

# Lithospheric shear zones as constant stress experiments

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## ABSTRACT

Below the seismogenic zone, plate boundaries are defined by lithospheric ductile shear zones. These localize strain, and are the main reason that convective motions in the Earth's interior are expressed at the surface as plate tectonics. We argue here that lithospheric shear zones operate at approximately constant stress, equal to the yield strength of the surrounding rocks. This places constraints on the bulk strength of the lithosphere, and allows us to calculate the cumulative width of the shear zones as a function of depth. The concept applies most clearly to strike-slip shear zones, but is also applicable to thrust- and normal-sense shear zones, although these evolve with time due to temperature changes associated with burial or exhumation. If shear zones operate at constant stress, this affects their microstructural evolution, and hence their rheology. Decreases in grain size due to dynamic recrystallization may cause a switch in the dominant deformation mechanism to grain-size-sensitive creep, leading to weakening and strain localization. A common argument, based on a constant strain rate approach, has been that such transitions will be inhibited by grain growth. In a constant stress shear zone, however, dynamic recrystallization continues to maintain the low grain size even after the switch has occurred. Grain growth is inhibited under these conditions, and hence the switch is permanent as long as the boundary conditions remain the same. Grain-size reduction in shear zones is therefore a critical factor in maintaining plate tectonic processes.

## INTRODUCTION

Lithospheric deformation is highly localized, and this is a defining feature of plate tectonics: numerical models of mantle convection fail to reproduce plate-like behavior in the thermal boundary layer unless the material shows a significant degree of strain localization (e.g., Bercovici, 2003). The primary cause of strain localization is weakening caused by microstructural modification during deformation, resulting in the formation of faults or ductile shear zones (Poirier, 1980; Rutter, 1999); this leads to a conundrum, however. Is the strength of the lithosphere controlled by the strength of the undeformed rocks (granite, gabbro, or peridotite) that make up the bulk of the plates, or by the strength of the deformed and weakened material found in shear zones? Or should we treat the lithosphere as a complex two-phase system, with a strength controlled by the volume proportion of weakened material, and the orientation and connectivity of the shear zones?

A second conundrum arises from the fact that the mechanical properties of rocks in the ductile field (which makes up much of the lithosphere) are commonly expressed in terms of a relationship between stress and strain rate (rheology). To evaluate strength (the stress required to produce a given strain rate), we have to specify the strain rate. Do we use an average strain rate for the entire plate, or the much higher strain rate within the shear zones that accommodate the deformation?

In this paper we develop the concept that ductile shear zones on the lithospheric scale are constant stress experiments. This leads us to the conclusion that the bulk strength of the

lithosphere is controlled by the yield strengths of the undeformed rocks that make it up, and does not depend on the mechanical properties of the shear zones. It also allows us to calculate the cumulative width of the shear zones at any temperature, and therefore depth, within the lithosphere, which answers some vexing questions about the deep structure of major fault zones. We show that under constant stress conditions, the likely cause of weakening in shear zones is a switch to grain-size-sensitive deformation mechanisms. This conclusion addresses doubts as to the efficacy of this process, which arose from analyzing the grain-size evolution in terms of constant strain rate, rather than constant stress (e.g., De Bresser et al., 2001).

## LITHOSPHERIC SHEAR ZONES AS CONSTANT STRESS EXPERIMENTS

Many rock mechanics experiments are carried out at constant strain rate. Under these conditions, weakening causes a drop in the applied stress (Tullis and Yund, 1977). At constant stress, however, weakening causes an increase in strain rate, and any intermediate situation is possible. Consider the simplest possible case of the initiation of a new transform plate boundary cutting a plate of pristine lithospheric material consisting of homogeneous coarse-grained crystalline rock. At low temperatures such materials exhibit macroscopically plastic behavior; that is, deformation is elastic until the stress exceeds a critical value (the yield stress,  $\sigma_y$ ) (Ranalli, 1987). The yield stress is a function of temperature and depth, following the Coulomb failure law down to the brittle-ductile transition, and then decreasing below that with tem-

