experiments
and nature, the creep behavior of wet quartzite is a continuous spectrum from dominant basal
<a> to a mixture of basal <a> and prism <a> slip and dominant prism <a> slip. We propose a
self-consistent micromechanics-based approach to determine the creep behavior in such
situations and plot the results in a 3D strength profile. In nature, the temperature, pressure, and
the relative contribution of dominant slip systems profoundly affect the strength of wet quartzite.

43 1. Introduction

An accurate expression for the creep of quartzite is of paramount importance in 44 geodynamics (e.g., Kohlstedt et al., 1995; Ranalli, 1987; Beaumont et al., 2001; Lavier et al., 45 2013; Chowdhury et al., 2017). Despite many decades of effort, our knowledge on this subject is 46 still incomplete. The dislocation creep of wet quartizte has been described by a flow law of the 47 form $\dot{\varepsilon} = A f_w^m \exp\left(-\frac{Q}{RT}\right) \sigma^n$, where $\dot{\varepsilon}$ is the strain rate, A the pre-exponential parameter, f_w 48 the water fugacity, m the water fugacity exponent, Q the activation energy, R the universal gas 49 constant, T the absolute temperature, σ the differential stress, and n the stress exponent. A major 50 problem has been that past experiments have yielded very different values of Q, n, m, and A51 (flow law parameters, hereafter), with n varying between 2 and 4, Q between 130 kJ/mol and 52 53 240 kJ/mol, and m between 0.372 and 2.8 (Kronenberg and Tullis, 1984; Koch et al., 1989; Gleason and Tullis, 1995; Luan and Paterson, 1992; Post et al., 1996; Rutter and Brodie, 2004; 54 Chernak et al., 2009; Holyoke and Kronenberg, 2013). Such a large range of flow law 55 parameters translates to very large uncertainties in the predicted strength of the continental crust 56

and may have profoundly affected the outcome of many geodynamic models where a quartz flowlaw is used.

Recently, Lu and Jiang (2019) have shown that the large difference in *Q* among
experiments may be explained by the pressure dependence of the activation enthalpy. Using the
following flow law (e.g., Karato and Jung, 2003):

$$\dot{\varepsilon} = A f_w^m \exp\left(-\frac{Q+PV}{RT}\right) \sigma^n = A f_w^m \exp\left(-\frac{H}{RT}\right) \sigma^n \tag{1}$$

where P is the pressure, V the activation volume, and H (= Q + PV) the activation enthalpy, 63 they applied an iterative method to the experiment data of Gleason and Tullis (1995), Luan and 64 Paterson (1992), and a set of pressure stepping experiments of Kronenberg and Tullis (1984) to 65 obtain the following set of flow law parameters: n = 4, m = 2.7, V = 35.3 cm³/mol, and Q = 13266 kJ/mol. Furthermore, there is growing evidence that in the dislocation creep regime *n* may 67 depend on the active dominant slip system (Goldsby and Kohlstedt, 2001; Hirth and Kohlstedt, 68 2003) and the reported variation of *n* between 2 and 4 for quartzite may reflect a switch of the 69 dominant slip systems (Tokle et al., 2019). 70

In this contribution, we re-examine the existing high-quality quartz creep experiments data by considering *both* the activation enthalpy dependence on pressure and the stress exponent dependence on dominant slip systems. We select 19 creep experiments on quartz samples with grainsize of 20-200 μ m, deformed from 700 °C to 1200 °C and interpreted to represent steadystate regimes 2 and 3 dislocation creeps (Hirth and Tullis, 1992), to determine the flow law parameters using Eq.1. A detailed description of the selection criteria of previous creep experiments for our analysis is given in supplementary information and Lu and Jiang (2019). We

found that current experimental dataset is consistent with two endmember dislocation creep flow

respectively to a dominant prism a slip and a dominant basal a slip.

80 Furthermore, both experiments and natural quartz-bearing mylonites suggest that both basal $\langle a \rangle$

81 and prism <a> slips can be significant. We develop a self-consistent micromechanics-based

82 method to numerically determine the flow behavior of wet quartzite in such situation.

83 2. Dislocation creep flow laws determined from experiments

The flow law expressed in Eq.1 implies that *n*, *m*, and *V* cannot be determined in isolation, because they are related by (Lu and Jiang, 2019):

86
$$m = -n \left(\frac{\partial \ln \sigma}{\partial \ln f_w} \right)_{T, \dot{c}} + \frac{V}{RT} \left(\frac{\partial P}{\partial \ln f_w} \right)_{T, \dot{c}}$$
(2)

87 In addition, to determine *n* through the relation $n = \left(\frac{\partial \ln \dot{\varepsilon}}{\partial \ln \sigma}\right)_{T,P,f}$ requires experiment runs under

constant T, P, and f_w conditions. As experiments (Dataset S1) were carried out under different 88 P-T conditions, one needs to normalize the data to a common reference P-T condition. The 89 following iterative approach, which is a refined one from that in Lu and Jiang (2019), is used to 90 determine all the parameters in Eq.1: First, starting with an initial input of n, the values of m and 91 V are solved using Eq.2, based on a minimum of three sets of P and f_w stepping experimental 92 data at constant $\dot{\varepsilon}$ and T (Lu and Jiang, 2019). Second, with the values of n, m and V, the 93 activation energy Q is determined from T stepping experimental data (see supplementary 94 information for details). Third, the values of m, V, and O allow for normalization of all data in 95 Dataset S1 to a reference *P*-*T* condition. With the normalized strain rates and normalized 96 stresses, an updated value of *n* is obtained by linear regression. A new round of iteration is 97

initiated with the updated *n*. The iteration continues until the output and input *n* values are within a specific tolerance of $(|n_{output} - n_{input}| \le 0.1)$. The pre-exponential term *A* is a final fitting parameter, which can be obtained once *n*, *m*, *V*, *Q* are determined. The final set of *n*, *m*, *V*, *Q*, and *A* are the best-fit flow law parameters for the data.

