Rheology of two-phase systems: A microphysical and observational approach

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1 **Abstract**

2 Ductile shear zones commonly contain distinctive bands of high strain rock characterized
3 by intimately mixed fine-grained two-phase or polyphase material. These ultramylonite
4 bands are weaker than the surrounding material, and may play a critical role in strain
5 localization. How such zones develop, how the phases become evenly dispersed, the bulk
6 rheology, and the controls on grain size, are all unclear. The following generic scenario
7 may resolve some of these questions.

8 1) Dislocation creep and dynamic recrystallization cause grain-size reduction: commonly,
9 the recrystallized grain sizes of the two phases differ.

10 2) Grain size reduction causes a switch to grain-boundary diffusion creep, which requires
11 grain-boundary sliding. Diffusion allows one phase to fill spaces that open between
12 grains of the other: this will happen most rapidly in the finer-grained phase. The grain
13 size of the resulting mixture is therefore controlled by that of the finer-grained phase.
14 This leads to mixing and dispersion of the two phases, producing a fine-grained, evenly
15 dispersed two-phase aggregate.

16 4) The bulk rheology will be controlled by grain-boundary diffusion creep of the two
17 phases, with the grain-size controlled by the finer-grained phase.

18 Bulk flow laws can be developed for quartz-feldspar and olivine-orthopyroxene
19 ultramylonites based on these concepts, using appropriate mixing laws.

20

21 **1. Introduction**

22 Much of our understanding of the rheology of rocks deformed in the ductile regime
23 comes from experimental data on single-phase materials, such as quartz, feldspar, calcite,

and olivine. It is now widely recognized that strain localization in crystalline materials is largely a result of microstructural changes consequent upon deformation, including grain-size reduction (e.g., Schmid, 1982; Rutter and Brodie, 1988; Braun et al., 1999), and the development of crystallographic preferred orientation (CPO) (e.g., Muto et al., 2011).

The dominant rock-types in the lithosphere are polyphase, however, and the bulk mechanical properties of these materials are likely to be different from those of the individual minerals. A common observation is that polyphase materials form ultramylonite bands within which the phases are finely and intimately mixed, and that these fine-grained mixtures are significantly weaker than the individual phases making up the undeformed rock (Figure 1). This phenomenon raises a number of questions, which need to be addressed if we are to understand how ductile shear zones in the lithosphere initiate and localize strain.

- How do these intimate mixtures of fine-grained phases form?
- What controls the grain size?
- How is the grain-size maintained during deformation, or how does it evolve?
- Can we define a bulk rheology for two-phase or polyphase mixtures?

The purpose of this paper is to address these questions based on microphysical considerations and observational data.

2. Previous work

The rheological properties of polyphase materials have attracted a lot of interest from observational, experimental, and theoretical standpoints over several decades. It is generally agreed that strain-localization in these materials results from grain-size reduction, produced by some combination of metamorphic reactions, fracture and

dynamic recrystallization, leading to a switch to grain-size sensitive creep (e.g., Stünitz and Fitz Gerald, 1993; Fliervoet et al., 1997; Kruse and Stünitz, 1999; Kenkmann and Dresen, 2002; Toy et al., 2010; Kilian et al., 2011; Linckens et al., 2011). Behrmann & Mainprice (1987), for example, documented observational evidence for simultaneous crystal plastic deformation in quartz, and superplastic deformation, which involves unrestricted grain-boundary sliding accommodated by some combination of diffusional and dislocation creep, in fine-grained quartz-feldspar aggregates. Diffusion creep in fine-grained two-phase mixtures has also been demonstrated in a number of experimental studies (e.g., Bruhn et al., 1999; McDonnell et al., 2000; Ji et al., 2001; Xiao, 2002; Farla et al., 2013; Tasaka et al., 2013).

Handy (1994), in a general review of the rheology of two-phase aggregates, made the case that the rheology depends on whether the aggregate contains a load-bearing framework of the stronger phase (LBF microstructure), or layers of interconnected weak phase (IWP microstructure). This suggestion was supported by an experimental investigation into the mechanical properties of aplite (one third quartz, two thirds feldspar) by Dell'Angelo and Tullis (1996), who showed that the material initially has a strength comparable to pure feldspar, but progressively weakens with strain due to interconnection of deformed quartz aggregates (see also Holyoke and Tullis, 2006). Similarly, Bons and Urai (1994), working with camphor-octochloropropane mixtures, showed that ~45% of the weak phase is needed for connectivity at the start of deformation, but only 25% after 50% shortening. Bloomfield and Covey-Crump (1993), on the other hand, found that strain is partitioned nearly equally between the phases in deformed calcite-halite mixtures.

70 Considerable theoretical attention has been paid to the distribution of stress and
71 strain-rate within a deforming two-phase material. This can range between the Reuss
72 model, in which the deviatoric stress is assumed to be constant throughout the aggregate
73 and the strain-rate varies according to the mechanical properties of the different phases;
74 and the Voigt model, in which the strain-rate is assumed to be constant and the stress
75 varies between the phases. Numerous intermediate models based on various theoretical
76 considerations have also been proposed (e.g., Hill, 1965; Yoon and Chen, 1990; Tullis et
77 al., 1991; Handy, 1994; Ji and Zhao, 1994; Takeda, 1998; Ji et al., 2003; Montési, 2007);
78 useful reviews and assessments of these models are provided by Ji et al. (2001); Dimanov
79 and Dresen (2005), and Takeda and Grier (2006). Numerical modeling by Takeda and
80 Grier (2006), for example, predicts that weak-phase supported materials approximate to
81 the Voigt model, whereas strong-phase supported materials show behavior intermediate
82 between the Reuss and Voigt models. This suggests that the mechanical properties are
83 likely to evolve with strain, as the weak phase deforms and creates an IWP
84 microstructure. None of these models allow for changes in rheology resulting from
85 decreases in grain size during deformation, however, which may result in the strength of
86 the two phase composite being less than that of either of the two phases taken
87 individually (e.g., Stünitz and Fitz Gerald, 1993; Fliervoet et al., 1997; Bruhn et al., 1999;
88 Farla et al., 2013).

89 In a very detailed experimental investigation of synthetic anorthite-diopside materials,
90 Dimanov and Dresen (2005) compared their results with a range of models, and they also
91 concluded that the rheology is likely to evolve with deformation. For materials in which
92 both phases deform by dislocation creep, their data broadly support the volume averaging

technique of Tullis et al. (1991), who predicted a composite flow law with parameters that are volume averages of those for the two phases. This method gives results that appear fit the data over the full range of volume fractions. For systems that can be approximated by rigid particles in a weak matrix, the model of Yoon and Chen (1990) works well for volume fractions of the strong phase in the range 0-70%. The self-consistent model of Hill (1965) is useful for mixtures of two linear viscous phases, although it overpredicts the strength for volume fractions around 50% (Dimanov and Dresen, 2005). Most of the other models give very poor fits to the experimental data.