perature. The external boundary condition is the imposed relative plate velocity,  $V$ . If this is to be accommodated, the stress within the plate at any given depth must reach the yield stress at that depth. The plate will then yield on one or more subparallel faults or shear zones parallel to the direction of plate motion (the pattern of faults within the 400-km-wide San Andreas transform of California is characteristic of this). Within the developing shear zones, microstructural changes such as grain-size reduction cause the material to become weaker, and at depths below the brittle-ductile transition its behavior evolves toward a ductile flow law that can be expressed by a relationship between strain rate and stress of the form  $\dot{\epsilon} = A_s \sigma_s^n$ , where  $A_s$  describes the material parameters of the shear zone mylonite, including temperature dependence following the Arrhenius law,  $\sigma_s$  is the flow stress in the shear zone, and  $n$  has a value in the range 1–10 (e.g., Hirth et al., 2001). The minimum cumulative width  $w$  of shear zone material has to be such that it can accommodate the plate velocity at a flow stress no higher than the yield stress. The strain rate in the shear zones will then be  $\dot{\epsilon} = V/w$ , and if  $\sigma_s = \sigma_y$ ,

$$w = \frac{V}{A_s \sigma_y^n} \quad (1)$$

During the weakening process, the strain rate in the deforming zones increases, so they get progressively narrower. The value of  $w$  cannot drop below the value given by Equation 1, however, or  $\sigma_s$  will exceed  $\sigma_y$  in the surrounding rock, which will then deform, increasing  $w$ . When  $w$  has the value given by Equation 1, the yield stress is not exceeded, and there is no reason for the shear zones to widen. Hence the flow stress in the shear zones will remain at  $\sigma_y$ , and their evolution can be described as a constant stress experiment. This conclusion may seem counterintuitive, as the plate-boundary condition invoked is one of constant velocity. The constant velocity boundary condition is converted to a constant stress boundary condition within the shear zones because of the buffering effect of the yield stress in the surrounding rock.

There are tectonic situations in which this concept does not apply. Continental lithosphere is commonly cut by numerous preexisting faults and shear zones, and may also contain boundaries between rocks of different properties, which are likely to concentrate stress and deformation on various scales. If the cumulative width of preexisting suitably oriented shear zones,  $w_0$ , exceeds the minimum value for  $w$  specified by

Equation 1, then the stress required to accommodate plate motion will be less than the yield stress of the undeformed material. The average strain rate in the shear zones will be  $V/w_0$ , and the flow stress will be:

$$\sigma_s = \left( \frac{V}{w_0 A_s} \right)^{\frac{1}{n}} \quad (2)$$

A decrease in the relative plate velocity may also result in the flow stress in the shear zones dropping below the yield stress in the undeformed material. The value of  $w$  will then will be too great for the new velocity, and the flow stress will be controlled by Equation 2, with  $w_0$  being the width of the shear zones before the change in velocity.

Normal and reverse faulting regimes exhumate or bury the footwall of the shear zone, leading to changes in pressure and temperature that complicate the situation. Exhumation and cooling up to the brittle-ductile transition cause all rocks to become stronger (because of the temperature dependence of ductile deformation), but continuing microstructural change within the evolving shear zones increases the strength contrast with the surrounding rock. As a result, strain becomes more localized ( $w$  decreases, and strain rate increases). The value of  $w$  continues to be governed by the process of microstructural weakening relative to the surrounding rock (which may be older shear zone material formed under higher temperature conditions). Flow stress in the shear zones at any depth therefore continues to equal the yield stress in the surrounding material at that depth and temperature, and increases with exhumation until the brittle-ductile transition is reached. Conversely, thrust-sense shear zones are likely to undergo burial and temperature increase, causing all rocks to become weaker. Strain becomes less localized, flow stress and strain rate decrease, and  $w$  increases. Widening of the shear zones requires previously intact rock to be deformed, and hence the flow stress continues to equal the yield stress in the surrounding rock, but decreases as the rocks become warmer.