102 On the basis of Tokle et al. (2019, their Fig.1), the high-temperature and low-temperature data point to two different values of n. We thus consider the high-temperature and low-103 temperature data in Dataset S1 separately. For experimental data associated with samples 104 deformed at higher temperatures (900-1200°C) and lower stresses (Luan and Paterson, 1992; 105 Gleason and Tullis, 1995; Stipp and Tullis, 2003; Heilbronner and Tullis, 2006; Nachlas and 106 Hirth, 2015; Kidder et al., 2016; Richter et al., 2018), we used an initial value of n = 4, three sets 107 of P and f_w stepping experimental data from 820 MPa to 1590 MPa (Kronenberg and Tullis, 108 1984), and the T stepping experimental data from 827 °C to 1050 °C (Luan and Paterson, 1992; 109 Gleason and Tullis, 1995) in the above described iterative method. We obtained a final set of 110 $n = 3.9 \pm 0.2$, m = 2.6, V = 35.8 cm³/mol, $Q = 132 \pm 19$ kJ/mol, and 111 $A = 2.5 \times 10^{-(14\pm0.4)} \text{MPa}^{-n-m} \text{s}^{-1}$. For data associated with samples deformed at lower temperatures 112 113 (700-900 °C) and higher stresses (Kronenberg and Tullis, 1984; Koch et al., 1989; Post et al., 1996; Chernak et al., 2009; Tokle et al., 2013; Richter et al., 2016; Richter et al., 2018), we used 114 an initial value of n = 2, the three sets of P and f_w stepping experimental data from Kronenberg 115 and Tullis (1984) performed between 820 to 1590 MPa, and the T stepping experimental data 116 from 750 °C to 900 °C (Koch et al., 1989) in the iteration. We got a final set of $n = 2.5 \pm 0.1$, 117 m = 1.7, V = 23.1 cm³/mol, $Q = 126 \pm 16$ kJ/mol, and $A = 6.3 \times 10^{-(12 \pm 0.4)}$ MPa^{-n-m}s⁻¹. Therefore, 118 the best-fit flow laws are respectively: 119

120
$$\dot{\varepsilon} = 2.5 \times 10^{-14} f_w^{2.6} \exp\left(-\frac{132000 + 35.8P}{RT}\right) \sigma^4$$
 dominant prism slip \(3a\)

121
$$\dot{\varepsilon} = 6.3 \times 10^{-12} f_w^{1.7} \exp\left(-\frac{126000 + 23.1P}{RT}\right) \sigma^{2.5}$$
 dominant basal slip \(3b\)

122 with f_w , *P*, and σ all in MPa.

Eq.3a differs from the flow law obtained in Lu and Jiang (2019) slightly in *m* and *V* (m = 2.7, V = 35.3 cm³/mol in Lu and Jiang, 2019) and more in A ($A = 6.0 \times 10^{-15}$ in Lu and Jiang 2019) because Lu and Jiang (2019) applied the stress calibration of Holyoke and Kronenberg (2010) for the experimental data. Such calibration is not applied here for reasons given in supplementary information.

Using the two flow laws of Eqs.3, the strain rate versus differential stress plots for all data in Dataset S1 at a reference *P*-*T* condition of T = 900 °C and P = 1500 MPa are shown in Fig.1. Clearly, the two distinct flow laws are consistent with the current dataset: Eq.3a is the best-fit flow law for wet quartzite deforming predominantly by prism <a> slip and producing characteristic Y-max c-axis fabrics (Fig.2), whereas Eq.3b is the best-fit flow law for wet quartzite deforming predominantly by basal <a> slip with characteristic cluster of c-axes in the periphery (Fig.2).

3. Self-consistent determination of creep behavior of wet quartzite in the transitional regime

A steady-state dislocation creep is always associated with the activation of multiple slip systems (von Mises, 1928). Quartz c-axis fabrics from both experimental samples (Fig.2) and natural mylonites (e.g., Stipp et al., 2002; Law et al., 2010; Toy et al., 2010; Behr and Platt, 2011; Whitney et al., 2014) also suggest that common slip systems in quartz are basal <a>, prism
<a>, and rhomb <a>. Quartz c-axis fabrics suggest that rhomb <a> cannot be a dominant slip
system and prism <c> is not important unless at very high-temperature conditions (Lister and
Dornsiepen, 1982) and is rare in mylonites. In most natural quartz-bearing mylonites, one may
assume that at least one of basal <a> and prism <a> must be dominant to achieve a steady-state
dislocation creep. In such a case, we must consider the creep due to simultaneous operation of
both basal <a> and prism <a> systems that cannot be described by either Eq.3a or Eq.3b.

Where deformation mechanisms differ from one grain to another, the stress (and strain rate) field varies among grains accordingly. On a suitable Representative Volume Element (RVE) containing a large number of grains such that the average strain rate and average stress over the RVE represent the macroscale fields, the macroscale (or average) fields can be expressed as:

152
$$\overline{\dot{\varepsilon}} = \sum r_i \dot{\varepsilon}_i, \ \overline{\sigma} = \sum r_i \sigma_i$$
 (4)

where $\dot{\varepsilon}_i$, σ_i , and r_i are the strain rate, stress, and volume fraction of the *i*th grain, respectively; $\overline{\varepsilon}$ and $\overline{\sigma}$ are, respectively, the macroscale (or average) strain rate and stress on the RVE. The macroscale creep behavior of the sample is then a relation of the form $\overline{\varepsilon} = f(\overline{\sigma}, \lambda)$ where λ is a parameter describing the current *fabric state* of the aggregates. In the simplistic Voigt situation where all grains are assumed to have a uniform stain rate ($\dot{\varepsilon}_1 = \dot{\varepsilon}_2 = \overline{\varepsilon}$, Voigt, 1887), the grains with prism <a> slip as the dominant deformation mechanism will have a uniform stress σ_1 after Eq.3a, and those grains with dominant basal <a> slip will have another uniform stress σ_2 after 160 Eq.3b. Denoting the volume fractions for two grain types by α and $1-\alpha$, the macroscale 161 (average) stress on the RVE is:

162
$$\overline{\sigma} = \alpha A_1^{-\frac{1}{n_1}} f_w^{-\frac{m_1}{n_1}} \exp\left(\frac{Q_1 + PV_1}{n_1 RT}\right) \overline{\dot{\varepsilon}}^{-\frac{1}{n_1}} + (1 - \alpha) A_2^{-\frac{1}{n_2}} f_w^{-\frac{m_2}{n_2}} \exp\left(\frac{Q_2 + PV_2}{n_2 RT}\right) \overline{\dot{\varepsilon}}^{-\frac{1}{n_2}}$$
(5)

where subscripts "1" and "2" stand for the flow laws for dominant prism <a> slip and dominant
basal <a> slip. Similarly, if the stresses in all grains are assumed uniform – the Reuss situation
(Reuss, 1929), the macroscale (average) strain rate on the RVE is expressed as:

166
$$\overline{\dot{\varepsilon}} = \alpha A_1 f_w^{m_1} \exp\left(-\frac{Q_1 + PV_1}{RT}\right) \overline{\sigma}^{n_1} + (1 - \alpha) A_2 f_w^{m_2} \exp\left(-\frac{Q_2 + PV_2}{RT}\right) \overline{\sigma}^{n_2}$$
(6)

Fig.3 shows the deformation mechanism maps based on the Voigt and Reuss situations. One can see that in the Reuss case, the two strain rates ($\dot{\varepsilon}_1$ and $\dot{\varepsilon}_2$) are on a comparable order. Similarly, in the Voigt case, the two stresses (σ_1 and σ_2) are also comparable in magnitude. Therefore, in the *P*-*T* range where natural mylonites form, the simultaneous operation of both slips must be considered for the macroscale strain rate and stress relationship.

The Voigt and Reuss averages only provide the upper and lower bounds for the 172 macroscale flow behavior where both basal $\langle a \rangle$ and prism $\langle a \rangle$ are significant because the 173 assumption of either uniform stress or uniform strain in all grains is unrealistic. Even in a single-174 phase polycrystal aggregate like quartzite, $\dot{\varepsilon}_i$ and σ_i are distinct in each grain due to its unique 175 shape, orientation, and rheology (here determined by the dominant slip system) (Mura, 1987; 176 Nemat-Nasser & Hori, 1999). A self-consistent micromechanics approach based on the Eshelby 177 formalism (Jiang, 2014, 2016, Qu et al., 2016) is utilized to quantitatively describe such 178 interactions between a grain and the polycrystal aggregate. Briefly, the $\dot{\varepsilon}_i$ and σ_i in any 179

individual grain is related to the macroscale fields and macroscale rheology by a set of 180 *partitioning equations*. The macroscale rheology is expressed in terms of the constituent grains 181 182 rheology by a set of *homogenization equations*. The partitioning and homogenization equations are then solved simultaneously so that a self-consistent solution for the macroscale rheology 183 (here the macroscale flow law) is obtained. A more detailed description of the specific approach 184 185 is in supplementary information. As an example, we consider an RVE of 500 quartz grains subjected to a macroscale strain rate of 10^{-12} s⁻¹. A geothermal gradient of 20° C/km is considered 186 so that P is not an independent variable. The water partial pressure is assumed to equal to the 187 lithostatic pressure, and the water fugacity is calculated based on Pitzer and Sterner (1994). We 188 assume that the quartz grains are rheologically isotropic (following either of the two flow laws of 189 190 Eq.3), randomly mixed, and have ellipsoidal shapes with a maximum axial ratio of 2. Although our homogenization method allows tracking of grain shapes and orientations and hence 191 macroscale rheological anisotropy due to preferred shape fabric development, we limit the 192 calculation here to the initial state of the quartzite where the grains are randomly oriented in 193 space, and the macroscale rheology is isotropic. This is because quartz flow laws (Eq.3) do not 194 consider any anisotropy. 195

Fig. 4 shows the numerically obtained stress-strain rate relation as a plot of $\overline{\sigma}$ verses α and *T* in 3D. The strength of wet quartzite varies significantly with temperature as well as the relative dominant slip systems (α). At lower temperatures like $T = 300^{\circ}$ C, the variation in strength due to α can be nearly an order of magnitude, from more than 2GPa at $\alpha = 0$ to ~300MPa at $\alpha = 1$. As higher temperatures like $T = 600^{\circ}$ C, the quartzite strength is significantly reduced, and the effect of dominant slip systems is subdued as well. At $T = 600^{\circ}$ C, the strength changes from ~10MPa at $\alpha = 0$ to ~5MPa at $\alpha = 1$.

203 4. Discussion

We have obtained two endmember dislocation creep flow laws (Eq.3) from well-204 controlled creep experiments (Dataset S1) using a flow law expression (Eq.1) that includes the 205 pressure dependence of the activation enthalpy. Our approach differs from that of Tokle et al. 206 (2019). They obtained two "laboratory fit" flow laws by fitting experimental data into the 207 expression of $\dot{\varepsilon} = A f_w^m \exp\left(-\frac{Q}{RT}\right) \sigma^n$ that does not include the pressure effect on activation 208 209 enthalpy. However, the effect of pressure through the activation enthalpy term and that through the f_w term are in opposite directions (Lu and Jiang, 2019). As is clear from Eq.1, a greater 210 pressure results in a higher f_w^m that decreases σ but a reduced $\exp\left(-\frac{H}{RT}\right)$ that increases σ . 211 Therefore, considering the pressure effect on f_w alone cannot adequately explain the great 212 discrepancies in flow law parameters determined from creep experiments conducted under 213 214 different confining pressures. Tokle et al. (2019) also used estimates of stress, strain rate, and temperature from natural samples to obtain two "extrapolated fit" flow laws for wet quartzite. 215 We have shown that such estimates are associated with much greater uncertainties than well-216 217 controlled creep experiments and cannot allow an accurate determination of flow law parameters 218 (Lu and Jiang, 2019).

Our numerical calculation for the creep behavior of quartzite with simultaneous basal A and prism <a> slip is based on micromechanics and is consistent with the common observation that quartz c-axis fabrics vary continuously from Y-max to periphery cluster patterns in both experiments and nature. Tokle et al. (2019) attributed those creep data that plot between endmembers qualitatively to the contribution of the grain boundary sliding mechanism. The

grain boundary sliding mechanism may be more significant in some experiments considered by
Tokle et al. (2019) as they included ultrafine-grained samples (Rutter and Brodie, 2004; Fukuda
et al., 2018; Richter et al., 2018) which we did not. However, the grain boundary sliding
mechanism alone is inconsistent with combined Y-max and periphery cluster c-axis fabrics from
experiments and mylonite samples.