Much attention has also been paid to the question of grain-size evolution in polyphase systems. This arises from the perceptions firstly, that surface-energy-driven grain growth is likely to counteract grain-size reduction in deforming rocks, and hence limit the potential weakening effect associated with a switch to grain-size sensitive creep (e.g., De Bresser et al., 2001), and secondly that interspersed phases are likely to inhibit grain growth. Much of the discussion has been couched in terms of the concept of Zener pinning, in which a dispersed rigid particles of a secondary phase inhibit surface-energy-driven grain-boundary migration in the primary phase (e.g., Olgaard, 1990; Evans et al., 2001; Herwegh and Berger, 2004; Herwegh et al., 2005; Herwegh et al., 2011; Bercovici and Ricard, 2012). In aggregates such as quartz + feldspar, olivine + orthopyroxene, or calcite + dolomite, however, the grain-boundary diffusivities of the two phases are fairly similar, and it may be better to describe the grain-size evolution in terms of Ostwald ripening (e.g., Voorhees, 1992), or by surface-energy-driven grain-boundary migration in both phases controlled by their relative diffusivities (Evans et al., 2001; Hiraga et al., 2010).

Another important issue that has received some attention is the role of chemical reactions during the deformation of polyphase mixtures (e.g., Toy et al., 2010). Setting aside the possibility of net transfer or exchange reactions related to changes in extensive variables (P , T , water fugacity), differential diffusion of different chemical components may result in phase changes at the local scale. Behrmann and Mainprice (1987), for example, document the formation of tails of K-feldspar around plagioclase porphyroclasts in high- T mylonites from the San Gabriel Mtns in California: this is likely to result from exchange of Na^+ and K^+ between the two feldspars. Differential diffusion of Si and Mg in deformed peridotites may result in migration of olivine and orthopyroxene boundaries from one phase into the other (Sundberg and Cooper, 2008; Hiraga et al., 2010).

3. Grain-size evolution: basic principles

Materials deforming by the motion of lattice dislocations tend to undergo dynamic recrystallization by a variety of mechanisms (e.g., Hirth and Tullis, 1992), which commonly leads to grain-size reduction. For a wide variety of both minerals and metals empirical relationships have been determined showing a relationship between the mean size of the dynamically recrystallized grains d_r and the differential stress σ of the form:

$$d_r = K_r \sigma^p, \quad 1$$

(e.g., Twiss, 1977). K_r is a material constant, and experimental evidence suggests little or no dependence on temperature or other environmental variables (e.g., van der Wal et al., 1993; Stipp and Tullis, 2003; Stipp et al., 2006). The exponent p varies according to the material, but has values of around -1. I refer here to d_r as the piezometric grain size. The empirical relationship between d_r and stress in several minerals has been calibrated

experimentally, but only over a limited range of conditions, and the theoretical basis for the relationship is still uncertain, although it is likely to be related to the relationship between subgrain size and stress (e.g., Platt and Behr, 2011a).

Under static conditions the grain size in crystalline materials tends to increase. Two processes contribute to this, both of them driven by the tendency of the system to minimize surface energy. One process is the migration of grain boundaries in a single phase, driven by surface energy (e.g., Evans et al., 2001). This process, referred to as γ GBM by Platt and Behr (2011a), results in the straightening of grain boundaries, the development of equilibrium interfacial angles where three or more crystals meet, and the shrinkage of small grains and growth of large grains leading to an overall increase in the average grain size. Grain-boundary migration requires diffusion across the boundary zone, a distance commonly of the order of a few nanometers. Migration may be inhibited by the concentration of impurity ions near grain-boundaries (Skemer and Karato, 2007), or by the presence of secondary phases along them. The second process is Ostwald ripening, involving diffusion between isolated grains of a secondary phase such that small grains shrink and large grains grow (Voorhees, 1992). Diffusion distances are much larger, and Ostwald ripening is therefore likely to be much slower than γ GBM.

The evolution of grain size due to γ GBM is commonly described in terms of a grain-growth law of the form:

$$d^n - d_0^{-n} = K_g t, \quad 2$$

(Evans et al., 2001), where d is grain size, d_0 is the grain size at time t_0 , and K_g includes temperature dependence following an Arrhenius function. The exponent n is generally

thought to be around 2 on theoretical grounds (Humphreys and Hatherly, 2004), but experimental data suggest values between 2 and 5.

In a polyphase system γ GBM in the dominant (matrix-forming) phase is likely to be inhibited by the presence of secondary phases along grain boundaries, a process known as Zener pinning. In general, Zener pinning is thought to lead to a limiting grain size of the matrix phase following an expression with a form such as:

$$d_m = K_z d_p / f_p, \quad 3$$

(Evans et al., 2001) where d_m is the limiting grain size in the matrix, K_z includes an Arrhenius type temperature dependence, d_p is the grain size of the pinning particles, and f_p is the volume fraction of the pinning phase. The ratio d_p/f_p is known as the Zener parameter. The efficacy of Zener pinning is complicated by whether or not the secondary phases move with the grain-boundaries or become included within the matrix grains, and by the degree to which the secondary phases grow by amalgamation or by Ostwald ripening. This has led to various modifications of the Zener equation, involving raising either the volume fraction or the Zener parameter to some empirically determined power (e.g., Herwegh et al., 2011). The uncertainties in the Zener equation and the parameter values limit its value as a predictive tool.

There is a widespread perception that the grain size in a deforming material is controlled by an equilibrium between grain-size reduction by dynamic recrystallization and grain growth (e.g., Derby, 1992; De Bresser et al., 1998). There are solid grounds to suggest that this is not true for the small grain sizes characteristic of many mylonites (e.g., <80 microns for quartz, and <100 microns for olivine). At the stresses corresponding to these grain sizes, grain-boundary migration is driven by the strain

energy associated with lattice dislocations (ρ GBM), rather than by surface energy. The two driving forces (ρ GBM and γ GBM) produce incompatible grain-boundary topologies, and act in opposite directions on curved grain-boundaries: during γ GBM the grain-boundaries migrate in the direction of their concavity, whereas during ρ GBM they migrate in the direction of convexity (Platt and Behr, 2011a). Whichever driving force is greater will therefore dominate the average microstructural evolution of the aggregate. This relationship predicts a boundary in stress – grain-size space for which the driving forces will be equal and opposite, referred to by Platt and Behr (2011a) as the D_{min} line, given by:

$$D_{min} \gg \frac{C_1 \mu \gamma}{S^2}. \quad 4$$

C_1 is a material constant, and is independent of temperature; μ is the elastic modulus; and γ is the interfacial energy.

At stresses above the D_{min} line, ρ GBM will be dominant; γ GBM will only affect the smallest grains in the aggregate, or it may not take place at all. ρ GBM is unlikely to cause grain growth in a recrystallized aggregate: the direction of grain-boundary migration is largely independent of grain size, and different boundaries of the same grain may migrate inwards or outwards with respect to the grain centre. Estimated locations of the D_{min} lines for quartz and olivine are shown on stress–grain-size maps presented later in this paper.

These essentially theoretical considerations are supported by observation: in materials that have undergone dynamic recrystallization by either grain-boundary bulging (BLG) or subgrain rotation (SGR), the average grain size of the recrystallized aggregate

is usually close to or identical to the newly formed grains (e.g., Platt and Behr, 2011a). Evidence for grain-growth is restricted to materials that have experienced some degree of post-deformational annealing. At high temperatures this relationship breaks down; rapid and irregular grain-boundary migration produces the microstructure described by Hirth and Tullis (1992) as Regime 3, and by Stipp et al. (2002b, a) as GBM. This may reflect deformation at stress/grain-size ratios below or along the D_{min} line.