The aim of this section has been to show that within a wide range of tectonic settings, the flow stress in a lithospheric-scale shear zone equals the yield strength of the surrounding rock, and there will not be a stress drop as a result of weakening. We use the term “constant stress experiment” to describe this condition, but accept that exhumation or burial will change the value of the yield stress, and that preexisting structures in the lithosphere may complicate this simple statement. The constant stress condition implies that the lithosphere behaves as a plastic medium, with a bulk strength controlled by the yield strengths of the undeformed rocks that

make it up. Its strength is therefore independent of the rheology of the shear zones that cut it.

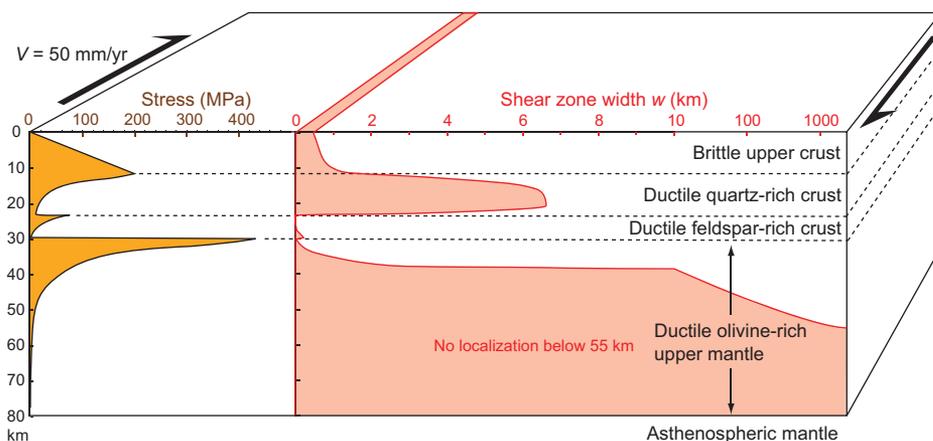
### WIDTH OF LITHOSPHERIC SHEAR ZONES AT DEPTH

The yield stress of undeformed rock decreases with increasing temperature and depth in the lithosphere, and the degree of strain localization also decreases. As a result,  $w$  is likely to increase with depth, and both the flow stress and the strain rate will decrease correspondingly. Where the shear zones cut compositionally stronger layers, however, they may become narrower. Quantitative data on yield stresses for rock are difficult to obtain; steady-state flow laws for time-dependent creep obtained in laboratory experiments can be extrapolated toward geologically realistic strain rates and temperatures (Hirth and Tullis, 1992), but data on plastic yielding, which is a non-steady-state process, cannot be extrapolated in the same way. Our concept of natural shear zones as constant stress experiments, however, suggests that measurements of paleostress in shear zone mylonites can give us a proxy for the yield stresses at various temperatures (and hence depths) in the lithosphere. This should not be affected by the tectonic setting of the shear zone, although it will be a function of lithospheric composition. We therefore use a stress-temperature profile recently measured by Behr and Platt (2011) from mylonites in the Whipple Mountains core complex in southeast California, using the dynamically recrystallized paleopiezometer of Stipp and Tullis (2003). If, as we propose, these flow stresses provide a measure of how the yield strength of undeformed rock varies with temperature in the orogenic lithosphere of western

North America, then the stress-temperature profile from this area can be converted to a strength-depth profile using an appropriate geotherm. The resulting strength profile for the quartz-rich part of the crust (Behr and Platt, 2011) is shown in the upper left of Figure 1.

From the stress-depth data we can calculate  $w$  at any depth for a transform boundary such as the San Andreas system, assuming that it cuts lithosphere of comparable composition to that in the Whipple Mountains. This requires choice of a flow law appropriate for the shear zone, which is complicated by our limited understanding of the weakening processes that create these zones. For illustrative purposes, we use a flow law similar to that estimated by Hirth et al. (2001) for quartz-rich mylonites (for details, see the GSA Data Repository<sup>1</sup>). The Hirth et al. (2001) flow law is based on experimental data, with the material parameters adjusted to match estimated stresses, strain rates, and temperatures from the Ruby Gap duplex in Australia. The Whipple mylonites were formed under similar conditions and show similar microstructures, but have somewhat higher strain rates for a given level of stress. We calculate  $w$  versus depth for the San Andreas transform system (50 mm/yr relative motion) to a depth of 23 km, assuming an appropriate lithospheric structure and geotherm, and this is shown with the red curve in the upper part of Figure 1.