The values of Q for the two endmember flow laws in Eq.3 are essentially the same within 229 error ($Q = 132 \pm 19$ kJ/mol in Eq.3a and $Q = 126 \pm 16$ kJ/mol in Eq.3b). This is expected and 230 consistent with the understanding that dislocation creep, regardless of slip systems, is ultimately 231 accommodated by the same lattice diffusion of vacancy defects (Dorn 1954; Sherby and Burke 232 1968; Freer, 1981). The V values are reasonable both from a theoretical consideration (Lu and 233 Jiang, 2019) and in comparison with the activation volumes of other silicates 234 $(V = 14 \sim 24 \text{ cm}^3/\text{mol} \text{ for olivine, Karato and Jung, 2003, and } V = 24 \sim 38 \text{ cm}^3/\text{mol} \text{ for}$ 235 anorthite, Rybacki et al., 2006). The difference in V between two flow laws is poorly understood 236 but may suggest that point defect migration associated with the two slip systems is distinct 237 because the total activation volume is the sum of activation volume for point defect formation 238 239 and that for point defect migration (Béjina et al., 2003). The values and variation of m in Eq.3 are also poorly understood. The original introduction of the f_w^m term into the flow law (Eq.1) by 240 Kohlstedt et al. (1995) was phenomenological. There is still a great uncertainty regarding the 241 speciation of water-related defects and exactly how they facilitate the dislocation creep 242 243 (Paterson, 1989; Kronenberg, 1994; Tullis, 2002). Hobbs (1981,1985a) proposed that the 244 charged water-related defects, acting as dopants, influence the concentrations of charged defects, such as the dislocation kinks and jogs or vacancies and interstitials, and hence the deformation 245 246 rate. The value of m depends on the specific rate-controlling process and the nature of charged

247 defects, with *m* being 1.5 or 1.7 if the deformation is glide-controlled, 1.5 or 2.5 if the

deformation is climb-controlled, and 0.33 or 0.4 if the rate-controlling process is the diffusion of oxygen or silicon defect (Ord and Hobbs, 1986). Our values of m = 2.6 for dominant prism <a> slip and m = 1.7 for dominant prism <a> slip appear to be consistent that the dislocation creep in regimes 2 and 3 (Hirth and Tullis, 1992) is essentially climb-controlled.

It remains difficult to test quartz flow laws with natural data because the latter is limited 252 and associated with great uncertainties related to existing quartz grainsize piezometric relations 253 254 (Austin and Evans, 2007, 2009; Cross et al., 2017; Heilbronner and Kilian, 2017; Shimizu, 2008, 255 2012; Stipp and Tullis, 2003). We have calculated the paleostresses using the grainsize data from WMCC samples (Behr and Platt, 2011) and the piezometric relations of Heilbronner and Kilian 256 257 (2017) and Shimizu (2008, 2012) which seem to be in good agreement with the flow law of Lu and Jiang (2019). Estimates of α and T are required to plot the natural data in Fig.4. The 258 259 deformation temperature of WMCC samples was taken from Behr and Platt (2011). How to quantitatively estimate α for natural samples is unclear. We used the following qualitative 260 estimates here. Samples having pure Y-max fabrics are assigned to $\alpha = 1$ and those with pure 261 262 periphery cluster fabrics $\alpha = 0$. Samples with a mixture of Y-max and periphery c-axis patterns are given an α value between 0 and 1 depending on the relative strength of the two orientations. 263 264 For samples with no c-axis fabrics, we assume they deform by either dominant prism $\langle a \rangle$ slip $(\alpha = 1)$ or basal <a> slip ($\alpha = 0$) according to the sample descriptions. The estimations of $\overline{\sigma}$, α 265 and T of WMCC samples, with their error bars, are plotted in the 3D strength profile (Fig.4). 266 Because of the large uncertainties in $\overline{\sigma}$, α , T, and the choice of the macroscale strain rate and 267 geothermal gradient, it is not fruitful to quantitatively compare natural estimates and the 268 269 predictions of our flow laws. Our purpose of plotting natural estimates in Fig.4 is to present our

flow laws in the context of current limited knowledge and uncertainties of continental strength inorder to facilitate further investigation on the subject.

In fact, α is expected to vary with time in nature as a result of temperature change and/or 272 finite strain buildup. Hobbs (1985b) suggests that the critical resolved shear stress (CRSS) for 273 basal <a> slip is insensitive to temperature whereas that for prism <a> slip decreases with 274 275 increasing temperature. This is supported by the observation in both experiments and nature that basal <a> slip is dominant at lower temperatures and prism <a> slip at higher temperatures. On 276 277 the other hand, fabric buildup as the finite strain increases may also be responsible for slip 278 system switch and hence α variation. Heilbronner and Tullis (2006) observed in experiments a 279 progressive fabric transition from the periphery to Y-max with increasing strain and degree of 280 recrystallization. Regardless of the cause of α variation in a continuous deformation, our work suggests that it affects the flow strength of wet quartzite significantly. Because of the likely 281 variation of α , one should not expect a simple analytical expression for "the flow law" of 282 quartzite. The flow strength of wet quartzite evolves on the curved surface, as a function of 283 $\overline{\sigma} = f(T, P, \alpha)$ as in Fig.4. Based on the fabrics of WMCC samples (Fig.4), α decreased as the 284 temperature dropped and led to a remarkable increase in the flow strength at shallower depths. 285

286 **5.** Conclusions

By considering both the pressure dependence of the activation enthalpy and slip system dependence of the stress exponent, we have obtained two best-fit endmember flow laws (Eqs.3), corresponding respectively to dominant basal <a> slip and dominant prism <a> slip systems, from a dataset of carefully selected creep experiments of wet quartzite interpreted to represent steady-state regimes 2 and 3 dislocation creep.