There has been widespread discussion of the possibility that dynamic recrystallization may result in a sufficient reduction of grain size that the dominant deformation mechanism switches to some form of grain-size sensitive creep (GSSC). Possible GSSC mechanisms include classical diffusion creep (Coble or Nabarro-Herring creep), pressure solution creep, dislocation- or diffusion-assisted grain-boundary sliding (e.g., Langdon, 2006; Warren and Hirth, 2006; Hansen et al., 2011; Tasaka et al., 2014), or dislocation creep in which the dominant recovery mechanism is ρ GBM, which gives the flow law a dependence on d_r (DRX creep, Platt and Behr, 2011a). The concept of a deformation mechanism switch is supported by the fact that the piezometric line for olivine, for example, falls in a grain-size sensitive creep field over a significant part of the calibrated range. There is no reason why this should not happen: the switch in mechanism only occurs once a fine-grained dynamically recrystallized aggregate has formed, and the recrystallization process affects grains that may themselves be much larger, and hence lie in the dislocation creep field. On the other hand, De Bresser et al. (1998) and De Bresser et al. (2001), among many others, have suggested that once the switch occurs, grain-size reduction processes will cease, and grain growth will take over, possibly reversing the switch in mechanism. This question needs to be examined carefully. The argument

advanced by De Bresser et al. (2001) was based on the assumption that a switch in deformation mechanism would result in a stress drop. This might cause the material to drop below the D_{min} line in stress – grain-size space, in which case γ GBM could drive grain growth. In a natural shear zone, the effect of a switch in deformation mechanism is more likely to cause an increase in strain-rate, and depending on the external boundary conditions, the stress may not drop at all (Platt and Behr, 2011b). In this case, dislocation activity and dynamic recrystallization will continue at approximately the same level - the switch in mechanism is a result of an increase in the rate of GSSC, rather than a decrease in the rate of dislocation creep. Several GSSC mechanisms depend on dislocation activity in any case (e.g., disGBS and DRX creep).

In a deforming polycrystalline aggregate the situation is more complicated. Many studies have shown that the grain size in the multi-phase aggregates is substantially less than the dynamically recrystallized grain size for at least one of the phases where it occurs in single-phase layers (e.g., Behrmann and Mainprice, 1987; Stünitz and Fitz Gerald, 1993; Kruse and Stünitz, 1999; Kenkmann and Dresen, 2002; Kilian et al., 2011; Linckens et al., 2011). The aggregate may therefore be subject to a driving force for grain growth. Growth will be slowed or limited, however, by phase mixing, and also by grain-boundary sliding. Inhibition of grain growth by phase mixing has been widely cited as the reason fine-grained ultramylonites persist (Olgaard, 1990; Stünitz and Fitz Gerald, 1993; Krabbendam et al., 2003; Herwegh and Berger, 2004; Farla et al., 2013), and for this reason in the analysis below I neglect the effects of grain growth (I discuss the limits of this assumption and a possible test below). The question remains, however, how the fine-grained mixtures form in the first place.

4. Formation of fine-grained two-phase mixtures

The critical question raised by polyphase ultramylonites is the process by which fine-grained intimate mixtures of two or more phases can form. A variety of mechanisms can be envisaged, including crystallization of phases from initially fine-grained or glassy material, such as volcanic rocks or pseudotachylite; exsolution of a second phase (Linckens et al., 2011) or the crystallization of new phases during retrogressive metamorphism (White and Knipe, 1978; Brodie and Rutter, 1987); and the formation of symplectites around initially coarse-grained phases. The grain size produced by these processes is not readily predictable, and may subsequently evolve as the result of either dynamic recrystallization or surface-energy-driven grain growth.

A common process in initially coarse-grained two-phase systems that does not necessarily involve phase changes is the formation of dynamically recrystallized tails around porphyroclasts. A number of workers have suggested that the nucleation of new phases within porphyroclast tails can produce mixed-phase aggregates (e.g., Kruse and Stünitz, 1999; Kenkmann and Dresen, 2002). An example is illustrated in Figure 2, which shows a tail of dynamically recrystallized material forming around a large porphyroclast of sodic plagioclase in a mylonitic granitoid from the Aspromonte massif in southern Italy (Platt and Compagnoni, 1990). The tail contains a significant amount of dispersed quartz (Fig. 2a), and the resulting quartz-feldspar mixture is significantly finer-grained than in the ribbons of dynamically recrystallized quartz that formed from originally coarse-grained quartz in the protolith (Fig. 2b). It is common in mylonitic granitoids to find quartz infilling fractures in feldspar porphyroclasts: the relatively coarse grain size of the infilling quartz, and the lack of indications of crystal-plastic deformation, suggest that

it was emplaced by a solution-reprecipitation mechanism. An example of this is illustrated in Figure 3: quartz infills cracks at the margin of a K-feldspar porphyroclast in a mylonite from the Whipple Mtns metamorphic core complex, and the quartz is mixed with recrystallized feldspar in the tail (Figure 4a). EBSD images of the quartz-feldspar mixture in the tail show the intimate mixing of the two phases (Figs 4b, c). The effect of this mixing on the grain size and CPO are shown by an ultramylonitic granitoid from the Aspromonte massif in Calabria (Figures 5 and 6). Tails around plagioclase porphyroclasts (Figure 5) show fine-grained mixtures of quartz and dynamically recrystallized plagioclase; the relatively coarse grained quartz in adjacent quartz ribbons has a grain size of ~40-50 μm and a strong CPO, whereas both quartz and feldspar in the mixed tails have a grain size of 13-15 μm and weak and disorganized CPO (Figure 6).

In neither of the two examples discussed above is there any evidence of metamorphic reactions contributing to the process. The feldspars have deformed by a combination of brittle fracture and intracrystalline plasticity, leading to dynamic recrystallization, without any change in composition, and without breaking down to other phases. The admixture of quartz in the recrystallized tails appears to be the result of diffusion (probably by dissolution-precipitation) of silica from outside the porphyroclast system.

These observations suggest a relatively simple mechanism for the formation of fine-grained mixtures by deformation of initially coarse-grained two-phase aggregates, as set out in the following generic scenario (Figure 7).

1. Dislocation creep and dynamic recrystallization of both phases reduces the overall grain size. The stronger phase will tend to form porphyroclasts, and if it has a dynamically recrystallized grain size smaller than that of the matrix phase, the grain

size in the recrystallized tails will be less than that in the surrounding matrix. This relationship is common observed in quartz-feldspar and olivine-orthopyroxene rocks. Brittle fracture processes and metamorphic reactions may also play a role in grain size reduction, and in feldspars it can be difficult to distinguish grain refinement by frictional slip on cleavage planes from dynamic recrystallization (Dell'Angelo and Tullis, 1996).

2. If the grain size is sufficiently reduced by dynamic recrystallization, one or both phases may deform more rapidly by grain-boundary diffusion creep rather than dislocation creep (by grain-boundary diffusion creep I refer here both to classical Coble creep and to pressure solution creep).
3. Grain-boundary diffusion creep requires grain-boundary sliding, and tends to open up potential voids between the grains, which are filled by diffusion. This allows diffusion of the matrix phase into the dynamically recrystallized tails around the porphyroclasts. Note that this process implies the nucleation and growth of grains of the diffusing phase. This process was documented by Behrmann and Mainprice (1987), where K-feldspar fills pull-aparts in fine-grained plagioclase aggregates. The size of the new grains of the diffusing phase is controlled by the spaces that form between grains of the phase making up the tail, and hence can lead to even mixing and dispersion of the two phases, with a grain size controlled by that of the porphyroclastic phase.
4. The rheology of the resulting aggregate will therefore be controlled by a combination of grain-boundary diffusion creep and grain-boundary sliding, with a grain size controlled by dynamic recrystallization of the porphyroclastic phase.