Comparable data sets are not available for lower crustal and mantle rocks, but we estimate a stress profile through this part of the lithosphere (Fig. 1) using published flow laws for dislocation creep in feldspar and olivine, and a strain rate of  $10^{-15} \text{ s}^{-1}$ , which is equivalent to distributing a relative plate velocity of 50 mm/yr



**Figure 1. Differential stress and cumulative width  $w$  of transform boundary in compositionally layered lithosphere. Flow stress in shear zones is equal to yield strength of lithosphere, so stress profile applies to both. Note change in scale for  $w$  above 10 km.  $V$ —velocity. (For tabulated results and basis for calculations, see the Data Repository [see footnote 1].)**

<sup>1</sup>GSA Data Repository item 2011060, supplementary information, is available online at [www.geosociety.org/pubs/ft2011.htm](http://www.geosociety.org/pubs/ft2011.htm), or on request from [editing@geosociety.org](mailto:editing@geosociety.org) or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

across a zone the width of the North American Cordillera (1500 km). This gives some indication of the stress required to initiate weakening (but see caveat above about determination of yield stresses). We then assume that grain-size reduction results in a switch to diffusion creep in these minerals, and use flow laws for this mechanism to estimate strain rates, and hence  $w$ , for the San Andreas transform through the lower crust and uppermost mantle (red curve in Fig. 1) (for details of these calculations, see the Data Repository).

Our calculations suggest that the cumulative width of the shear zones making up the San Andreas transform may exceed 6 km in the middle crust, narrows to a few tens of meters in the lower crust and uppermost mantle, and then widens again to 80 km at a depth of 45 km. At depths >55 km, we predict that weakening mechanisms do not operate, and hence will not contribute to strain localization. These conclusions depend on the choice of lithospheric composition and thermal structure, and are very sensitive to the flow-law parameters, which are not well constrained, but they are consistent with the limited available data on the width of plate-boundary faults at depth (e.g., Vauchez and Tomassi, 2003; Herquel et al., 1999; Rumpker et al., 2003; Baldock and Stern, 2005).

## IMPLICATIONS FOR RHEOLOGY OF SHEAR ZONES

Our concept of lithospheric shear zones as constant stress experiments has consequences for our understanding of their microstructural evolution, and hence of weakening mechanisms. The processes that lead to weakening and strain localization include shear heating, changes in mineral assemblage due to hydration and retrogressive metamorphic reactions (Jefferies et al., 2006), development of crystallographic preferred orientations due to dislocation creep (Poirier, 1980), and grain-size reduction caused by brittle fracture and dynamic recrystallization (Rutter and Brodie, 1988; Stewart et al., 2000). Here we focus on one of the most potent causes of microstructural weakening, which is associated with grain-size reduction. Crystal-plastic deformation leads to an increase in the internal strain energy of the deforming crystals, due to the build-up of lattice defects (dislocations). This causes dynamic recrystallization, which involves the creation of new grains and the formation of lattice that is largely free of dislocations, reducing the total free energy of the system. These new grains are in general much smaller than preexisting grains, and grain-size reduction is one of the characteristic features of the rocks that occupy shear zones, known as mylonites (White et al., 1980). A large body of experimental evidence from metals, salts, and rock-forming minerals such as calcite,

quartz, feldspar, and olivine demonstrates that the recrystallized grain size is an inverse function of the flow stress (De Bresser et al., 1998; Kohlstedt and Weathers, 1980; Post and Tullis, 1999; Stipp and Tullis, 2003; Twiss, 1977; van der Wal et al., 1993), and this relationship has been used as a way of measuring paleostresses in mylonites.