We developed a self-consistent micromechanics-based method to numerically determine the flow strength of wet quartzite creeping by a continuous combination of basal $\langle a \rangle$ and prism $\langle a \rangle$ slips that are common in nature and experiments. The relative contribution of basal $\langle a \rangle$ slip and prism $\langle a \rangle$ slip is parameterized by α with $\alpha = 0$ and $\alpha = 1$ representing pure basal $\langle a \rangle$ slip and pure prism $\langle a \rangle$ slip respectively. The continuous mixture of the two dominant slip systems is represented by $0 < \alpha < 1$.

Our approach has reconciled the large discrepancies among flow parameters derived from different experiments, and our results are broadly consistent with c-axis fabrics from quartz-rich mylonites and experiment samples as well as theoretical considerations. The flow strength of wet quartzite over the entire range of $0 \le \alpha \le 1$ cannot be expressed in a simple analytical form but is best represented in a 3D plot as a function of temperature, pressure, and strain rate. The flow strength of wet quartzite is particularly sensitive to the dominant slip systems (α value) at lower temperatures.

305

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313

314 FIGURE CAPTION

Figure 1. Plots of normalized strain rate versus normalized stress for quartz creep experiments 315 and the dislocation creep flow laws. The circles represent data collected from axial compression 316 experiments and the diamonds from general shear experiments. All data are normalized to a 317 reference condition of T = 900 °C, P = 1500 MPa. (a). All data are normalized using the flow 318 law parameters m = 2.6, V = 35.8 cm³/mol, and Q = 132 kJ/mol. The dashed line represents the 319 best-fit line for the high-temperature experimental data. (b) All data are normalized using 320 m = 1.7, V = 23.1 cm³/mol and Q = 126 kJ/mol. The dashed line represents the best-fit line for 321 the low-temperature experimental data. 322 323 Figure 2. Plots of stress and strain rate data of eight experimental runs and the corresponding

quartz c-axis fabrics. All data are normalized to T = 900 °C and P = 1500 MPa using Eq.3a (for data ≥ 900 °C) and Eq.3b (for data < 900 °C). Variation in quartz c-axis fabrics suggest combined basal <a> and prism <a> slips, from Y-max pattern indicating dominant prism <a>slip at 1000 °C and 900 °C, through a mixture of basal <a>, prism <a>, and rhomb <a> c-axis fabric pattern at 915 ~ 875 °C, to a strong cluster of c-axes in the periphery reflecting dominant basal <a> slip at 800 °C and 700 °C. Grain boundary sliding mechanism cannot explain the caxis variation.

Figure 3. Deformation mechanism maps for wet quartzite for (a) Voigt and (b) Reuss cases. (a) A plot of differential stress versus temperature, with strain-rate contour lines $(10^{-10} \text{ s}^{-1}, 10^{-12} \text{ s}^{-1}, 10^{-12} \text{ s}^{-1}, 10^{-13} \text{ s}^{-1})$ of prism <a> slip (solid lines) and basal <a> slip (dashed lines). (b) A plot of strain rate versus temperature, with stress contour lines (10 MPa, 100 MPa, and 1000 MPa) of prism <a> slip (solid lines) and basal <a> slip (dashed lines). The strain rates/differential stresses of prism <a> and basal <a> slips are equal on the red lines. In the shaded areas, the difference of
the strain rates/differential stress between prism <a> and basal <a> slips is less than one order of
magnitude.

Figure 4. Plot of $\overline{\sigma}$ of wet quartzite versus T and α . The cross-sections of the $\overline{\sigma} - T$ plot at

- 340 $\alpha = 0$ and $\alpha = 1$ are two endmember situations (Eqs.3). At a given T, $\overline{\sigma}$ decreases as α
- 341 increases. As *T* increases, both the absolute strengths of prism $\langle a \rangle$ and basal $\langle a \rangle$ slips and the
- 342 strength contrast decrease. Estimations of paleostress, T and α of WMCC samples (Behr and
- 343 Platt, 2011) with error bars are also plotted.

344 **References**

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Supporting Information for

Flow strength of wet quartzite in steady-state dislocation creep regimes

Lucy X. Lu, Dazhi Jiang^a

Department of Earth Sciences, Western University, London Ontario, Canada N6A 5B7

^aState Key Laboratory of Continental Dynamics, Department of Geology, Northwest University, Xi'an 710069, China

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Introduction

A detailed description of the selection criteria of previous creep experiments for our analysis and the dataset of 19 creep experiments on quartz samples are given in supplementary information. Additionally, more detailed descriptions of the determination of tow flow laws and the self-consistent micromechanics-based homogenization approach are available in this supplementary information.

Text S1.

A total of 19 creep experiments, deformed from 700 °C to 1200 °C and corresponding to steady-state regimes 2 and 3 dislocation creep, on quartz samples are analyzed in this paper. To minimize complications from the effect of water on quartz, we only include data from samples with 0.1-0.4 wt.% H₂O added, following Tokle et al. (2019). Dislocation creep is grainsize-independent, whereas other deformation mechanisms, like grain boundary sliding and diffusion, are sensitive to the grainsize. To diminish the contributions of other mechanisms, we exclude ultrafine-grained samples (Rutter and Brodie, 2004; Fukuda et al., 2018; Richter et al., 2018) and only use samples with grainsize between 20-200 µm. These creep experiments were performed in either uniaxial compression or general shear setup. For the general shear experiments, the shear stresses (τ) and the shear strain rates ($\dot{\gamma}$) were converted to von Mises equivalent stresses (

$$\sigma_{eqv} = \sqrt{3}\tau$$
) and equivalent strain rates ($\dot{\varepsilon}_{eqv} = \frac{\dot{\gamma}}{\sqrt{3}}$) so that the results could be directly

compared with those from uniaxial compression experiments where the differential stresses and axial strain rates were used. Tokle et al. (2019) have shown that converting the shear stresses and shear strain rates to von Mises equivalent stresses and equivalent strain rates causes no systematic difference between the uniaxial compression experiments and general shear experiments. The experiments we used in this paper were conducted in the Griggs-type apparatus with solid salt assemblies, molten salt, or gas confining media. The stress calibration of Holyoke and Kronenberg (2010) for Griggs apparatus with solid medium or molten-salt medium was not applied for the following three reasons: First, it is not clear yet if such stress calibration, which was determined under uniaxial compression experiments, is applicable to general shear experiments (Tokle et al., 2019). Second, Holyoke and Kronenberg (2010) pointed out that applying their calibration for molten-salt medium to the data of Gleason and Tullis (1995) only changed the fitting term A (form 1.1×10^{-4} MPa⁻ⁿs⁻¹ to 5.1×10^{-4} MPa⁻ⁿs⁻¹) but not the values of *n* or *Q* substantially. Third, some experimental studies suggested that this stress calibration might be overdone (Kidder et al., 2016; Richter et al., 2016), an observation consistent with our recent results (Lu and Jiang, 2019).