The process suggested above is a postulate, and would be difficult to test experimentally, as it requires large strains to reach steady state microstructures and grain sizes. In the next section I follow the reasoning set out above to characterize the rheology of a mylonitic granitoid formed by this process, and I then suggest a possible observational test.

5. Flow law for a mylonitic granitoid

Both quartz and feldspar in a granitoid rock at temperatures above ~450°C will deform, initially at least, by dislocation creep. If dislocation climb is the main recovery mechanism, the flow law is likely to be power-law with a stress exponent in the range 3-5, and will not be sensitive to grain size. Tullis et al. (1991) proposed that a two-phase material in which both phases deform by dislocation creep will obey a bulk flow law in which the parameters are averaged by volume as follows (subscript a refers to the aggregate, subscripts 1 and 2 to the two phases, and ϕ is volume fraction):

The power law exponent n_a , the activation energy Q_a , and the prefactor A_a are the given by:

$$\log n_a = \phi_1 \log n_1 + \phi_2 \log n_2,$$

$$Q_a = [Q_2 (n_a - n_1) - Q_2 (n_a - n_2)] / (n_2 - n_1),$$

$$\log A_a = [\log A_2 (n_a - n_1) - \log A_1 (n_a - n_2)] / (n_2 - n_1). \quad 5$$

As discussed above, this averaging technique corresponds reasonably well to experimental data, particularly for high volume fractions of the stronger phase, which is the case for granitoid rocks. It tends to overestimate the strength for low volume fractions (<40%) of the stronger phase (Dimanov and Dresen, 2005). It has the disadvantage that it falls down if the power law exponents for the two phases are too close. For the systems discussed here, however, the two minerals show significantly different exponents. Flow

law parameters used in this paper for dislocation creep in quartz and feldspar are given in Table 1.

Experimental evidence suggests that solid-state grain-boundary diffusion creep (Coble creep) is very unlikely in quartz under geological conditions, even at fine grain sizes (Rutter and Brodie, 2004). An important alternative process of diffusion creep in a granitic mylonite at low or moderate temperatures is likely to be water-assisted diffusion of silica along grain-boundaries and fractures (den Brok, 1998; Farver and Yund, 2000). This process, commonly known as pressure-solution creep, has been described in terms of a number of flow laws, depending on the nature of the diffusion pathway, the source-to-sink distance of diffusion, and the presence or absence of advection (e.g., den Brok, 1998; Gratier et al., 1999). Strain-rate–grain size, stress–strain-rate, and stress–T plots for the three main flow laws (island-channel, stress corrosion, and thin-film) proposed for pressure solution are shown in Figure 8 (methods of calculation are given in the Appendix), together with the dislocation creep flow law of Hirth et al. (2001). In general, at temperatures and stresses characteristic of ductile deformation, strain-rates in quartz predicted by the island-channel and stress-corrosion models are ~5 orders or more of magnitude faster than dislocation creep, which rules them out as viable processes in a rock that has clearly deformed largely by crystal plasticity. Under these same conditions, grain-boundaries are likely to be tight, and advection limited, in which case the thin-film model for pressure solution is most likely to be applicable. The flow law for the thin-film model has a form analogous to that for Coble creep:

$$\dot{\epsilon} = \frac{C_2 V r_f c D_w S}{r_s R T d^3},$$

where C_2 is a shape constant (taken to be 44 for spherical grains), V is molar volume, ρ_f and ρ_s are the densities of fluid and solid phases; c is the solubility of the solid in the fluid phase as a mole fraction; D is the diffusivity of the solid in the grain-boundary fluid film; w is the grain-boundary width, σ is differential stress, R is the gas constant, T is temperature, and d is the grain size or source to sink distance (parameter values are given in the Appendix). Note that the strain-rate has a linear relationship to stress and a grain-size sensitivity exponent of 3. The grain size of the quartz is assumed to be the same as the feldspar, and hence governed by the piezometric relationship for sodic plagioclase (Post and Tullis, 1999), as discussed in the previous section. It is likely that the piezometric relationship varies with the composition of the plagioclase, and is different for K-feldspar, but experimental data are not available.

The experimental data of Rybacki et al. (2006) for wet fine-grained anorthite suggests that rates of diffusion creep in feldspar may be comparable to those produced by pressure-solution creep in quartz under the conditions discussed here (Figure 9). Granites contain sodic plagioclase and K-feldspar, rather than anorthite, but the detailed observational evidence provided by Stünitz and Fitz Gerald (1993) and Kilian et al. (2011) suggests that the rates of granular flow in quartz, albite, and K-feldspar are all very similar in fine-grained mixtures of these minerals. This suggests that a quartz-feldspar aggregate can be treated as a mixture of two linear viscous phases. I have therefore used the experimental data for diffusion creep of anorthite together with the flow-law for thin-film pressure solution creep in quartz, and applied the self-consistent mixing law:

$$3\mu_a^2 + [2(\mu_1 + \mu_2) - 5(\phi_1\mu_1 + \phi_2\mu_2)]\mu_a - 2\mu_1\mu_2 = 0,$$

where μ_a is the bulk viscosity of the aggregate, and μ_1 and μ_2 are the viscosities of the two phases (see discussion by Dimanov and Dresen, 2005). This mixing law may overpredict the strength for volume fractions around 50% (Dimanov and Dresen, 2005), but this is unlikely to be important if the viscosities of the two phases are similar.

We therefore use two different mixing laws to determine composite flow laws for dislocation creep and diffusion creep. These two mechanisms are independent, so it is reasonable to assume that the total aggregate strain-rate is the sum of the strain-rates produced by the two mechanisms (Montési and Hirth, 2003). For the purposes of this discussion I have not allowed for any additional component of strain-rate accommodated by grain-boundary sliding (GBS). GBS is in any event an essential component of grain-boundary diffusion creep (Langdon, 2006), and is not normally treated separately in flow laws for this mechanism. Unrestricted grain-boundary sliding, which is thought to be responsible to superplastic behavior, may contribute substantially to the strain-rate. The approach taken here is therefore conservative, in that it does not take this contribution to the strain-rate into account. Eqns 5, 6, and 7 lead to a bulk flow law as follows:

$$\dot{\epsilon} = A_a S^{n_a} \exp\left(\frac{-Q_a}{RT}\right) + \frac{S}{2m_a}. \quad 8$$

The bulk viscosity μ_a is calculated from Eqn 8 using viscosity terms for thin-film pressure solution creep in quartz and for diffusion creep in feldspar given by:

$$m_{ps} = \frac{r_s RT d^3}{2C_2 V r_f c D w} \quad \text{and} \quad m_{diff} = \frac{d^3}{2A f(H_2O) \exp(-Q/RT)}. \quad 9$$

Parameter values for these equations are listed in the Appendix (for pressure solution creep) and in Table 1, and the volume fraction of feldspar is taken to be 0.4. As discussed

above, the mixing laws I have used appear be approximately valid over the whole range of volume fractions.