Dynamic recrystallization is thought to contribute to weakening in three specific ways. (1) It acts to reduce dislocation density, and hence to counteract work hardening during low-temperature dislocation creep (Fliervoet and White, 1995; Tullis and Yund, 1985). (2) Grain-size reduction may increase the contribution of grain-boundary sliding to the deformation, and this has been incorporated into flow laws for olivine as a discrete deformation mechanism (Hirth and Kohlstedt, 2003). (3) Grain-size reduction could also lead to a switch to grain-boundary diffusion creep (Coble creep), which is strongly grain-size sensitive, the strain rate increasing with the inverse cube of the grain size (Rutter and Brodie, 1988). This could result in a marked decrease in strength in olivine and calcite rocks (Montési and Hirth, 2003; Schmid et al., 1977).

In a constant strain rate experiment, these weakening processes cause a decrease in flow stress (path A to C in Fig. 2). This in turn results in a decrease in the rate of dislocation creep, and hence of grain-size reduction. De Bresser et al. (2001) argued that as a result, grain growth driven by surface energy will counteract grain-size reduction, and move the material back toward the boundary with the dislocation creep field (C to D in Fig. 2). They therefore maintain that deformation mecha-

nism switches, and the weakening they cause, are unlikely to be permanent.

If natural shear zones operate at constant stress, the argument changes. The switch in the dominant deformation mechanism causes an increase in strain rate, not a stress drop (path A to B in Fig. 2). Hence the rate of dislocation creep, the dislocation density, and the dynamically recrystallized grain size, all of which are functions of stress, remain the same. At the levels of differential stress typical of mylonites (50–250 MPa) (Dunlap et al., 1997; Fitz Gerald et al., 2006; Stipp and Kunze, 2008), strain energy-driven grain-boundary migration ( $\rho$ GBM) is orders of magnitude faster than surface-energy driven migration ( $\gamma$ GBM), and will dominate the microstructure (Humphreys and Hatherly, 2004). Grain growth driven by  $\gamma$ GBM is inhibited by  $\rho$ GBM, and there is no reason for further evolution of the grain size, which will be maintained at the level associated with the flow stress. We therefore argue that grain-size reduction in constant stress shear zones is likely to lead to a substantial degree of weakening, and may be a key cause of strain localization in the lithosphere.

## CONCLUSIONS

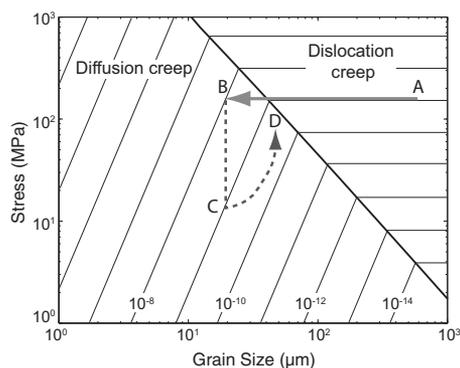
Lithospheric shear zones approximate constant stress experiments; the flow stress in the shear zones equals the yield strength of the surrounding rock. The bulk strength of the lithosphere is therefore controlled primarily by the plastic yield strength of undeformed rock, even though the deformation takes places almost exclusively in shear zones that are much weaker.

The yield stress and the degree of weakening decrease with depth and temperature below the brittle-ductile transition, which means that the shear zones will in general widen with depth, until they reach temperatures at which ductile deformation is completely distributed. This provides us with a quantitative way of addressing the deep structure of plate-boundary faults.

In constant stress shear zones the microstructural evolution involves an abrupt decrease in grain size due to dynamic recrystallization. In some rock types, this may cause a transition in the dominant deformation mechanism to grain-size-sensitive creep; the strain rate at constant stress then increases, leading to strain localization. At constant stress, the rate of dislocation creep remains the same, so dynamic recrystallization continues to maintain the low grain size. This addresses the questions that have been raised about whether deformation mechanism switches could lead to permanent weakening in the lithosphere.

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**Figure 2. Stress–grain-size map for olivine at 1000 °C, based on flow laws from Hirth and Kohlstedt (2003). Grain-size reduction caused by dynamic recrystallization (path A to B) can result in a switch to diffusion creep. If strain rate remains constant, grain growth may return material to dislocation creep field (C to D), as suggested by De Bresser et al. (2001). At constant stress, material remains at B, and switch is permanent.**

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