Data Set S1. A Dataset of 19 Creep Experiments on Quartz Samples.

	Sample	Sample ID	Stress (MPa)	Strain Rate (s ⁻¹)	Tempera ture (°C)	Confining Pressure (GPa)	Notes	Apparatus
Kronenberg & Tullis (1984)	Arkansas Novaculite 1-	NV-16	250	1.60E-06	800	1.59	0.4wt% water	Griggs-type solid medium apparatus
	- 60μm	NV-46	280	1.60E-06	800	1.22	added, stress	(NaCl, CaCO ₃)
		NV-36	330	1.60E-06	800	0.82	determine d at 20% strain	
Koch et al. (1989)	Simpson Orthoquartzite 0.21±0.01mm	K874	582	1.52E-06	800	0.95		Griggs-type solid medium apparatus
		K875a	445	1.52E-06	770	1.05		
		K875b	1166	1.75E-05	770	1.03		
		K875c	176	2.11E-07	770	1.04		
		K875d	356	1.99E-06	770	1.03		
		K903	338	1.81E-06	850	0.91		
		K904a	595	1.61E-05	850	1.03		
		K904b	155	2.05E-07	850	1.04		
		K905	230	1.83E-06	850	1.07		
		K908	500	2.20E-05	875	1.27		

K909	224	1.51E-06	900	1.16	
K910	251	2.19E-06	900	1.16	
K911	871	1.59E-04	900	1.14	
K912a	87	1.48E-07	900	1.2	
K912c	291	1.63E-06	850	1.21	
K912d	155	2.17E-07	850	1.2	
K919a	421	1.64E-06	750	0.94	
K919b	212	1.87E-07	750	0.91	
K922a	326	1.64E-06	800	1.05	
K922b	117	1.25E-07	800	1.07	
K922C	1138	1.86E-05	800	1.07	
K923	472	1.82E-05	900	1.18	
K924	927	1.90E-05	800	1.2	
K929a	200	1.73E-06	900	1.2	
K929b	56	2.46E-07	900	1.2	
K929c	505	1.90E-05	900	1.2	
K932a	800	1.44E-04	900	1.25	
K932b	1326	1.64E-04	800	1.25	
K933a	530	1.65E-06	750	1.15	
K933b	225	1.57E-07	750	1.15	
К933с	1315	1.64E-05	750	1.13	

		K934	180	1.80E-06	900	1.2	
		K936	485	1.69E-05	850	1.24	
		K938	416	1.74E-05	900	1.17	
		K939	680	2.32E-05	850	1.15	
Luan & Paterson (1992)	Silicic acid 20- 30µm	5417a	350	6.00E-05	1027	0.3	Griggs-type gas medium apparatus
		5417b	500	3.00E-04	1027	0.3	
		5424	300	5.80E-05	1027	0.3	
		5433a	360	7.50E-05	1027	0.3	
		5433b	290	3.50E-05	1027	0.3	
		5433c	544	3.50E-04	1027	0.3	
		5485a	380	5.10E-05	1027	0.3	
		5485b	182	1.00E-05	1027	0.3	
		5485c	350	1.00E-04	1027	0.3	
		5490a	423	5.30E-05	1027	0.3	
		5490b	130	5.00E-06	1027	0.3	
		5490c	180	1.00E-05	1027	0.3	
		5490d	220	2.00E-05	1027	0.3	
		5490e	260	4.00E-05	1027	0.3	
		5490f	300	8.00E-05	1027	0.3	
		5490g	380	2.20E-04	1027	0.3	

5490h	445	4.00E-04	1027	0.3	
5493	280	5.70E-05	1027	0.3	
5568a	160	1.00E-05	1027	0.3	
5568b	260	1.00E-04	1027	0.3	
5568c	205	1.20E-05	927	0.3	
5568d	320	1.10E-04	927	0.3	
5574a	149	1.00E-05	1027	0.3	
5574b	238	4.30E-05	1027	0.3	
5574c	320	1.20E-04	1027	0.3	
5575a	270	1.00E-05	927	0.3	
5575b	150	1.00E-05	1027	0.3	
5575c	450	1.00E-04	927	0.3	
5575d	295	1.00E-04	1027	0.3	
5577a	230	1.00E-05	1027	0.3	
5577b	420	1.00E-04	1027	0.3	
5577c	320	5.00E-05	1027	0.3	
5578a	168	1.00E-05	1027	0.3	
5578b	270	1.00E-04	1027	0.3	
5578c	348	2.00E-04	1027	0.3	
5579a	200	1.00E-05	1027	0.3	
5579b	420	1.00E-04	1027	0.3	

		5580a	320	1.00E-05	927	0.3	
		5580b	570	1.00E-04	927	0.3	
		5581a	400	1.00E-05	827	0.3	
		5581b	705	1.00E-04	827	0.3	
		5582a	185	1.00E-05	1027	0.3	
		5582b	280	1.00E-05	927	0.3	
		5582c	400	1.00E-05	827	0.3	
		5583a	270	5.00E-05	1027	0.3	
		5583b	380	5.00E-05	927	0.3	
		5583c	490	5.00E-05	827	0.3	
Hirth & Tullis (1992)	Black Hills Quartzite 100µm	CQ-82	100	1.00E-06	900	1.5	Griggs-type solid medium apparatus
	Heavitree Quartzite 210µm	W339	290	1.00E-06	800	1.5	Griggs-type molten salt apparatus
Tullis & Wenk (1994)	Heavitree Quartzite 180- 200µm	W-370	250	1.00E-06	800	1.2	Griggs-type solid medium apparatus
	Black Hills Quartzite 100- 120µm	W-502	125	1.00E-06	800	1.2	Griggs-type molten salt apparatus