A stress–grain-size deformation mechanism map for a granitoid rock obeying this composite flow law is shown in Figure 9a. In the right hand field labelled "dislocation creep", the grain size is large enough that the second term in Eqn 7 is negligible, and the rock deforms by dislocation creep in both quartz and feldspar. In the left-hand field labelled "diffusion creep", an ultramylonite formed by phase mixing as described in Section 4 deforms by thin-film diffusion creep in quartz, and by diffusion creep in feldspar, and the second term in Eqn 7 contributes the bulk of the strain rate. The map does not correctly predict the behavior of the material during the process of grain size reduction, when it may show a very large range of grain sizes, and the two styles of deformation co-exist in different parts of the rock. It does, however, provide a useful guide to the change in material properties consequent upon the conversion of the rock from a coarse-grained granitoid (point A in Figure 9c), to an ultramylonite (point B), assuming constant stress during deformation. In this case, there is >4 orders of magnitude increase in strain rate resulting from the change. A map for quartz under the same conditions is shown for comparison (Figure 9a) - in the dislocation creep field pure quartzite deforms several thousand times faster than granite, but under the conditions of point B, the two materials deform at similar rates.

These maps and the conclusions drawn from them are only as good as the flow laws and the piezometers on which they are based, and will need revision as more experimental data for quartz and feldspar become available. Also, the flow law for a granitoid ultramylonite derived here only applies to that part of the rock that consists of a

fine-grained two-phase mixture, formed by the mechanism proposed in this paper. Real mylonitic rocks, including the ones discussed and illustrated here, are commonly composed of layers with different compositions, formed by deformation and dynamic recrystallization from different protolith minerals. The bulk rheology of such composite materials depends on the relative volume proportions of the different layers, and their degree of connectivity. The rheology of the two-phase mixture is an essential component of the bulk rheology, and may dominate it, but it is not the whole story.

Another issue is the role of pressure solution in the process discussed here. Pressure solution is a very poorly understood process, and there is little reliable experimental data to constrain the flow laws. Three different flow laws have been proposed, based on different descriptions of the diffusion pathway, which predict strain rates differing by many orders of magnitude. It is possible that these different pathways are influenced or activated by temperature, deviatoric stress, and fluid pressure; at present we have no constraints on this. Pressure solution is critically dependent on the presence of water, either as a fluid phase or a grain-boundary film. Changes in fluid pressure driven by hydration or dehydration reactions may dramatically affect the rate of pressure solution creep. Partial melting, which in granitoid rocks may start at around 650°C, will severely lower water activity, as water partitions into the melt, and this may limit the activity of pressure solution creep at higher temperatures.

There are also limits on grain-size reduction processes in deformed rocks. At higher temperatures, deviatoric stress is lower, and the material may pass into the stress–grain-size field where surface-energy driven grain-boundary migration becomes dominant. The grain size will no longer be controlled by the piezometric relationship, and the switch to

grain-size sensitive creep may no longer occur. This point may correspond to the intersection of the D_{min} line (see above) with the piezometer - an estimated position for this in quartz (~60 μm grain size and 18 MPa shear stress) is shown in Figure 9a. In the crust, this is likely to occur at a temperature of ~450°C (Behr and Platt, 2011). The position of the D_{min} line for feldspar, unfortunately, is not known.

A possible test for the operation of the mechanism described here is to determine the flow stress from the dynamically recrystallized grain size of pure quartz bands or aggregates, and then see if the grain size of quartz feldspar mixtures in the same rock corresponds to that predicted by the feldspar piezometer. I have done this for a number of published descriptions of ultramylonitic granitoids, as well as the two examples described in this paper (Table 2). The results, though few in number, seem consistent with the hypothesis.

6. Olivine-orthopyroxene mixtures

The olivine-orthopyroxene system differs from the quartz-feldspar system in several ways. The two minerals share many deformational characteristics, and the contrast in dislocation creep strength between them is likely to be much less than between quartz and feldspar; the presence of an aqueous phase is unlikely under most conditions, though a melt phase may be present; and the similarity in composition of the two minerals may allow grain-boundary migration and grain-boundary diffusion creep in a mixed aggregate to proceed by the diffusion of a single component (e.g., MgO). Recent observational and experimental work on the behavior of the system requires some review.

Observational data

Several recent studies of mylonitic shear zones cutting peridotites and pyroxenites within the Semail ophiolite of Oman differ in detail in their interpretations (Toy et al., 2010; Linckens et al., 2011; Tasaka et al., 2014), but these workers all observed bands of polyphase ultramylonite that are finer grained than dynamically recrystallized olivine in the same rock. They attributed grain-size reduction to dynamic recrystallization, accompanied by compositional changes in the pyroxene, and they attributed the mixing of olivine and pyroxene to grain-boundary sliding accompanied by diffusion of one or both phases. They concluded that strain localization resulted from a switch to some form of grain-size sensitive creep in the fine-grained mixture.

Experimental work

McDonnell et al. (2000) and Ji et al. (2001) both worked on artificial fine-grained forsterite-enstatite mixtures. Ji et al. (2001) worked at low pressures in a gas-medium apparatus, which means that experimental conditions were in the stability field of protoenstatite rather than orthoenstatite; McDonnell et al. (2000) worked at 600 MPa confining pressure in a solid-medium apparatus, in the orthoenstatite stability field. In both sets of experiments a low stress exponent of the mixture suggested a component of diffusion creep, accompanied by grain-boundary sliding.

Tasaka and Hiraga (2013) carried out a very detailed experimental analysis of syndeformational grain-growth kinematics in extremely fine-grained forsterite-enstatite mixtures in a gas-medium apparatus, and discussed the results in terms of a Zener pinning model. This group also presented results from a long set of rheological experiments on similar materials (Tasaka et al., 2013), carried out at constant

displacement rate. Samples with very low enstatite contents show power law creep with stress exponent of 3 and no grain-size sensitivity, but at higher enstatite contents the stress exponent drops to around one and the grain-size sensitivity is 2. They interpret this as indicating grain-boundary sliding, with a change in the accommodation mechanism from dislocation creep to diffusion creep with increasing enstatite content. All their experiments were carried out at low pressures, and hence outside the stability field of orthoenstatite.

Farla et al. (2013) also carried out experiments on fine-grained olivine-orthopyroxene mixtures, using a solid-medium direct-shear apparatus at high pressure, so that the experiments lay in the orthoenstatite stability field. Their experiments suggest that the mixtures are significantly finer-grained and weaker than pure olivine deformed under the same conditions, due to a switch to grain-size sensitive creep. They attribute the mixing of the two-phases to neighbour-switching during grain-boundary sliding.

Linckens et al. (2014) carried out experiments on initially coarse-grained olivine-orthopyroxene mixtures, also using a solid-medium apparatus. They noted that the dynamically recrystallized orthopyroxene is 5-10 times finer grained than the olivine, and calibrated an orthopyroxene piezometer using the grain size of the olivine and the experimental piezometer for olivine of van der Wal et al. (1993). They saw very minor mixing of the two phases during the experiments, and concluded that the processes leading to a well-mixed microstructure are inefficient at the conditions achieved in their experiments.

Bruijn and Skemer (2014) use the results of experiments reported in an unpublished thesis (Lawlis, J.D., 1998, High temperature creep of synthetic olivine-enstatite aggregates, PhD thesis, Dept. of Geosciences, Pennsylvania State University, State College, PA.) to construct flow laws for orthopyroxene. The dislocation creep flow law is based directly on experiments designed to determine the flow-law parameters; they then use deviations from this law at fine grain sizes to determine two data points for a diffusion creep flow law, based on assumptions about the factors controlling the grain size. The experiments were carried out in a solid medium apparatus in the stability field of orthopyroxene, and hence are probably the most reliable results available, so I use the dislocation creep flow law in the following discussion (Table 1). For the diffusion creep flow law (Table 1), I chose values of the parameters given by Tasaka et al. (2013) within the uncertainties such that the flow law fits the two data points inferred by Bruijn and Skemer (2014).

Summary of conclusions reached to date

There is a general consensus that phase mixing in olivine-pyroxene aggregates leads to substantial weakening, and that this is likely to be a result of a switch to GSSC, probably involving an important component of grain-boundary sliding. The mechanism of mixing remains uncertain, and has been attributed to solution-redeposition involving an unspecified fluid phase, grain-boundary diffusion, or grain-boundary sliding with neighbour-switching. Controls on the grain size of the mixture remain uncertain, with most theoretical discussion focusing on grain growth governed by Zener pinning, rather than the processes of grain-size reduction that produce the fine-grained mixture. Flow

laws for the fine-grained mixtures remain debated: stress exponents in the range 1-2, and grain-size exponents in the range 2 to 3, have been suggested or assumed.

A postulate for the behavior and flow laws of olivine-orthopyroxene mixtures

Microstructural evidence suggests that the processes and effects involved in the formation of fine-grained olivine-orthopyroxene mixtures may be analogous to those suggested above for quartz-feldspar mixtures. This is illustrated in Figure 10, which shows an orthopyroxene porphyroclast system in a xenolith from the Jagersfontein kimberlite in South Africa. These xenoliths were studied in some detail by Skemer and Karato (2008), and Figure 10 is redrawn from an EBSD map of the porphyroclast and its tail published by Bruijn and Skemer (2014). The map shows dynamic recrystallization on the margin of the porphyroclast, probably largely by the SGR mechanism, with some contribution from ρ GBM. Grain size in the orthopyroxene is reduced by dynamic recrystallization from several mm to $\sim 50\mu\text{m}$, and further grain size reduction takes place in the tail with distance away from the porphyroclast. These values suggest a differential stress of 15-20 MPa, according to the paleopiezometer constructed by Bruijn and Skemer (2014). The dramatic increase in strain of the orthopyroxene in the tail relative to the porphyroclast suggests a switch to grain-size sensitive creep. The interesting aspect of the recrystallized tail from the perspective of this paper is the presence of small grains of olivine mixed in with the orthopyroxene. These grains are completely isolated within the orthopyroxene tail, but tend to decrease in abundance with distance away from the coarse-grained olivine on either side of the tail (Figure 10). This suggests a diffusional mechanism for mixing. Given the widespread development of olivine-orthopyroxene

mixtures in high-strain peridotites, I suggest that diffusional mixing, with grain size controlled by dynamic recrystallization in the orthopyroxene, is a common mechanism.

To assist the discussion, deformation mechanism maps for pure olivine and pure orthopyroxene are shown in Figure 11a-d, using the parameters listed in Table 1. These show that in both the dislocation creep and diffusion creep fields, olivine deforms about two orders of magnitude faster than orthopyroxene, with the difference in rate decreasing at high temperature and low stress. The stress-temperature maps (Figures 11b and d) are constructed assuming that the grain size follows the piezometric laws for the two minerals. These suggest that at strain-rates in the range 10^{-15} to 10^{-11} , dynamic recrystallization will cause olivine to switch to diffusion creep below a temperature in the range 800-900°C; whereas orthopyroxene will switch below 1000-1050°C. Above these temperature ranges the dominant deformation mechanism in each mineral is dislocation creep. The reason for this behavior is illustrated in the stress-grain-size maps (Figures 11a and c): these show that if the grain size follows the piezometric relation, there is a switch to diffusion creep at higher stresses and hence smaller grain sizes, which for any given strain-rate are required at lower temperatures.

At any given stress, below ~ 500 MPa, the piezometer for orthopyroxene predicts significantly finer grain sizes than the piezometer for olivine. If, as I postulate above for the two-phase mixture, the grain size in the olivine is controlled by the piezometric relationship for orthopyroxene, then the recrystallized olivine will lie in the diffusion creep field over a wide range of stress and temperature space. This therefore suggests a scenario based on the following assumptions, some of which still need to be tested.

1. In a coarse-grained peridotite, both olivine and orthopyroxene deform by dislocation creep, and the grain size will be reduced following the respective piezometers for the two minerals (Figure 11a and c).
 2. Over a wide range of conditions, orthopyroxene will switch into the diffusion creep field as a result of grain-size reduction (Figure 11c). Diffusion rates in the olivine are likely to be around two orders of magnitude faster than in the orthopyroxene, however, and hence olivine can diffuse into the spaces that open up around the orthopyroxene grains due to grain-boundary sliding.
 3. The grain size of the olivine where it is mixed with the orthopyroxene is approximately that of the dynamically recrystallized orthopyroxene, and this governs the rate of diffusion creep in the olivine.
 4. If the shear zone is operating at constant stress, dislocation creep and dynamic recrystallization will continue in both phases, even though diffusion creep is the dominant deformation mechanism. This, and the mixing of the two phases, will largely inhibit surface-energy driven grain growth.
- I have not introduced any other grain-size sensitive creep mechanisms into this discussion. The most widely discussed mechanism is Dis-GBS (e.g., Hansen et al., 2011), but the flow-law parameters for this mechanism are uncertain, and there is limited microstructural evidence for grain-boundary sliding in mylonitic peridotites. The other possibility is DRX creep (Platt and Behr, 2011a), but the flow law is very sensitive to the grain-boundary mobility, which for olivine is uncertain to within at least 3 orders of magnitude (Evans et al., 2001). Addition of these mechanisms may change the details of the processes discussed here, but are unlikely to change the overall conclusions.

At temperatures below 1000-1050°C, a peridotitic ultramylonite is likely to deform predominantly by diffusion creep in both olivine and orthopyroxene. The fine-grained aggregate can therefore be treated as a mixture of two linear viscous phases, and so I have used the self-consistent mixing law (Eqn 9).

Stress–grain-size and stress-temperature deformation mechanism maps for harzburgite, incorporating dislocation creep in both phases using the parameter-averaging method of Tullis et al. (1991), and diffusion creep for both phases after dynamic recrystallization, are shown in Figure 11 e and f. As noted for above for a granitoid, the stress–grain-size map does not correctly predict the behavior of the material during the process of grain-size reduction, but it shows how the material properties of the mixture will change during dynamic recrystallization to an ultramylonite. At constant stress, the change would result in an increase in strain rate by a factor of ~20-40 at 900°C. This result is weakly dependent on stress, but decreases significantly with temperature: at 700°C the strain-rate increases by three orders of magnitude; above ~1050°C, deformation remains in the dislocation creep field, and no weakening is predicted.

The shaded area in Figure 11f is a region of stress-T space in which the deformation mechanism map for orthopyroxene predicts that it will be in the dislocation creep field, but the olivine, if it has a grain size controlled by the piezometric relation for orthopyroxene, will be deforming by diffusion creep. In that case mixing of the two phases might be much slower, and the self-consistent mixing law for two linear viscous phases would not apply. The predicted deformation mechanism and strain-rates in this area are therefore likely to be unreliable.