Gleason & Tullis (1995)	Black Hills Quartzite 100µm	W611a	84	6.31E-06	1050	1.5	samples with no melt	Griggs-type molten salt apparatus
		W611b	100	6.31E-06	1000	1.45		
		W611c	132	7.94E-06	950	1.44		
		BA96a	43	1.58E-06	1050	1.56		
		BA96b	61	1.58E-06	1000	1.54		
		BA96c	68	1.58E-06	950	1.53		
		BA96d	89	1.99E-06	900	1.5		
		BA94a	76	1.58E-06	1000	1.56		
		BA94b	109	6.31E-06	1000	1.56		
		BA94c	141	2.00E-05	1000	1.59		
		BA94d	228	7.94E-05	1000	1.6		
		BA95a	74	1.58E-06	1100	1.68		
		BA95b	110	6.31E-06	1100	1.68		
		BA95c	142	2.00E-05	1100	1.7		
		BA95d	225	7.94E-05	1100	1.7		
Post et al. (1996)	Black Hills Quartzite 100µm	W614	290	1.00E-05	900	1.69		Tullis-modified Griggs Apparatus
		W625	310	1.00E-05	900	1.72		
		W728	330	1.00E-05	900	1.6		

Stipp & Tullis (2003)	Black Hills Quartzite 100µm	W1030	207	2.00E-06	800	1.5	
		W1085	198	2.25E-06	850	1.5	
		W1049	268	2.30E-05	900	1.5	
		W1099	149	2.35E-06	900	1.5	
		W1051	189	2.40E-05	1000	1.5	
		W1024	102	2.05E-06	1000	1.5	
		W1025	87	2.10E-06	1050	1.5	
		W1119	257	2.10E-04	1100	1.5	
		W1029	130	2.45E-05	1100	1.5	
		W1022	130	6.00E-06	1100	1.5	
		W1066	60	2.25E-06	1100	1.5	
		W1126	34	2.20E-07	1100	1.5	
Stipp et al. (2006)	Black Hills Quartzite 100µm	W1172	48	2.25E-06	1000	1.5	Griggs-type molten salt apparatus
		W1142	66	2.00E-06	950	1.5	
		W1081	139	2.30E-06	900	1.5	
		W1089	177	2.40E-06	850	1.5	
		W1082	168	2.40E-06	800	1.5	
		W1140	156	2.25E-06	750	1.5	

Chernak et al. (2009)	Black Hills Quartzite	W1341	260	1.39E-05	900	1.5	Griggs-type solid medium apparatus
Holyoke & Kronenberg (2013)	Black Hills Quartzite 100µm	TMQ - 7	162	1.60E-06	800	1.6	Griggs-type solid medium apparatus
			422	1.60E-06	800	0.85	
			175	1.60E-06	800	1.56	
Kidder et al. (2016)	Black Hills Quartzite 70µm	W1341	242	2.17E-05	900	1.3	Griggs-type solid medium apparatus
		W1505	100	1.70E-06	900	1.3	
		W1509a	120	1.60E-06	900	1.3	
		W1510	87	2.27E-06	900	1.3	
		W1525a	94	1.69E-06	900	1.3	
		W1525b	192	4.50E-05	900	1.3	
		W1526a	337	1.68E-05	900	1.3	
		W1526b	124	2.30E-06	900	1.3	
		W1518b	58	2.20E-06	900	1.3	
Heilbronner & Tullis (2002)	Black Hills Quartzite 100µm	W872	310	1.50E-05	900	1.5	Griggs-type solid medium apparatus
		W858	180	1.50E-06	900	1.5	
		W946	364	1.73E-05	875	1.5	

		W920	165	8.66E-06	900	1.5		
		W935	182	1.73E-05	915	1.5		
Holyoke & Tullis (2006)	Black Hills Quartzite 100µm	W1105	788	2.00E-05	800	1.5	minimum strength after	Griggs-type solid medium apparatus
		W1106	139	2.00E-06	800	1.5	yield	
		W1153	346	2.00E-06	745	1.5		
Heilbronner & Tullis (2006)	Black Hills Quartzite 100µm	W920	121	7.22E-06	900	1.5		Griggs-type solid medium apparatus
		W1010	182	1.14E-05	915	1.55		
		W935	208	1.14E-05	915	1.5		
		W965	191	1.16E-05	915	1.55		
Nachlas & Hirth (2015)	Silica gel (20- 40µm)	W1674	242	2.89E-05	900	1		Griggs-type solid medium apparatus
		W1678	286	2.89E-05	900	1		
		W1680	251	2.89E-05	900	1		
		W1699	234	2.89E-05	900	1		
		W1700	225	2.89E-05	900	1		
		W1701	199	2.89E-05	900	1		
Tokle et al. (2013)		LT379	486	5.40E-06	800	1.5		

Richter et al. (2016)	Crushed quartz crystal <100µm	BR452	1329	1.56E-05	700	1.064		Griggs-type solid medium apparatus
Richter et al. (2018)	Crushed quartz crystal <100µm	BR383	940	1.62E-05	700	1.59	viscous regime	Griggs-type solid medium apparatus
		Br388	685	1.62E-05	800	1.53		
		Br419	880	1.62E-05	800	1.56		
		Br448	630	1.67E-05	800	1.07		
		Br412	312	1.62E-05	900	1.53		
		Br337	127	1.73E-05	1000	1.51		

Text S2.

Once the values of *n*, *m*, and *V* are determined, the activation energy *Q* can be obtained based on *T* stepping experimental data. To evaluate the value of *Q*, we first take the natural logarithm of the flow law $\dot{\varepsilon} = Af_w^m \exp\left(-\frac{Q+PV}{RT}\right)\sigma^n$ and rearrange it as $\ln(\dot{\varepsilon}) - m\ln(f_w) - n\ln(\sigma) + \frac{PV}{RT} = -\frac{Q}{RT} + \ln(A)$. For each run, we can measure the temperature (*T*), the confining pressure (*P*), the strain rate ($\dot{\varepsilon}$), and the differential stress (σ). The water fugacity f_w is determined using the state equation of water (Pitzer and Sterner, 1994), assuming the partial pressure of water is equal to the confining pressure. Plotting $-\ln(\dot{\varepsilon}) + m\ln(f_w) + n\ln(\sigma) - \frac{PV}{RT}$ versus $\frac{1}{RT}$, the value of *Q* is obtained by linear regression. We use the *T* stepping experimental data of Luan and Paterson (1992) and Gleason and Tullis (1995) to determine the value of *Q* for high-temperature data. The *T* stepping experimental data of Koch et al. (1989) is used to determine *Q* for low-temperature data. The results are shown in Figure.S1.