633 **Discussion and Conclusions**

634 The analysis presented here suggests that the formation of fine-grained two-phase or
635 polyphase mixtures in ultramylonites may in part result from a combination of dynamic
636 recrystallization accompanying initial crystal-plastic deformation in both phases, which
637 leads to grain-size sensitive creep accompanied by grain-boundary sliding, which in turn
638 leads to diffusion of each phase into spaces opened up by grain-boundary sliding in the
639 other phase. Flow in the resulting mixture takes place by diffusion creep of both phases,
640 while the grain size is controlled by the piezometric relationship for whichever phase
641 recrystallizes to the finer grain size. This can lead to an increase in strain rate (at constant
642 stress) relative to the initial coarse-grained material by more than four orders of
643 magnitude, particularly at low temperatures. This is likely to be of major significance in
644 promoting strain localization in the ductile field.

645 The process described here is likely to be most applicable to materials such as
646 granitoid rocks, peridotites, and calcite-dolomite rocks, in which syn-kinematic
647 metamorphic reactions and phase changes do not play a significant role. In mafic and
648 pelitic rocks, for example, ductile deformation is commonly accompanied by
649 metamorphic reactions, and the rheology is likely to be strongly modified by diffusional
650 processes accompanying those reactions, by the strength (or weakness) and anisotropy of
651 the resulting new phases, and by fluctuations in fluid pressure and water activity
652 accompanying hydration and dehydration reactions. These complexities are not addressed
653 in this paper.

654 The composite flow laws presented here are subject to the uncertainties in the
655 experimental and theoretical data for the individual phases, and they depend on the

656 mixing laws that I have applied, which are approximate and are not universally valid. As
657 far as I can tell, for the two systems studied here, the mixing laws should be valid over
658 the full range of volume fractions. The Tullis et al. averaging technique, however, falls
659 down if the power law exponents for the two phases are too close.

660 There are several problems with the flow laws I have used. In the absence of
661 experimentally determined flow laws for sodic plagioclase and K-feldspar, I have used
662 experimental data for anorthite, and applied it to granitic rocks. The pressure solution
663 flow law I used has not been experimentally verified. I believe that the estimated strain-
664 rates are likely to be conservative, however, as several factors may result in higher
665 values. Firstly, I have neglected several contributors to grain-size sensitive creep, which
666 may result in higher strain rates. These include unlimited grain-boundary sliding with
667 neighbor swapping (superplasticity), and dislocation creep with grain-boundary migration
668 as a recovery mechanism (DRX creep). Second, I have chosen conservative values for
669 pressure solution creep. Thin-film pressure solution is the slowest of the postulated
670 pressure-solution mechanisms: much higher rates may occur if fluid pressure and content
671 are sufficient to produce fluid-filled channels. The diffusion creep flow law for olivine is
672 likely to be rate-controlled by diffusion of SiO_2 , which is relatively slow (Hiraga et al.,
673 2010). At least the initial stages of diffusional mixing of olivine and orthopyroxene,
674 however, may be controlled by the more rapid diffusion of MgO , which converts
675 orthopyroxene to olivine at the sink position, and olivine to orthopyroxene at the source
676 position.

677 Another limitation to the method is the assumptions that the grain size is controlled
678 by experimentally determined piezometers, and that there is no surface-energy-driven

grain growth. These assumptions are likely to break down at low stresses, equivalent to dynamically recrystallized grain sizes above $\sim 60 \mu\text{m}$ for quartz, and $100 \mu\text{m}$ for olivine, corresponding to the intersection of the piezometer with the D_{min} line for each mineral.

As pointed out earlier, the flow laws for two-phase ultramylonites derived here only apply to the relatively narrow bands of ultramylonite that are interleaved at all scales with layers with different compositions and deform by different mechanisms. Much work remains to be done to determine how the bulk rheology in deforming zones is controlled by this complexity in the microstructure.

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Appendix

Calculation of pressure solution flow laws

Pressure solution flow laws were calculated from expressions given by den Brok (1998) for the thin-film, island channel, and stress-corrosion models, using the following methods and parameters:

701 V : molar volume of solid $2.269\text{e-}5$ (Berman, 1988),
 702 ρ_f : density of fluid phase 0.923 ,
 703 ρ_s : density of solid phase 2.65 ,
 704 grain-boundary width $w = 2\text{E-}9$ m (Farver and Yund, 2000).

705 The solubility c of the solid in the fluid phase was calculated using expressions from
 706 Fournier and Potter (1982), using values for the specific volume of water interpolated
 707 from tabulated values from Burnham et al. (1969). The grain-boundary diffusivity D_{gb} of
 708 the solid in the grain-boundary fluid film was calculated from the bulk diffusivity,
 709 following Farver and Yund (2000), as follows:

710 $D_b = D_0 \exp(-Q_b/RT)$, and $D_{gb} = D_b * 1.2\text{e-}6 / (\tau * 2\text{e-}9)$, where $D_0 = 3.7\text{E-}10$ and
 711 $Q_b = 137000$; and the tortuosity factor $\tau = 1.75$.

712 The diffusivity of silica in water $= 2.8\text{e-}5 * \exp(-6300./T)$ (Watson and Wark, 1997).
 713 Length scales of the diffusion paths for the different flow laws are grain size $= 20 \mu\text{m}$ for
 714 the thin film model, the channel width for the island channel model $= 0.1 \mu\text{m}$, and the
 715 island diameter $= 0.5 \mu\text{m}$ for the stress-corrosion model.

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Figure and table captions

Figure 1. Peridotite mylonite from the Pyrenees. Porphyroclast of orthopyroxene (Opx)

indicates primary grain size; the weak matrix flowing around it is a mixture of

dynamically recrystallized olivine (Ol) and orthopyroxene. Sample courtesy of Reinoud

Vissers; image courtesy of Whitney Behr.

Figure 2. a) Recrystallized tail on margin of plagioclase porphyroclast in granitoid

mylonite (PC32, Aspromonte massif, Calabria, Italy). Plagioclase (p) is recrystallizing by

subgrain rotation; recrystallized grains become separated by quartz (q). b) Quartz ribbons

in the same rock (top) deform by dislocation creep, and have a mean recrystallized grain

size of 70 μm ; quartz-feldspar mixture in porphyroclast tails (bottom) have a grain size of

~10 μm .

944 Figure 3. a) Optical micrograph, b) electron backscatter image of the tail on a K-feldspar
 945 porphyroclast in mylonitic gneiss (PW25, Whipple Mtns core complex, SE California).
 946 Quartz was precipitated in cracks in the feldspar, and becomes mixed with recrystallized
 947 feldspar in the tail. Grain size in the mixture is $< 20 \mu\text{m}$; in pure quartz ribbons the grain
 948 size is $> 100 \mu\text{m}$.

949 Figure 4. a) Intimate mixing of fine-grained quartz (dark grey), albite (medium grey) and
 950 K-feldspar (light grey) in tail of porphyroclast shown in Figure 3. b) EBSD map of K-
 951 feldspar in the same area. Black areas are quartz. c) EBSD map of quartz in the same
 952 area. Note dispersion of finely recrystallized quartz within feldspar (black).

953 Figure 5. a) Plagioclase porphyroclasts in granitoid mylonite (PC32, Aspromonte massif,
 954 Calabria, Italy). Tails are a mixture of dynamically recrystallized quartz and plagioclase.
 955 b) EBSD map of quartz in a tail (location outlined in a). Note relatively coarse-grained
 956 quartz in ribbon on left (mean grain size $44 \mu\text{m}$), compared with finer grained quartz
 957 (mean grain size $15 \mu\text{m}$) mixed with plagioclase (black) on right. c) EBSD map of
 958 plagioclase, same area as b. Finely recrystallized plagioclase forms thin ribbons; larger
 959 grains are partly recrystallized relicts from the porphyroclast at top. Mean grain size of
 960 the fully recrystallized plagioclase is $13 \mu\text{m}$.

961 Figure 6. a) CPO in quartz from the coarse-grained ribbon at left in Figure 5b. b) CPO in
 962 quartz from the fine-grained quartz-feldspar mixture on the right in Figure 5b. c) CPO in
 963 plagioclase from the fine-grained quartz-feldspar mixture on the right in Figure 5b.

Figure 7. Cartoon illustration of the process of two-phase mixing in a granitoid rock, based on the example shown in Fig. 2. Dynamic recrystallization of primary feldspar creates an aggregate with a grain size controlled by the piezometric law for feldspar. The aggregate starts to deform by grain-boundary sliding, and quartz diffuses into the spaces between the grains. The end product has a grain size controlled by the feldspar, but the dominant deformation mechanism is diffusion creep in quartz.

Figure 8. Strain-rate–grain-size (a), stress–strain-rate (b), and stress-T plots for dislocation creep (Hirth et al., 2001) and three different pressure solution flow laws in quartz (after Den Brok et al., 1998). See Appendix for the calculation method. Stress values are differential stress.

Figure 9. a) Stress–grain-size deformation mechanism map for quartz, using the flow law of Hirth et al. (2001) for dislocation creep, and the flow law of den Brok (1998) for thin-film pressure solution creep. Piezometric line for quartz from Stipp and Tullis (2003), D_{min} line from Platt and Behr (2011a). b) Stress–grain-size deformation mechanism map for wet feldspar (maximum water fugacity for the quoted pressure and temperature), using the flow laws for dislocation creep and diffusion creep of Rybacki et al. (2006). Feldspar piezometric line from Post and Tullis (1999). c) Stress–grain-size deformation mechanism map for wet granite/granitic ultramylonite, based on the composite flow laws discussed in the text. Dynamic recrystallization and mixing at 45 MPa differential stress from 1 mm grain size (A) to 4 μm on the feldspar piezometric line (B) would result in an increase of strain-rate by a factor of ~ 5000 . Note that granite in the dislocation creep field is several thousand times stronger than pure quartz, but in the diffusion creep field the

strengths are fairly similar. d) Stress-temperature map for wet granitic ultramylonite, with grain size calculated using the piezometric relationship for feldspar from Post and Tullis (1999). Deformation is predicted to be in the diffusion creep field through the whole range of stress and temperature. Note, however, that this map is likely to be unreliable at high temperature, where water activity will be limited by partial melting. Stress values in all maps are differential stress.

Figure 10. Orthopyroxene porphyroclast with dynamically recrystallized tail in a xenolith from the Jagersfontein kimberlite, redrawn from an EBSD image in Bruijn and Skemer (2014). Dynamically recrystallized pyroxene has a grain size of ~50 μm , and olivine ~200 μm , both of which correspond to differential stress in the 20-30 MPa range (Skemer and Karato, 2008), which suggests that the microstructure in both minerals has not been significantly modified by grain growth. Note the decrease in recrystallized grain size in the tail with distance from the porphyroclast, which suggests that a component of dislocation creep and dynamic recrystallization continued to operate in the tail. Note also the scattered grains of olivine within the tail, suggesting diffusional mixing of the two phases.

Figure 11 a) and b), stress–grain-size and stress–T maps for olivine; b) and c) for orthopyroxene; c) and d) for harzburgite with 70% olivine and 30% orthopyroxene, all at 1 GPa and 1000°C, based on the parameters in Table 1 and the composite flow laws discussed in the text. Olivine piezometer from van der Wal et al. (1993); orthopyroxene piezometer from Bruijn and Skemer (2014); D_{min} line for olivine calculated for a surface

1007 energy of 1 J/m^2 (Tasaka and Hiraga, 2013) using Eqn 4. Water content in olivine is
1008 calculated from the fugacity using the formulation of Hirth and Kohlstedt (2003).

1009

1010 **Table and table caption**

	Qtz wet dis creep	Fsp wet dis creep	Fsp wet diff creep	Opx dis creep	Opx diff creep	Ol wet dis creep	Ol wet diff creep
$A \text{ Pa}^{-n} \text{ m}^q \text{ s}^{-1}$	6.31E-36	1.58E-18	2.0E-25	2.75E-9	5.5E-17	3.00E-20	3.33E-19
$Q \text{ (J/mole)}$	1.35E5	3.45E5	1.59E5	6E5	3.8E5	4.8E5	3.35E5
$V \text{ (m}^3\text{/mole)}$		38E-6	38E-6			1.1E-6	4E-6
n	4	3	1	2.9	1	3.5	1
$K \text{ m Pa}^{-p}$	8.899E4	0.5016		17.3156		1.4325E6	
p	-1.26	-0.66		-0.7645		-1.33	
q	0	0	3	0	3	0	3
r	1	1	1			1.2	1
ref	1	2	2	3	4	5	5

1011 Table 1. Flow law parameters for quartz (qtz), feldspar (fsp), orthopyroxene (opx) and
1012 olivine (ol). Dis creep, dislocation creep; diff creep, diffusion creep. A , prefactor; Q ,
1013 activation energy; V , activation volume; n , stress exponent in flow law; K constant for
1014 piezometric relation; p , exponent in piezometric relation; q , grain-size exponent in flow
1015 law; r , exponent for water content/fugacity; k , conductivity; κ , diffusivity; ref, reference.
1016 1, Hirth et al. (2001); 2, Post and Tullis (1999); Rybacki et al. (2006); 3, Bruijn and
1017 Skemer (2014); 4, Tasaka et al. (2013); Bruijn and Skemer (2014); 5, Hirth and Kohlstedt
1018 (2003); Freed et al. (2012). Water fugacity for quartz and feldspar flow laws was
1019 determined as a function of P and T from the fugacity calculator at
1020 <http://www.geo.umn.edu/people/researchers/withe012/fugacity.htm> (Pitzer and Sterner,
1021 1994). Water content for olivine flow law is calculated as ppm H/Si from fugacity using
1022 the expression from Hirth and Kohlstedt (2003). The published values of the prefactors
1023 are based on these values.
1024

1025

sample	Qtz μm	Qtz MPa	QF mixtures μm	Fsp MPa
PC32	44	24.3	13	8.9
PW25	100	12.7	20	4.6
Ref 1	110	11.7	15	7.2
Ref 2	50	22.0	9	15.5
Ref 2	40	26.2	6	28.7
Ref 2	30	32.9	5	37.8

1026

1027 Table 2. Comparison of dynamically recrystallized grain-size in monomineralic quartz
 1028 ribbons and in quartz-feldspar ultramylonite bands in the same sample, together with the
 1029 differential stress inferred using the quartz and feldspar paleopiezometers respectively.
 1030 PC32 and PW25, this paper; Ref 1, Kilian et al. (2011); Ref 2, Stünitz and Fitz Gerald
 1031 (1993).

Platt Figure 1

Platt Figure 2

Platt Figure 3

Platt Figure 4

a

b

c

Platt Figure 5

Platt Figure 6

Platt Figure 7

Platt Figure 8

Platt Figure 9

Platt Figure 10

Platt Figure 11