Figure S1. Plots of *T* stepping experimental data. (a). The plots of *T* stepping data of Luan and Paterson (1992; sample 5582 and sample 5583) and Gleason and Tullis (1995; sample W611 and sample BA96) using n = 4, m = 2.6, and V = 35.8 cm³/mol,

which yields $Q = 132 \pm 19$ kJ/mol by linear regression. (b) The plots of *T* stepping data of Koch et al. (1989) using n = 2.5, m = 1.7, and V = 23.1 cm³/mol, which yields $Q = 126 \pm 16$ kJ/mol by linear regression.

Text S3.

In micromechanics, the overall rheology of quartzite is defined as the average property over a Representative Volume Element (RVE, Fig.S2), which contains a large enough number of quartz grains. When the quartzite is subjected to a remote deformation, the partitioned strain rates and stresses vary from one quartz grain to another. To evaluate the strain rate and stress in each grain, we regard a quartz grain as an ellipsoidal inclusion embedded in the matrix composed of the rest of quartz grains. The matrix is rheologically heterogeneous, but in micromechanics, it is represented by a hypothetic Homogenous Equivalent Medium (HEM, Fig.S2) whose rheology is obtained from the rheological properties of all constituent grains in an RVE. Generalized Eshelby formalism (Jiang, 2014, 2016) relates the partitioned strain rates and stresses in individual grains to the remote fields:

$$\boldsymbol{\varepsilon}_{i} - \mathbf{E} = \left[\mathbf{J}^{d} - \mathbf{S}^{-1} \right]^{-1} : \mathbf{C}^{-1} : \left(\boldsymbol{\sigma}_{i} - \boldsymbol{\Sigma} \right)$$
(S1)

In Eq.S1, the sign ":" stands for the double-index contraction of two tensors. ε_i and σ_i are the partitioned strain rate and deviatoric stress tensors in the *i*th grain. The uppercase symbols **E** and **\Sigma** are the corresponding quantities in the remote field. **C** is the 4th order viscous stiffness (viscosity) of the matrix (HEM). **S** is the 4th order symmetric Eshelby tensor for each grain, and it is related to the inclusion shape and **C**. The partitioned strain rates and stresses, in turn, are used to evaluate the overall strain rate ($\overline{\varepsilon}$) and the overall stress ($\overline{\sigma}$) of quartzite: $\overline{\varepsilon} = \sum r_i \dot{\varepsilon}_i$ and $\overline{\sigma} = \sum r_i \sigma_i$, where r_i is the volume fraction of *i*th grain. Numerical calculations for the self-consistent approach are realized using MATLAB scripts, the algorithms for which are in the literature (Jiang, 2014, 2016; Qu et al., 2016). The executable MATLAB files are available from authors.

As an example, we have considered a quartzite made of 500 ellipsoidal quartz grains subjected to a macroscale strain rate of 10^{-12} s⁻¹. The quartz grains have ellipsoidal

shapes with a maximum axial ratio of 2. We assume that all quartz grains are rheologically isotropic and randomly distributed; therefore, the whole quartzite is also isotropic. An individual grain follows either of the two flow laws of Eq.3. As the self-consistent homogenization method involves tensor calculations, the tensor-form flow law of quartzite is required. In order to generalize the tensor-form equation, one needs to

rewrite the flow laws of the form
$$\dot{\varepsilon} = A f_w^m \exp\left(-\frac{Q+PV}{RT}\right) \sigma^n$$
 in terms of second

invariants of the deviatoric stress (σ_E) and strain rate ($\dot{\varepsilon}_E$) as

$$\dot{\varepsilon}_E = A_i f_w^m \exp\left(-\frac{Q+PV}{RT}\right) \sigma_E^n, \text{ where } A_i = \frac{3^{(n+1)/2}}{2} A \text{ (Ranalli, 1987, p.70). The tensor-}$$

form constitutive equation of isotropic materials is expressed as (Ranalli, 1987, p.70):

$$\sigma_{ij} = 2\eta_{eff} \varepsilon_{ij} \tag{S2}$$

where σ_{ij} and ε_{ij} are the components of the deviatoric stress and strain rate tensors; η_{eff} is the effective viscosity defined at a given state of stress or strain rate:

$$\eta_{eff} = \frac{1}{2} A_i^{-1} f_w^{-m} \exp\left(\frac{Q + PV}{RT}\right) \sigma_E^{1-n} \text{ or } \eta_{eff} = \frac{1}{2} A_i^{-\frac{1}{n}} f_w^{-\frac{m}{n}} \exp\left(\frac{Q + PV}{nRT}\right) \dot{\varepsilon}_E^{\left(\frac{1}{n}-1\right)}$$
(S3)

The effective viscosities of all quartz grains at a given state of stress or strain rate can be obtained using the parameters in two dislocation creep flow laws (Eq.3). Then numerical simulations allow us to evaluate the overall stress of quartzite with varying contributions of dominant slip systems at a given deformation condition.



Figure S2. A conceptual diagram of the self-consistent homogenization approach. A quartzite made of N constituent quartz grains, denoted by i = 1, 2, ..., N, subjected to a remote deformation (a). The overall rheological property of the quartzite is defined as the average property over a Representative Volume Element (RVE) (b), which contains a large number of quartz grains. Each quartz grain is regarded as an ellipsoidal inclusion embedded in the Homogenous Effective Medium (HEM) (c), whose rheological properties are obtained from the overall rheological properties of the RVE. To get the overall rheology from the constituents' properties is called homogenization. ε_i and σ_i are the partitioned strain rate and deviatoric stress in a constituent grain. The uppercase symbols E and Σ are the corresponding quantities in remote field. $\overline{\dot{\varepsilon}}$ and $\overline{\sigma}$ are the overall strain rate and stress of the quartzite.